Possible Orogeny parallel Lower Crustal Flow and Thickening in the Central Andes

Muriel Gerbault, Joseph Martinod, Gérard Hérail

Possible Orogeny-parallel Lower Crustal Flow and Thickening in the Central Andes

M. Gerbault, J. Martinod, G. Hérail

IRD UR 104, LMTG , University P. Sabatier, 38 rue des 36-ponts , Toulouse, France
fax: 33(0)5 61 55 82 50, email: gerbault@lmtg.ups-tlse.fr, corresponding author: M. Gerbault

REVISED VERSION, may 20th 2003.

Abstract: The relatively low elevation and thick crust in the Altiplano, in comparison to the higher elevation, but thinner crust in the Puna plateau, together with geophysical data, suggest that isostatic equillibrium is achieved by cooler and denser lithospheric mantle in the Altiplano. Excess density in the Altiplano mantle could create differential horizontal stress in the order of 25 MPa between both lithospheric columns. Numerical models accounting for pressure and temperature-dependent rheology show that such stress can induce horizontal ductile flow in the lower crust, from the Puna towards the Altiplano. With a minimum viscosity of 10^{19}Pa.s, this flow reaches 1 cm/yr, displacing more than 50 km of material within 5 Ma. If the lower crust viscosity is smaller, the amount of orogeny-parallel lower crustal flow can be even greater. Such a mechanism of channel flow may explain that different amounts of crustal material have been accomodated by shortening in the Altiplano and in the Puna. Because of the strength of the - brittle - upper crust, this channel flow does not necessitate large amounts of suface deformation (except vertical uplift), making it difficult to detect from the geology.

Keywords: orogeny, isostasy, lower crust flow, mantle density.

1 Introduction

From about 15°S to 27°S, the Central Andes comprise five main morphological sections: the western forearc, the eastern subandean zone, and a central high plateau area, the Altiplano-Puna, which is delimited to the north and south by the coalescence of the Eastern and Western Cordilleras (Fig. 1). A noticeable feature of this high plateau area is that the mean height varies from 3.7 km in the Altiplano to 4.4 km further south in the Puna.

One of the misunderstood features concerning the Central Andes is that despite comparable direction, speed, and geometry of the subducting Nazca plate, the orogeny displays a large bend, associated to considerable variations in estimated amounts of crustal shortening. These estimates, based on two-dimensional balanced crossed sections along different latitudes, also vary with authors because of different reference distances and times that can be considered in geological reconstructions. However, the amount of shortening generally decreases from the Arica bend toward the south: the section from Arica at 16°-21°S to the thin-skinned Subandean belt has undergone maximum shortening, estimated to about 50 km in the Altiplano and more than 200 km in the Subandean zone since the Oligocene (e.g. Roeder, 1988; Roeder & Chamberlain, 1995; Baby et al., 1989, 1995; Rochat 2002). In contrast, further south across the Puna plateau at 23-25°S, total shortening has been estimated to be less than 75 km (Allmendinger et al., 1997), with as little as 40 km in the Subandean, thick-skinned ranges (Coutand et al., 2001).

Compilations by Allmendinger et al. (1997) and Kley & Monaldi (1998) illustrate a discrepancy of about 20% between crustal volumes deduced from these shortening estimates, and crustal volumes deduced from present day seismic Moho depth subtracted by an hypothetical constant initial crustal thickness. In fact, initial crustal thickness is poorly constrained, and geological observations indicate that the Altiplano and Puna have displayed structural differences for a long time. Andean shortening could vary according to preexisting heterogeneities (e.g. Kley et al., 1999; Coutand et al., 2001; Sempéré et al., 2002); for instance from ~21°S to 23°S, the dominantly Early Paleozoic sedimentary province in the eastern part of the orogen changes to a dominantly marine arc and younger Precambrian basement, which coincides with the termination of the thin-skinned Subandean belt.

Other mechanisms have been proposed to explain these ~20% lack of shortened crustal material (e.g. Allmendinger et al., 1997). According to a compilation by Giese et al. (1999), tectonic shortening in the Central Andes (at about 21°S) would contribute to at least 55% of the root formation, hydration of the mantle wedge in the forearc region would produce up to 20% of crustal material, and magmatic addition and tectonic erosion would contribute only for about 5%. Finally, uncertainties rise from the possibility that the classical seismic Moho discontinuity may not correspond to the petrological crust/mantle boundary, so that the present day crustal volume could be overestimated (Giese et al., 1999; Yuan et al., 2000, 2002; Beck & Zandt, 2002).

Another explanation is that mass transfer also occurs in a direction perpendicular to the direction of convergence (e.g. Dewey & Lamb 1992; Lamb & Hoke, 1997; Kley & Monaldi, 1998; Kley et al., 1999; Jaillard et al., 2001). However, observed strike-slip features within the Altiplano-Puna region do not seem to accommodate significant amounts of displacements (more than 10 km): from kinematical reconstructions and paleomagnetism, these strike-slip structures are linked with crustal block rotations that accommodate the regional increase in oroclinal bending (Lamb et al., 1997; Lamb & Hoke, 1997; Lamb, 200, 2001; Coutand et al., 1999, 2001). On the other hand, the development of a zone of continental NS dextral shear in the Puna of the northern Argentinian Andes, during at least the last 10 Ma and without regional bending, is understood by Lamb (2001) to be the consequence of along-strike buoyancy stress gradients between 19°S and 23°S.

Buoyancy stress gradients were discussed by Froidevaux et Isacks (1984), when analysing the quantitative relationship between the geoid, topography and stress regime in the region of the Altiplano-Puna. Froidevaux & Isacks (1984) showed that an average elevation near 4 km is consistent with compensation by an unresolved combination of density heterogeneities due to the crustal root and a hot uppermost mantle.

Since Froidevaux & Isacks (1984), considerable data were gathered that help clarify lithospheric scale heterogneities. In the following we first review geophysical data on the deep structure of the Altiplano and Puna lithospheres, and discuss their state of mechanical equilibrium. We then explore with numerical modelling, the possible consequences of an heterogeneous density in the mantle on orogeny-parallel lower crust flow. We then infer that the distribution of the crustal volume of the Central Andes may be partly due to longitudinal (N-S) mass transfers occurring in the deep parts of the crust, and in such a way that it has little surface expression.

2 A link between the surface and the mantle in the Central Andes ?

2.1 Geophysical measurements and heterogeneity of the mantle

The seismological lithospheric structure in the Central Andes appears complex. However, shortening variations estimated using balanced cross-sections has been linked with variations in the mantle for a while. First in 1992, Whitman et al. noticed that attenuation of P waves at depths between 100 to 150 km, is not only low under the narrow active magmatic arc, but is contrastingly high under the Altiplano area (Fig. 2a) and low under the Puna extending to the east as far as the Santa Barbara system (Fig. 2b). Whitman et al. (1996) then presented a longitudinal north-south cross-section (Fig. 2c) showing the geometrical correspondence between the Altiplano's topographic low (800 m lower than to the north and south) and the seismic strong attenuation pattern in the mantle (Fig. 2c). North and south of the central Altiplano where the average topography is higher, warm asthenospheric material would fill the gap between the crust and the subducting plate.

Watts et al. (1995) using gravity data to calculate along strike changes in flexural rigidity, and Whitman et al. (1996) interpreting their seismic attenuation models, argue that the Altiplano bend region is composed of mechanically strong lithosphere, whereas the Puna region is composed of a thin and hot lithosphere. The coincidence of thin lithosphere with the area affected by Cretaceous rifting (25°S, Santa Barbara system) suggests that thin lithosphere may have been inherited from Mesozoic extension (Kley, 1996; Sempéré et al., 2002), which would have been less active in the Altiplano area. Another possibility suggested by geochemical and geophysical data near 24°-27°S (where young mafic magmatism occurs) is that lithospheric delamination occurred at 2-3 Ma (Kay et al. 1994, Whitman et al. 1996).

More recent seismological investigations have improved our knowledge on the lithospheric structure across the entire plateau, using teleseismic or local earthquake receiver function analysis, regional waveform modeling, tomographic inversions, and Moho reflection studies. Crustal thickness varies from north to south: for Yuan et al. (2001), the Moho is generally deeper than 60 km north of 23°S, while south of 23°S it is generally shallower than 60 km. For Beck & Zandt (2002), with given average crustal Vp and Poisson's ratio (0.25), crustal thickness in the Central Altiplano is 59-64 km, and reduces to ~50 km at 20.5°S. Lithospheric structure is also interpreted differently depending on authors, although there is a general agreement on a more asthenospheric seismic character to the south: Myers et al. (1998) find that from 18°S to 20.5°S the Altiplano mantle is consistent with a mantle lithosphere down to about 125-150 km depth, and that its asthenospheric character begins at about 21°S. For Yuan et al. (2002), a best fit of altitude and Bouguer anomaly with isostatic-Moho depth is obtained, assuming a felsic crust, when the lithospheric thickness is closer to 100 km than 150 km. In general they conclude that the systematic difference in altitude between the Altiplano and the Puna, from north to south of 23°S, can be explained either by some 20 km thinner lithosphere beneath the Puna or by a less dense Puna crust. Beck & Zandt (2002) interpret their results between 18°S and 20.5°S as the strong Brazilian lithosphere underthrusting the western part of the Eastern Cordillera but not the entire Altiplano (as did Dorbath et al. 1993). These authors propose that along-strike crustal thickness and elevation variations are linked to the dynamical process of delaminating lithospheric mantle, that would have already occured in the southern Altiplano, but is not yet completely removed beneath the Central Altiplano.

Magnetotelluric images of the crust and upper mantle in the northwestern Argentinean Andes reveal a prominent conductor rising from depths of 180 km below the Chaco plain, to 80 km beneath the Santa Barbara system and the Eastern Cordillera (Lezaeta et al., 2000; Brasse et al., 2002). This conductor is interpreted as a rise of the conductive asthenosphere, in agreement with seismic attenuation results (Whitman 1992, 1996; Haberland & Rietbrock, 2001).

2.2 Channel flow and density variations in the upper mantle

Variations in crustal thickness create a potential energy that can induce material flow in order to return to equilibrium (Artyushkov, 1973). Bird (1991) showed that an orogenic crustal root can induce outward channelised flow in the lower crust when the vertical variation of viscosity is accounted for (resulting from changes in rock composition and increase in temperature). This mechanism is now invoked to explain the development of high plateaus (Royden, 1996; Shen et al., 2000; Medvedev et al., 2001), decoupled deformation between the upper crust and the lithospheric mantle (as in the India-Asia collision zone, Royden et al., 1997), or exhumation of lower crust enhanced by surface processes (Beaumont et al., 2001). Since the Altiplano Moho is deeper than the Puna Moho by about 10 km, lower crustal flow should develop from the Altiplano towards the Puna. But since the Altiplano is lower in altitude than the Puna by about 800 m, the isostatic equilibrium between both lithospheric columns must be more complex.

Because the Puna has higher elevation than the Altiplano, one could argue that the Puna crust is lighter. However, there are other arguments that indicate that this is not the case, despite the lack of seismic refraction data for the Puna crust: 1) Neogene sedimentary basins would be of greater thickness under the Altiplano (reaching 10 km) than under the Puna (of the order of 6 km) (Vandervoort, 1993; Coutand, 1999), 2) the basaltic volcanism present in the Puna (Kay et al., 1994), as well as the silicic caldera complexes between 21°S and 24°S (e.g. Haberland & Rietbrock, 2001), would tend to increase the mean density of the Puna crust, and 3) seismic investigations mentionned in the previous section (e.g. Beck & Zandt, 2002; Yuan et al., 2002), do not show sufficient decrease in crustal seismic velocities from north to south. As a matter of fact, if we assume that ~1 km higher surface elevation and 11 km thinner crust of the Puna with respect to the Altiplano are only compensated by the crust, then isostatic equilibrium would require an average density difference of 150-200 kg/m^3.

These informations suggest that isostatic equilibrium between the Altiplano and the Puna plateaus involves at least partly, density heterogeneities in the upper mantle.

From seismological investigations (see section 2.1), the absence of increase in lithospheric thickness in the Puna, where crustal thickness is almost twice that of a normal crust, strongly suggests a different geothermal gradient than in the Altiplano. As mentionned in section 2.1, a warm geotherm could either be explained by long-term tectonic history differences, or relatively recent (2-3 Ma ?) partial removal of the mantle lithosphere near 24°-27°S, such as by delamination (Kay et al., 1994).

Measured heat flow data allow to identify east to west variations, but they are too scarce to show significant variations from north to south: a local W-E-trending high would be located at the boundary between Altiplano and Puna plateaus, at about 21°S (Hamza & Munoz, 1996; Springer & Forster, 1998). According to Babeyko et al. (2002) and references therein, temperatures of 800°C at 25 km depth are likely to have occurred under the volcanic arc of the Puna plateau (these authors suggest that this could be due to crustal scale convection).

A relatively hot asthenospheric upper mantle (as under the Puna) could explain higher topography than above a cooler continental mantle lithosphere (as under the Altiplano), because of the associated change in density. The conventional relationship between temperature T, pressure P, and density (e.g. Petrini et al., 2002) is :

$$\rho = \rho_o . (1 - \alpha . T + \beta . P),$$

where ρ_o is the reference density, α is the thermal expansivity and β is the compressibility. To obtain an extra density of 50 kg/m^3, assuming $\alpha = 3.10^{-5}$ K^{-1} and under identical pressure conditions, the difference in temperature must then be of about 500°C. This is a relatively high value when considering usual Moho temperatures in the range of 500-700°C, but it is a reasonable value when considering asthenospheric temperatures (about 1350°C) versus continental temperatures around 100 km depth (about 750°C).

A similar situation has been proposed to explain the juxtaposition of low-elevation thick-crust in the North American Craton and high-elevation thin-crust southeastern Canadian Cordillera (Hyndman & Lewis, 1999). There, crustal seismic velocities together with surface heat flow data suggest that isostatic equilibrium is achieved by upper mantle differential density, due to mantle temperature differences that could reach 400°C (Hyndman & Lewis, 1999).

If we thus assume that the Altiplano mantle lithosphere is denser by about 50 kg/m^3 on a thickness of about 50 km (Fig. 3a), the corresponding additional pressure is of the order of $dp = \rho.g.h$ = 25 MPa. Equilibrium of forces in the vertical direction can be achieved by 1 km higher relief in the Puna (1 km of material at 2500 kg/m^3), as is indeed observed. However, equilibrium of forces must also be achieved horizontally, and 25 MPa differential pressure could be sufficient to generate displacements in areas of sufficiently low strength, such as in the lower crust. Post-glacial rebound shows that the earth is capable of rapidly (within thousands of years) readjusting isostatic equilibrium by uplift or subsidence, and due to the very low-viscosity underlying asthenosphere, while decay of orogenic reliefs shows that several millions of years are required to readjust lateral heterogeneities within the lithosphere (e.g. Burov & Diament, 1995).

How do the lithosphere and crust adjust in response to a situation where two lithospheric columns of different density distribution are equilibrated in the vertical direction, but not in the horizontal direction ?

3 Modelling hypotheses and results

3.1 Channel flow

Channel flow (or Couette flow) can result from an applied horizontal pressure gradient dp/dx within a low viscosity layer: similarly to Turcotte & Schubert (1982), the net pressure force on an element in the horizontal x direction balances the shear force τ acting on the upper and lower boundaries of a layer of thickness h, so that $dp/dx = d\tau/dy$. When assuming that horizontal velocity U depends on the vertical coordinate only, and that we have a newtonian fluid with constant viscosity μ, one obtains:

$$dp/dx = \mu.d^2U/dy^2.$$

By integration and assuming no-slip boundary conditions (base of the upper crust and top of mantle lithosphere), the velocity profile is that of a Couette flow: $U = dp/dx.(y^2-hy)/2\mu$. For dp/dx= 25 MPa over a horizontal distance of 50 km, and for $h = 30$ km and $\mu = 10^{20}$ Pa.s., maximum velocity occurs in the middle of the layer and equals 1.875 cm/yr. Therefore significant horizontal velocities may occur in the lower crust, independently of superficial deformation.

Numerical modelling will allow us to test the amplitude of this mechanism within a more realistic geodynamical context: we incorporate the entire lithospheric scale, the boundaries of the lower crust channel are not implicitely fixed, rheology of the lower crust responds to power-law viscosity which depends on composition and temperature, and we account for the dynamical effect of variations of geometries, stresses and rheology through time.

3.2 Assumptions for numerical modelling

We use a finite differences two-dimensional code modified from Parovoz (Poliakov & Podladchikov, 1992). It is based on the Fast Lagrangian Analysis Continuum method (FLAC, Cundall & Board, 1988), which incorporates an explicit time-marching scheme in which the equation of motion is resolved for each element and at each time-step. The heat equation is resolved with the initial temperature field being calculated according to an age dependent procedure (e.g. Burov & Diament, 1995). It allows the use of pressure-dependent, non-associative, Coulomb criterion for brittle failure, and temperature-dependent creep power-law for ductile behaviour derived by rock experimentalists (see values in figure caption 3), so that the brittle-ductile transition is self-consistently defined in the modelling. See for example Gerbault et al.(2003) for details on the numerical method. We do not account for surface processes since we want to evaluate the isolated effect of lateral differences in rheology on surface relief.

The model consists of a lithosphere-asthenosphere medium 800 km long and 150 km thick, gridded with square elements of size 5 km by 5 km. The left-hand side of the model from X = -400 km to X= 0 km represents the Puna lithosphere, while the right-hand side of the model from X= 0 km to X= +400 km represents the Altiplano lithosphere (Fig. 4). The Moho depth in both the Puna and Altiplano lithospheric columns is initially set at 65 km. Variation of this depth will be discussed further in consideration of the modelling results. The surface is stress-free and surface processes are not taken into account. Both lateral borders are fixed in the horizontal direction, and free to slip vertically. Hydrostatic boundary conditions are applied at the bottom of the model, since the modelled medium is assumed to float in a perfect-fluid asthenosphere within the gravity field.

The rheology of the crust and mantle are chosen according to the known geology, chemistry and seismic character of the Central Andes. A compilation by Lucassen et al. (2001) from 21°-27°S shows that the Paleozoic metamorphic and magmatic basement in northern Chile and northwestern Argentina is largely homogeneous and felsic in composition: a) thermodynamic calculations indicate that the most important minerals are quartz and feldspars (68-76% of SiO2) under all reasonable P-T conditions at least for the upper 40 km, b) the remaining 20 km of lower crust are not mafic in the sense of a basaltic or andesitic composition, but would rather comprise discrete felsic and mafic components. These results are in accordance with geophysical observations of the velocity-density distribution at 21°-24°S (e.g. Beck & Zandt, 1996, 2002; Yuan et al., 2001, 2002): more specific information indicates that the brittle-ductile transition zone would occur at 14 to 20 km (low velocity zone), while a change to a more mafic composition would occur at 50 to 55 km depth (higher velocity).

The power-law viscous behaviour of a crust of felsic composition can be modelled with a range of dominant rock content: from the weakest wet granite to the stronger diabase (e.g. Ranalli, 1995), we choose parameters of an intermediate rock, that of quartzite (see Fig. 2 caption). Because viscosity controls the amplitude of deformation, we will compare two models M1 and M2, M1 in which the viscosity in the lower crust has a lower cutoff value of 10^{19} Pa.s, and M2 in which this cutoff is 10^{20} Pa.s. (see Fig. 5b). Below the Moho, dry olivine power-law creep parameters are chosen to represent the mantle. Despite the argument that there are fluids in the mantle due to the nearby presence of the subducting slab, it is still matter of general debate wether wet or dry olivine is the most appropriate rheology for describing large-scale behaviour of the continental mantle.

The two columns of lithosphere differ by their geotherm: a thermal age of 50 Ma is assigned to the Puna lithosphere and an age of 300 Ma to the Altiplano lithosphere (Kley et al., 1996; Sempéré et al., 2002). Although the thermal age concept should correspond to the most recent date of lithospheric thermal rejuvenation (see Burov & Diament, 1995, for applicability of the 'thermal age'), it is also a convenient variable that allows to control the concavity of the geotherm: the 750°C isotherm is located at about 25 km depth and 80 km depth respectively in the Puna and Altiplano lithospheric sections, with Moho temperatures of the order of 1000°C and 600°C respectively (Fig. 5a, 6b).

Finally, following the discussion of section 2.3, we assume that the density is 50 kg/m^3 greater in the Altiplano mantle than in the Puna mantle, from 65 to 115 km depth. In other words, the model incorporates the asthenosphere from 115 km to 150 km depth in the Altiplano 'column', and from 65 km to 150 km depth in the Puna 'column'. This extra density is balanced in the vertical direction by a 1 km higher topography in the Puna (Fig. 4). The model is thus, at initial state, equilibrated in the vertical direction (i.e. classical isostatic equilibrium), and the driving force for internal deformation will be caused by deviatoric stresses rising from unbalanced forces in the horizontal direction.

3.3 Results

We first describe results of the model in which lower crustal viscosity reaches 10^{19}Pa.s. We let the model evolve for 5 Ma, when the physical transient state is well established. Figure 5 displays vertical profiles within the Altiplano (dark plain lines) and within the Puna (dark dashed lines). Figure 6 shows 2D contour variables. Although crust and mantle rock compositions are respectively identical in the Altiplano and Puna columns, we use different colours (Fig. 6d) in order to track the deformation of these columns with respect to each other. Because of the different geotherm in both lithospheric columns (Fig. 5a, 6b), viscosity differs (Fig. 5b) and controls the 'position' and intensity of stresses (Fig. 5c) and deformation (Fig. 5d, 6c, 6d). We describe stresses and deformations at the surface and at depth, which are linked to temperature and rock rheology.

a) The deviatoric shear stress (equivalent to $(\sigma_1\text{-}\sigma_2)/2$ with principal stresses) is highest in the upper crust and in the upper mantle (Fig. 5c, 6a), where rock rheology depends on pressure rather than on temperature. In these competent layers, deformation remains mostly elastic since the brittle yield strength is barely achieved: maximum shear stress remains of the same order as cohesion (chosen equal to 20 MPa). In contrast, deformation is ductile in the lower crust from about 20 to 65 km depth, with negligible shear stress. Deformation is also ductile in the mantle where temperature exceeds 750°C (see Fig. 6b for associated depths between 50 and 100 km).

b) From 20 to 65 km depth in the lower crust, velocity vectors are horizontal and exceed 1cm/yr (Fig. 5d, 6c), indicating a strong channelised flow from the Puna towards the Altiplano. This is associated to a shear strain rate (equivalent to $(d\varepsilon_1/dt\text{-}d\varepsilon_2/dt)/2$ with principal values) greater than $10^{-14}s^{-1}$ (Fig. 6c). Velocities in the competent upper crust and upper mantle are at least an order of magnitude smaller, and strain rates about 2 orders of magnitude smaller. Surface deformation is thus negligible. In the Puna mantle, relatively high strain rates and upward directed velocity vectors illustrate the link sketched in Figure 3, between pressure gradients at depth and crustal channel flow.

c) The 350°C isotherm (corresponding approximately to the crustal brittle-ductile transition) does not significantly move within 5 Ma, while the 750°C isotherm (corresponding approximately to olivine brittle-ductile transition) moves due to advection of material: at the boundary between both lithospheric columns this isotherm ascends at the Puna side and deepens in the Altiplano side (Fig. 5a, Fig. 6b).

d) Surface relief conserves the initial 1km step-topography in the center, because the upper crust is strong enough to sustain the 25 MPa associated to the step-topography (Fig. 6a). Flexure of the upper crust occurs adjacent to, and because of, the step-topography. The average topography difference between the two lithospheric columns reduces to about 800 m.

e) After 5 Ma in the model, 50 to 100 km of Puna lower crust has moved into the Altiplano crust, from 20 to 65 km depth (Fig. 6d). This material flow thickens the Altiplano crust in such a way that after 5 Ma a Moho depth difference of more than 5km depth has developped, with deepening in the Altiplano (~68 km), and thinning in the Puna area (~ 62 km).

3.4 Remarks on modelling assumptions and rheology

a) The viscosity of the lower crust controls the speed and amount of material displaced from the Puna to the Altiplano. Lower crustal flow in the first model M1 with viscosity as small as 10^{19}Pa.s., is about a third faster than in the second model M2 with viscosity greater than 10^{20}Pa.s. (Fig. 5b-d, grey lines, Fig. 7, grey line). One would expect that an order of magnitude increase in viscosity would lead to an order of magnitude decrease in velocity, but the numerical models here assume non-linear power-law rheology for the crust (power-law exponent equal to 2.4 and 3 respectively for the crust and mantle), and the minimal 10^{19}Pa.s. and 10^{20}Pa.s. viscosities are lower cutoff values to the temperature-dependent viscosity, that consequently define a different channel thickness (*h*, Fig. 5b).

b) Many geophysical studies (GPS, postseismic deformation, seismic waves indicating melts, thermo-mechanical models suich as Beaumont et al., 2001, Babeyko et al., 2002) suggest that the lower crust viscosity under high plateaus could be as low as 10^{17} Pa.s. With the same geotherm and power-law creep parameters of a wet quartzite, the lower crust viscosity between 45 and 65 km depth reaches about 10^{18} Pa.s., so that lateral flow after 5 Ma exceeds 2 cm/yr. In contrast, if we incorporate a dominant mafic granulite crustal composition from 50 km to 65 km depth, as Yuan et al. (2002) suggest, then the viscosity of the crust at these depths is greater than 10^{21}Pa.s. in both the Altiplano and Puna columns. This restricts channel flow to occur only from ~20 to 50 km depth. It is yet uncertain wether sufficient amounts of partial melts or eclogitization (see discussion Beck & Zandt, 2002; Babeyko et al., 2002) affect the large-scale behaviour of the lower crust.

c) Because the mean Altiplano elevation is lower than the average Puna elevation, the lithostatic pressure in the Puna is greater than in the Altiplano at every depth, and there is no way to obtain flow from the Altiplano towards the Puna. The fact that the Altiplano crust is thicker than the Puna crust could be explained by our modelling as the consequence of channelised crustal flow rather than a pre-existing feature.

d) The proposed scenario should not be confused with orogenic 'spreading' into adjacent areas of lower topography and thinner crust (e.g. Gratton, 1989; Bird, 1991; Assumpçao-Araujo, 1993). Stresses arising from topography contrasts are involved in the building of the Subandean belt, as part of global compressional tectonics of the Andes (e.g. Willet et al., 1993; Adam & Reuther, 1999). In the lowlands in the east-west direction, however, the geotherm is colder and the crust is thinner, so that the viscosity and the strength of the lower crust are likely to be greater. Therefore, horizontal ductile flow in the lower crust is prevented. Similarly mantle material in the lowlands is also shallower and colder, so that it is also competent: it strengthens the overall lithosphere, allowing it to sustain stresses rising from topographic differences.

e) The initial conditions of our modelling correspond to a situation in which horizontal body force desequilibrium is inserted in a sudden way, by assuming different density or rock composition in the Altiplano and in the Puna upper mantles. In nature, such differences would rather be generated by the relatively rapid delamination of mantle lithosphere under the Puna (Kay et al., 1994), or by the more progressive underthrusting of the Brasilian shield under the Altiplano (Isacks, 1988; Dorbath et al., 1993; Roeder, 1995; Beck & Zandt, 2002), rather than long-term pre-existing lithospheric heterogeneities. in our model, the progressive decrease in lower crustal velocities through time (Fig. 7) should then be adapted so that a progressive increase in density differences progressively triggers lower crust horizontal flow. We modelled a non-equilibrated situation because the present day structure of the Andes shows that they are not in equilibrum: by doing so we implicitely assumed that the lithosphere is capable of reajusting stress gradients in the vertical direction (isostasy) much faster than lateral stress gradients, and because of the perfect-fluid-like behaviour of the underlying asthenosphere.

f) If the dense Altiplano lithospheric mantle is of weaker rheology (wet olivine) and/or warmer, in such a way that it has a relatively low viscosity, then it could flow downwards driven by its negative buoyancy (gravitational instability, e.g. Houseman et al., 1981; Fleitout & Froidevaux, 1982; Houseman & Molnar, 1997; Pysklywec et al., 2000). Such a gravitational instability can induce in turn the development of lateral flow of Puna upper mantle towards the Altiplano, that replaces the 'falling' denser lithospheric mantle. This mechanism should be the subject of future investigations: it would be interesting then to quantify the net amount of material transfer between the entire lithospheric columns. However, geophysical data indicate that the mantle under the Altiplano has not yet developped such an instability (Beck & Zandt, 2002).

4 Conclusions

Although one cannot exclude density hterogeneities in the crust, geophysical data support the existence of density heterogneities in the mantle: we have explored with numerical modelling the consequences of this density heterogeneity on orogeny-parallel lower crustal flow. We obtain south-to-north flow from the Puna lower crust into the Altiplano lower crust, induced by colder and denser mantle lithosphere under the Altiplano. The velocity of this flow decreases very slowly, meaning that this transient phenomenon has a long characteristic time-scale: after 5 Ma it is still 1cm/yr in case the viscosity is as low as 10^{19} Pa.s. in a channel thickness of about 40 km.

Our modelling shows that such a mass transfer is limited to the lower crust, involving stresses low enough that it does not affect the more competent upper crust and produce significant displacements detectable on the surface. This mechanism suggests that although a similar amount of crustal material is brought into the Puna and Altiplano areas induced by the relative convergence of the Nazca and South American plates, part of this material may also move perpendicularly from the Puna segment into the Altiplano segment. Thus there is an additional volume of crust to be accommodated by surperficial shortening in the Altiplano section, while there is less volume to be shortened in the Puna section (Fig. 8). In the typical case of our model (Fig. 6), more than 50 km of lower crust flows from the Puna into the Altiplano, on a thickness of 40 km and within 5 Ma. This corresponds to an area of 1000 km^2 and to a volume of 4.10^5 km^3 if we assume that it occurs along the 200 km E-W width of the plateau. This mechanism may thus significantly contribute to lateral variations of E-W estimated shortening through different sections of the Central Andes.

Variations in length, width, and crustal thickness make it difficult to relate directly the dynamics of shortening and the evolution of crustal volumes. However, the problems encountered when extrapolating shortening deduced from geological reconstructions to deformation of the entire crust, together with heterogeneous seismological lithosphere, are the sign of some kind of decoupled deformation between the deep crust and the surface. Fully 3D studies are indeed necessary to establish satisfactorily mass transfers in the Central Andes.

Acknowledgements

We thank S. Lamb and M. Froidevaux for refering this manuscript and for offering complementary view points, as well as T. Sempéré for constructive advice.

References

Adam J., Reuther C.D., 2000. Crustal dynamics and active fault mechanics during subduction erosion. Application of frictional wedge analysis on to the North Chilean Forearc, *Tectonophysics 321*, 297-325.

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

Artyushkov E.V., 1973. Stresses in the lithosphere caused by crustal thickness inhomogeneities, *J. Geophys. Res., 78*, 7675-7708.

Assumpção, M., Araujo, 1993. Effect of the Altiplano-Puna plateau, South America, on the regional intraplate stress. *Tectonophys. 221,* 475-496.

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

Baby P., Hérail G., Lopez JM., Oller J., Pareja J., Sempéré T., Tufino D., 1989. Structure de la zone subandine de Bolivie: influence de la géométrie des séries sédimentaires antéorogéniques sur la propagation des chevauchements, *C. R. Acad. Sci.,Sér. II 309*, 1717-1722.

Baby P., Moretti I., Guillier B., Limachi R., Mendez E., Oller J., Specht M., 1995. Petroleum system of the northern and central Bolivian Sub-andean zone, in *Petroleum Basins of South America*, ed A.J. Tankard, S.R. Suarez, Welsink H.J., *AAPG Mem. 62*, 445-458.

Beaumont C., Jamieson R.A., Nguyen M.H., Lee B., 2001. Himalayan tectonics explained by extrusion of a low-viscosity crustal channel coupled to focused surface denudation, *Nature 414*, 738-742.

Beck S.L., Zandt G., Myers S.C., Wallace T.C., Silver P.G., Drake L., 1996. Crustal thickness variations in the central Andes, *Geology 24*, 407-410.

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

Bird P., 1991. Lateral extrusion of lower crust from under high topography, in the isostatic limit, *J. Geophys. Res. 96* (6), 10275-10286.

Brace W.F.,Kohlstedt D.L., 1980. Limits on lithospheric stress imposed by laboratory experiments, *J. Geophys. Res. 85*, 6248-6252.

Brasse, H., Lezaeta, P., Rath, V., Schwalenberg, K. Soyer, W. & V. Haak, 2002. The Bolivian Altiplano conductivity anomaly. *J. Geophys. Res., 107,* 10.1029/2001JB000391.

Burov E.B., Diament M., 1995. The effective elastic thickness of continental lithosphere: what does it really mean? *J. Geophys. Res. 100*, 3905-3927.

Coutand I., Roperch P., Chauvin A., Cobbold P.R., Gautier P., 1999, Vertical-axis rotations across the Puna plateau (northwestern Argentina) from paleomagnetic analysis of Cretaceous and Cenozoic rocks, *J. Geophys. Res., 104*, 22965-22984.

Coutand I., Cobbold P.R., de Urreiztieta M., Gautier P., Chauvin A., Gapais D., Rossello E., 2001. Style of history of Andean deformation, Puna plateau, northwestern Argentina, *Tectonics 20-2*, 210-234.

Cundall, P., and M. Board, 1988. A microcomputer program for modelling large-strain plasticity problems, *Numerical Methods in Geomechanics 6*, 2101-2108.

Dewey J.F., Lamb S.H., 1992. Active tectonics of the Andes. *Andean geodynamics, Tectonophysics 205*, 79-95.

Dorbath C., Granet M., Poupinet G., Martinez C., 1993. A teleseismic study of the Altiplano and the Eastern Cordillera in northern Bolivia: new constraints on a lithospheric model, *J. Geophys. Res. 98*, 9825-9844.

Elger K., Oncken O., 2002. The pattern of deformation related to growth of the Southern Altiplano plateau -Bolivia, *5th International Symposium on Andean Geodynamics, Toulouse, Extended Abstracts*, IRD Paris, 730 p.

Fleitout L., Froidevaux C., 1982. Tectonics and topography for a lithosphere containing density heterogeneities, *Tectonics 1*, 21-57.

Froidevaux C., Isacks B.L., 1984. The mechanical state of the lithosphere in the Altiplano-Puna segment of the Andes, *Earth Planet. Sci. Lett. 71*, 305-314.

Gerbault M., Henrys S., Davey F., 2003. Numerical models of lithospheric deformation forming the Southern Alps of New Zealand, *J. Geophys. Res*. 108, 2341, doi:10.1029/2001JB001716.

Giese P., Scheuber E., Schilling F.R., Schmitz M., Wigger P., 1999. Crustal Thickening Processes in the Central Andes and the Different Natures of the Moho-Discontinuity, *J. S. A. Earth Sci. 12*, 201-220.

Gratton J., 1989. Crustal shortening, root spreading, isostasy, and the growth of orogenic belts: a dimensional analysis, *J. Geophys. Res. 94*, 15627-15634.

Haberland C., Rietbrock A., 2001. Attenuation tomography in the western cnetral Andes: a detailed insight into the structure of a magmatic arc, *J. Geophys. Res. 106*, 11151-11167.

Hamza V.M., Muñoz M., 1996. Heat flow map of South America, *Geothermics, 6*, 599-646.

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

Houseman, G. McKenzie, D.P., Molnar P., 1981. Convective instatibility of a thickened boundary layer and its relevance for the evolution of continental convergent belts. *J. Geophys. Res. 86*, 6115-6132.

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

Hyndman R.D., Lewis T.J., 1999. Geophysical consequences of the Cordillera-Craton thermal transition in southwestern Canada, *Tectonophys. 306*, 397-422.

Isacks BL., 1988. Uplift of the Central Andean plateau and bending of the Bolivian Orocline, *J. Geophys. Res. 93*, 3211-3231.

Jaillard E., Herail G., Monfret T., Diaz-Martinez E., Baby P., Lavenu A., Dumont J.F., 2001. Tectonic evolution of the Andes of Ecuador, Peru, Bolivia and Northernmost Chile. In: *Tectonic evolution of South America, 31st International Geological Congress, Rio de Janeiro, Brazil 2000* (U.G. Cordani, E.J. Milani, A. Thomaz Filho, D.A .Campos eds), 481-559.

Kay, S. Mahlburg, B. Coira, and J. Viramonte, 1994. Young mafic back-ark volcanic rocks as guides to lithospheric delamination beneath the Argentine Puna Plateau, Central Andes. *J. Geophys. Res. 99*, 24323-39.

Kley J., 1996. Transition from basement-involved to thin-skinned thrusting in the Cordillera Oriental of Southern Bolivia, *Tectonics 15-4*, 763-775.

Kley J., Monaldi CR.,1998. Tectonic shortening and crustal thickening in the Central Andes: how good is the correlation?, *Geology 26-8*, 723-726.

Kley J., Monaldi C.R., Salfity J.A., 1999. Along-strike segmentation of the Andean foreland : causes and consequences, *Tectonophysics 301*, 75-94.

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

Lamb S., Hoke L., Kennan L., Dewey J., 1997. Cenozoic evolution of the Central Andes in Bolivia and northern Chile. *Orogeny through time ,Geol. Soc. Spec. Pub. 121*, 237-264.

Lamb S., 2000. Active deformation in the Bolivian Andes, South America. *J. Geophys. Res. 105*, 25,627-25,653.

Lamb S., 2001. Vertical axis rotation in the Bolivian orocline, South America; 2, Kinematic and dynamical implications. J. *Geophys. Res. 106*, 26,633-26,653.

Lamb S., 2001. Vertical axis rotation in the Bolivian Orocline, South America; 1, Paleomagnetic analysis of Cretaceous and Cenozoic rocks. J. *Geophys. Res. 106*, 26,605-26,632.

Lezaeta P., Munoz M., Brasse H., 2000. Magnetotelluric image of the crust and upper mantle in the backarc of the Northwestern Argentinean Andes, *Geophys. J. Int. 142*, 841-854.

Lucassen F., Becchio R., Harmon R., Kasemann S., Franz G., Trumbull R., Wilke H.G., Romer R.L., Dulski P., 2001. Composition and density model of the continental crust at an active continental margin--the Central Andes between 21° and 27°S, *Tectonophysics 341*, 195-223.

Martinez C., 1980. Structure et evolution de la chaîne Hercynienne et de la chaîne Andine dans le Nord de la cordillère des Andes de Bolivie, *Travaux et Documents de l'O.R.S.T.O.M. 119*, Paris.

Medvedev S., C. Beaumont, 2001. Growth of Continental Plateaux: Channel Injection or Tectonic Thickening? *AGU Fall Meeting*, San Francisco, USA. EOS Transactions.

Mercier J.L., Sebrier M., Lavenu A., Cabrera J., Bellier O., Dumont J-F., Machare J., 1992. Changes in the Tectonic Regime above a subduction zone of Andean Type: the Andes of Peru and Bolivia during the Pliocene-Pleistocene, *J. Geophys. Res 97*, 11945-11982.

Myers S.C., Beck S., Zandt G., Wallace T., 1998. Lithospheric-scale structure across the Bolivian Andes from tomographic images of velocity and attenuation of P and S waves. *J. Geophys. Res. 103*, 21233-21252.

Petrini K., Connoly A.D., Podladchikov Y.Y., 2001. A coupled petrological-tectonic model for sedimentary basin

evolution: the influence of metamorphic reactions on basin subsidence, *Terra Nova 13*, 354-359.
Poliakov,A., Podladchikov Y., 1992. Diapirism and topography, *Geophys. J. Int. 109*, 553-564.
Pysklywec R., Beaumont C., Fullsack P., 2000. Modeling the behavior of the continental mantle lithosphere during plate convergence, *Geology 28,* 655-658.
Ranalli G. *Rheology of the Earth*, 2nd ed.,Chapman & Hall eds., 413p, 1995.
Rochat P., Herail G., Baby P., Mascle G., 1999. Crustal balance and control of erosive and sedimentary processes on the formation of the Altiplano, *C. R. Ac. Sci. Paris, Sér. II 328*-3, 189-195.
Rochat P., 2002. Structures et cinématique de l'Altiplano Nord-Bolivien au sein des Andes Centrales, *Geologie Alpine edited by the Laboratoire de Geoogie de l'universite de Grenoble, Memoire H.S. 38*.
Roeder D., Andean-age structure of the Eastern Cordillera(Province of La Paz, bolivia), *Tectonics 7*, 23-39, 1988.
Roeder D. Chamberlain RL., 1995. Structural geology of sub-andean fold and thrust belt in northwestern Bolivia. In: *Petroleum basins of South America,* eds A.J. Tankard, S.R. Suarez, H.J. Welsink, A.A.P.G. Memoir 62, Tulsa Oklahoma, 459-479.
Royden, L. H., 1996. Coupling and decoupling of crust and mantle in convergent orogens: implications for strain partitioning in the crust, *J. Geophys. Res. 101*, 17679-17705.
Royden L.H., Burchfiel C., King R.W., Wang E., Chen Z., Shen F., Liu Y., 1997. Surface deformation and lower crustal flow in Eastern Tibet, *Science 276*, 788-790.
Shen, F. L-H. Royden, and B-C Burchfiel, 2001. Large-scale crustal deformation of the Tibetan Plateau. *J. Geophys.Res. 106,* 6793-6816.
Schmid, S. M. and E. Kissling, (2000). The arc of the western Alps in the ligh²t of geophysical data on deep crustal structure, *Tectonics 19*,62-85.
Sempéré T, Carlier G., Soler P., Fornari M., Carlotto V., Jacay J., Arispe O., Neraudeau., Cardenas J., Rosas S., Jimenez N., 2002. Late Permian - Middle Jurassic lithospheric thinning in Peru and Bolivia, and its bearing on Andean-age tectonics. *Tectonophysics 34*5, 153-181.
Springer M., Förster A., 1998. Heat-flow density across the Central Andean subduction zone, *Tectonophysics 291*, 123-139.
Turcotte D., Schubert G., 1982. *Geodynamics, Applications of continuum physics to geological problems*, Wiley New York, 450pp.
Vandervoort, D.S., 1993, Non-marine evaporite basin studies, southern Puna plateau, Central Andes *[Ph.D. dissertation]: Ithaca, New York, Cornell University*.
Watts AB., Lamb S.H., Fairhead J.D., Dewey J.F., 1995. Lithospheric flexure and bending of the Central Andes, *Earth Planet. Sci. Lett. 134*, 9-21.
Willet S., Beaumont C., Fullsack PP., 1993. Mechanical model for the tectonics of doubly vergent compressional orogens, *Geology 21*, 371-374.
Whitman D., Isacks B.L., Chatelain J.L., Chiu J.M., Perez A., 1992. Attenuation of high-frequency seismic waves beneath the central Andean plateau, *J. Geophys. Res. 97*, 19929-19947.
Whitman D., Isacks B.L., Kay S.M., 1996. Lithospheric structure and along-strike segmentation of the Central Andean plateau: seismic Q, magamatism, flexure, topography and tectonics, *Tectonophysics 259*, 29-40.
Yuan X., S. V. Sobolev, R. Kind, O. Oncken, G. Bock, G. Asch, B. Schurr, F. Graeber, A. Rudloff, W. Hanka, K. Wylegalla, R. Tibi, Ch. Haberland, A. Rietbrock, P. Giese, P. Wigger , P. Röwer, G. Zandt, S. Beck, T. Wallace, M. Pardo, D. Comte, 2000. Subduction and collision processes in the central Andes constrained by converted seismic phases, *Nature 408,* 958-961.
Yuan X., Sobolev S.V., Kind R., 2002. Moho topography in the Central Andes and its geodynamic implications, *Earth Planet. Sci. Lett. 199*, 389-402.

Figure Captions

Figure 1: Relief of the Central Andes.

Figure 2: **a**) Whitman et al. (1992) seismic attenuation models in the Altiplano (**a**) and in the Puna (**b**). High attenuation related to continental mantle lithosphere is absent under the Puna. Location of the MT profile which supports this deep structure (see text for details), modified from Lezaeta et al.(2000) **c**) North-South vertical cross section showing the correspondence between relatively low topography in the Altiplano and thick lithospheric mantle lid (after Whitman et al., 1992, 1996). The hatched zone represents regions of high seismic wave attenuation.

Figure 3: **a**) Lithostatic pressure (P) is uncompensated at the base of columns of different crustal thickness (with density ρ_c), but can become progressively compensated at the base of heavier mantle lithosphere (along thickness *dh*) present below the thicker crust (density ρ_m greater than asthenospheric density ρ_a that lies under the thinner crust). P1 and P2 would be the lithostatic pressure for the Puna and the Altiplano columns respectively (modified from Bird, 1991). **b**) Sketch of a possible situation in the Puna-Altiplano longitudinal cross-section. Plain arrow indicates additional stress created by $d\rho$ along *dh*, and the dashed line is the resulting possible flow.

Figure 4: Model setup. the Puna and Altiplano lithospheric columns are sketched to the left and to the right of the modelled plate, respectively. Initial Moho depth is 65 km, free slip lateral boundary conditions, hydrostatic compensation at the base, and initial topography that allows vertical isostatic equilibrium. Densities in the crust and mantle are 2800 and 3300 kg/m3 respectively, with an additional 50kg/m^3 as shown on figure for the first 50 km of Altiplano mantle. Lamé's parameters λ=μ=3.10^{10} Pa, Mohr-Coulomb cohesion *So* = 20 MPa, friction angle ϕ =30°, dilatation angle ψ = 0°, creep-law parameters for quartzite *A* = 3.2 10^{-4} MPa^{-n}/s, *E* = 1.54 10^5 kJ/mol, *n* = 2.3 (crust), and for olivine *A* = 7 10^4 MPa^{-n}/s, *E* = .2 10^5 kJ/mol, *n* = 3 (mantle) (values taken in compilation by Ranalli, 1995).

Figure 5: Vertical profiles after 5 Ma. From left to right, **a**) temperature, **b**) effective viscosity ($\log_{10}$), **c**) deviatoric shear stress, and **d**) horizontal velocity as a function of depth. Dark lines: profiles for model M1 in which lower crust viscosity can be as low as 10^{19}Pa.s (at about 30-65 km depth). Along a profile at X=-50 km in the Puna section (dashed lines), temperature is high and general strength (deviatoric shear stress) is small. Along a profile at X=+50 km in the Altiplano section (plain lines), temperature is low but increases with time in the lower crust, and due to horizontal advection of temperatures at more than 1 cm/yr. High shear stress in the upper crust and the lithospheric mantle is due to pressure dependent rheology (maximum is 25 Mpa at depth 70 km) in response to strain in the other ductile layers (mainly the lower crust). Grey lines represent model M2 in which lower crust viscosity cannot be lower than 10^{20} Pa.s. for a profile along X=+50 km in the Altiplano section. In this model, the initial geotherm does not change significantly because of little horizontal displacements in the lower crust (less than 0.5 cm/yr).

Figure 6: Results after 5 Ma for M1, 2D contours cross sections. **a**) topography, **b**) 350°C and 750°C isotherms, **c**) shear strain rate (second invariant) and velocity vectors (local maximum is 2.8 cm/yr), and **d**) displacements of coloured layers. Note that the base of the model (asthenospheric material) undulates, as heavy Altiplano mantle tends to sink, while lighter Puna mantle tends to flow upwards to replace it (see text 3.4 f).

Figure 7: Evolution of the mean velocity in the lower crust with time, for the 2 models in which the viscosity is either limited to 10^{19}Pa.s (dark line, M1) or to 10^{20}Pa.s (grey line, M2).

Figure 8: Map-view sketch of the possible effect of south-north mass transfer in the Altiplano-Puna region: because of mass influx in the Altiplano, more material (the ideal 50 km on the figure) needs to be accommodated (shortened) than that resulting from the relative displacement between the forearc and the Brazilian craton (the ideal 100 km on the figure). In turn, less material is to be accomodated (shortened) in the Puna. The numbers correspond to an equivalent amount of shortening which depends on local crustal thicknesses and relative lengths and widths, and not to a strict displacement-volume relationship. Because of the extended N-S length of the Altiplano, the equivalent shortening amount in the Altiplano should be smaller than 50 km.

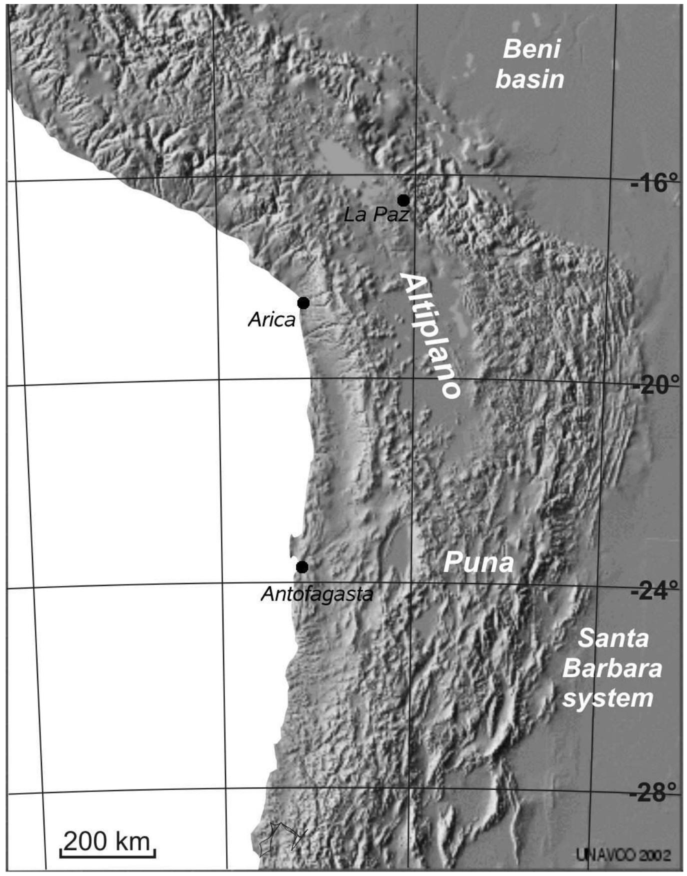

Beni
basin
-16°
La Paz
Altiplano
Arica
-20°
Puna
Antofagasta
-24°
Santa
Barbara
system
-28°
200 km
UNAVCO 2002

FIGURE 1

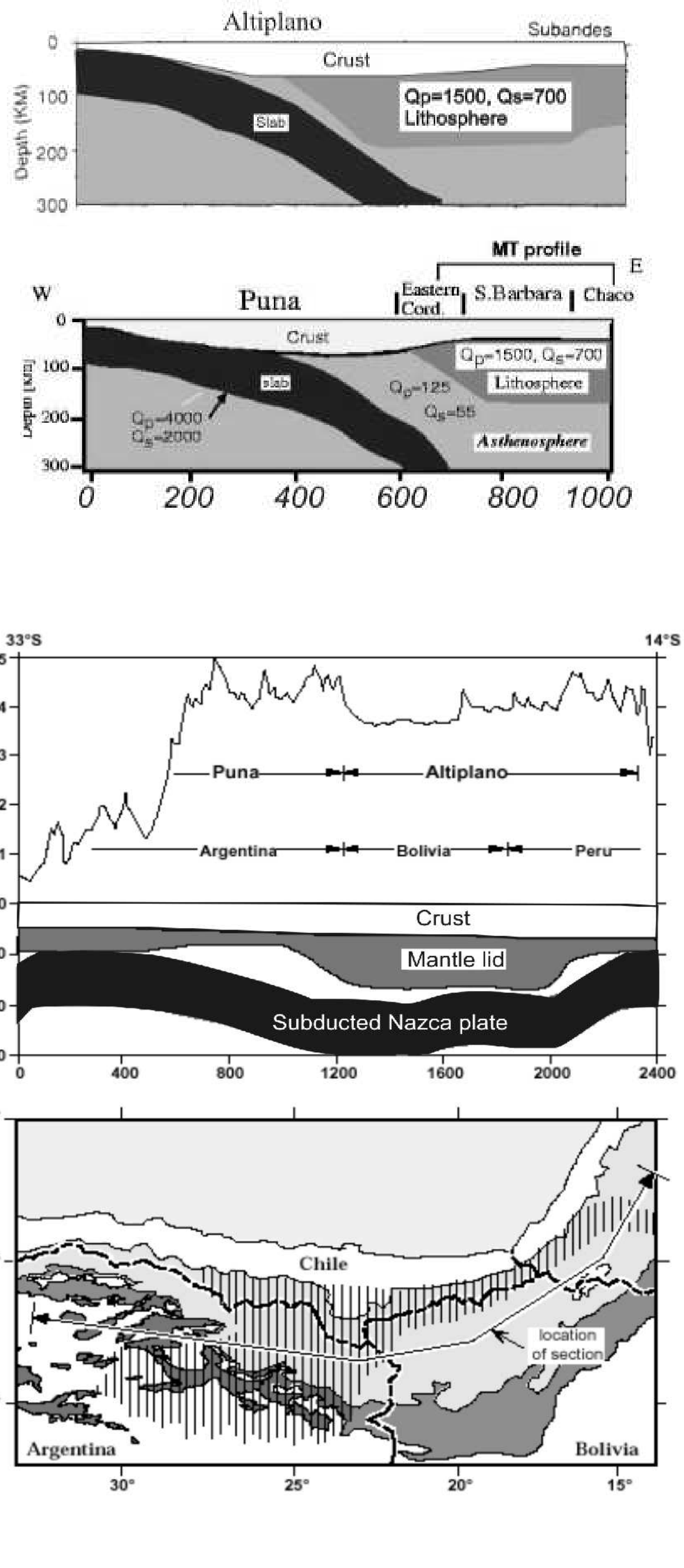
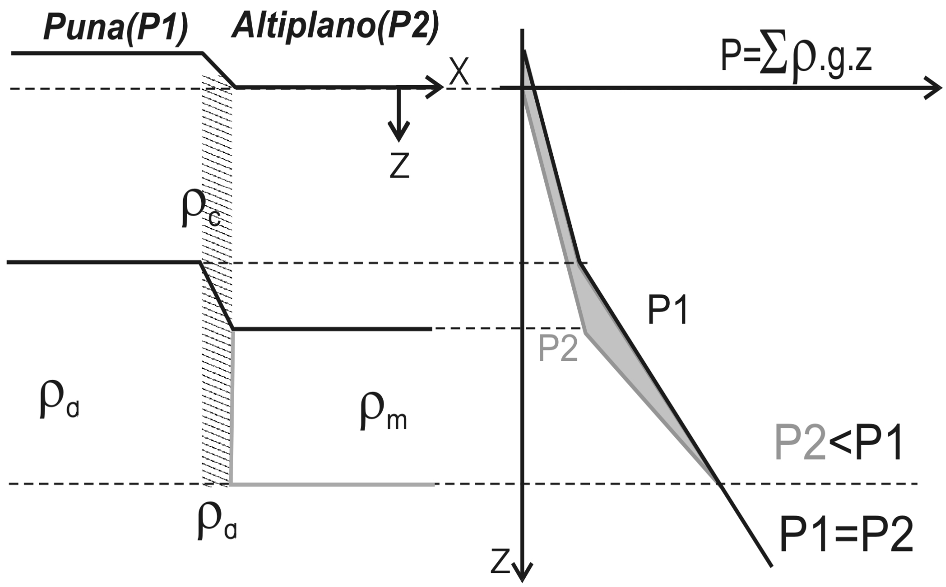

Altiplano
Subandes
Crust
Qp=1500, Qs=700
Lithosphere
Slab
Depth (KM)
0
100
200
300
MT profile
W
E
Puna
Eastern Cord.
S.Barbara
Chaco
Crust
Qp=1500, Qs=700
Lithosphere
slab
Qp=125
Qs=55
Qp=4000
Qs=2000
Asthenosphere
0
200
400
600
800
1000
33°S
14°S
Puna
Altiplano
Argentina
Bolivia
Peru
Crust
Mantle lid
Subducted Nazca plate
0
400
800
1200
1600
2000
2400
75°
70°
65°
Chile
location of section
Argentina
Bolivia
30°
25°
20°
15°

FIGURE 2 -abc


Puna(P1)
Altiplano(P2)
X
Z
ρc
ρa
ρm
ρa
P=∑ρ.g.z
P1
P2
P2<P1
P1=P2
Z

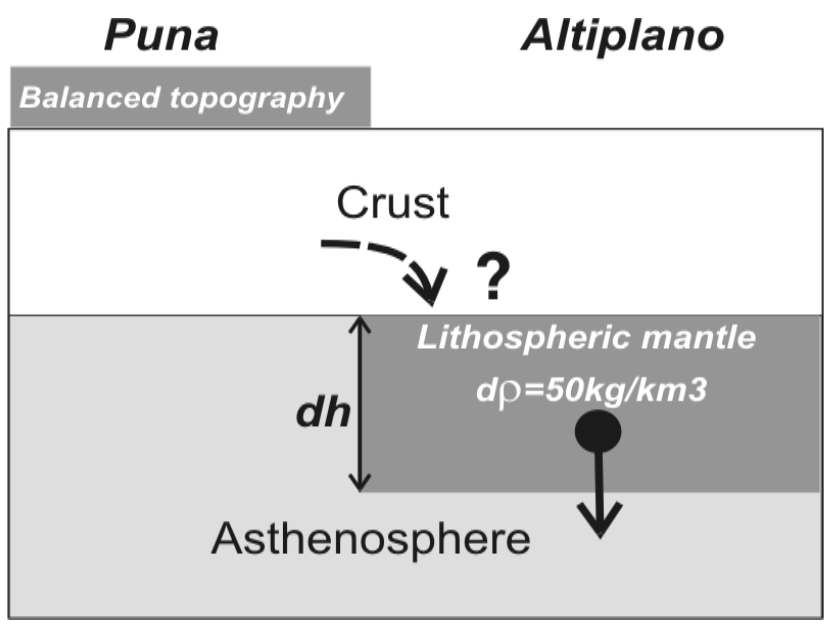

Puna
Altiplano
Balanced topography
Crust
?
Lithospheric mantle
dρ=50kg/km3
dh
Asthenosphere

FIGURE 3 -ab

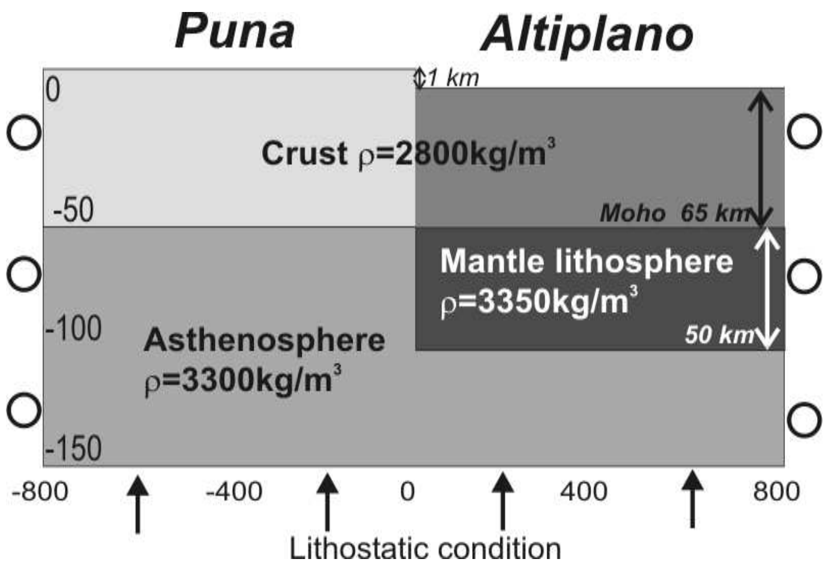


FIGURE 4

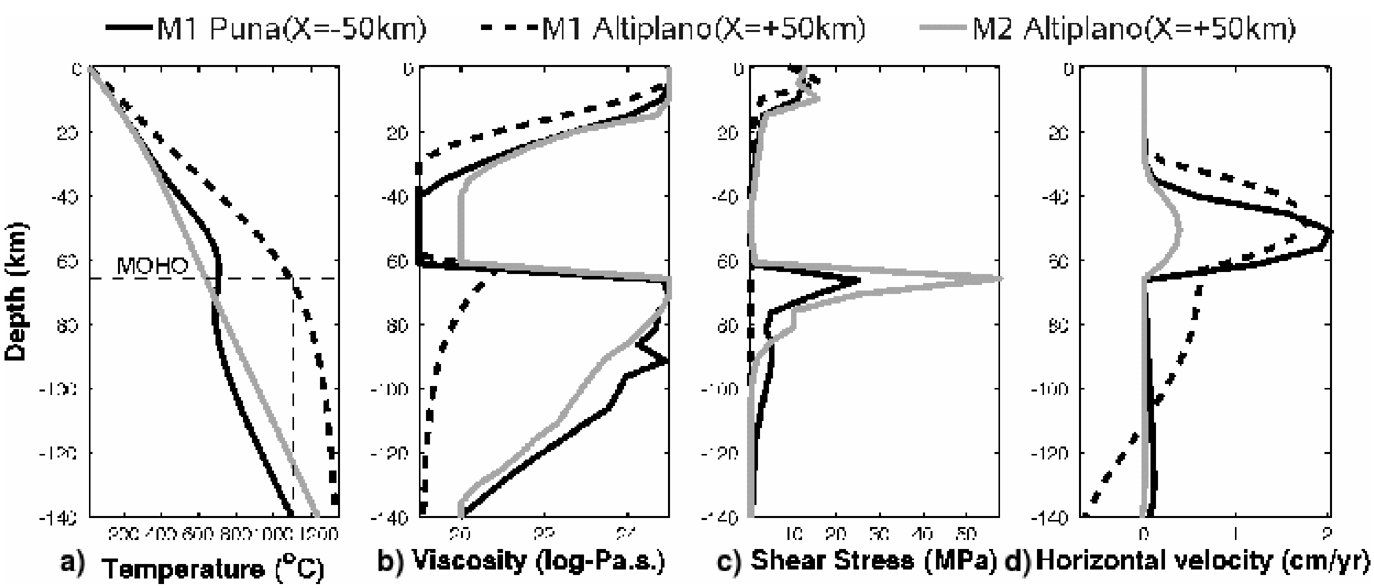


FIGURE 5

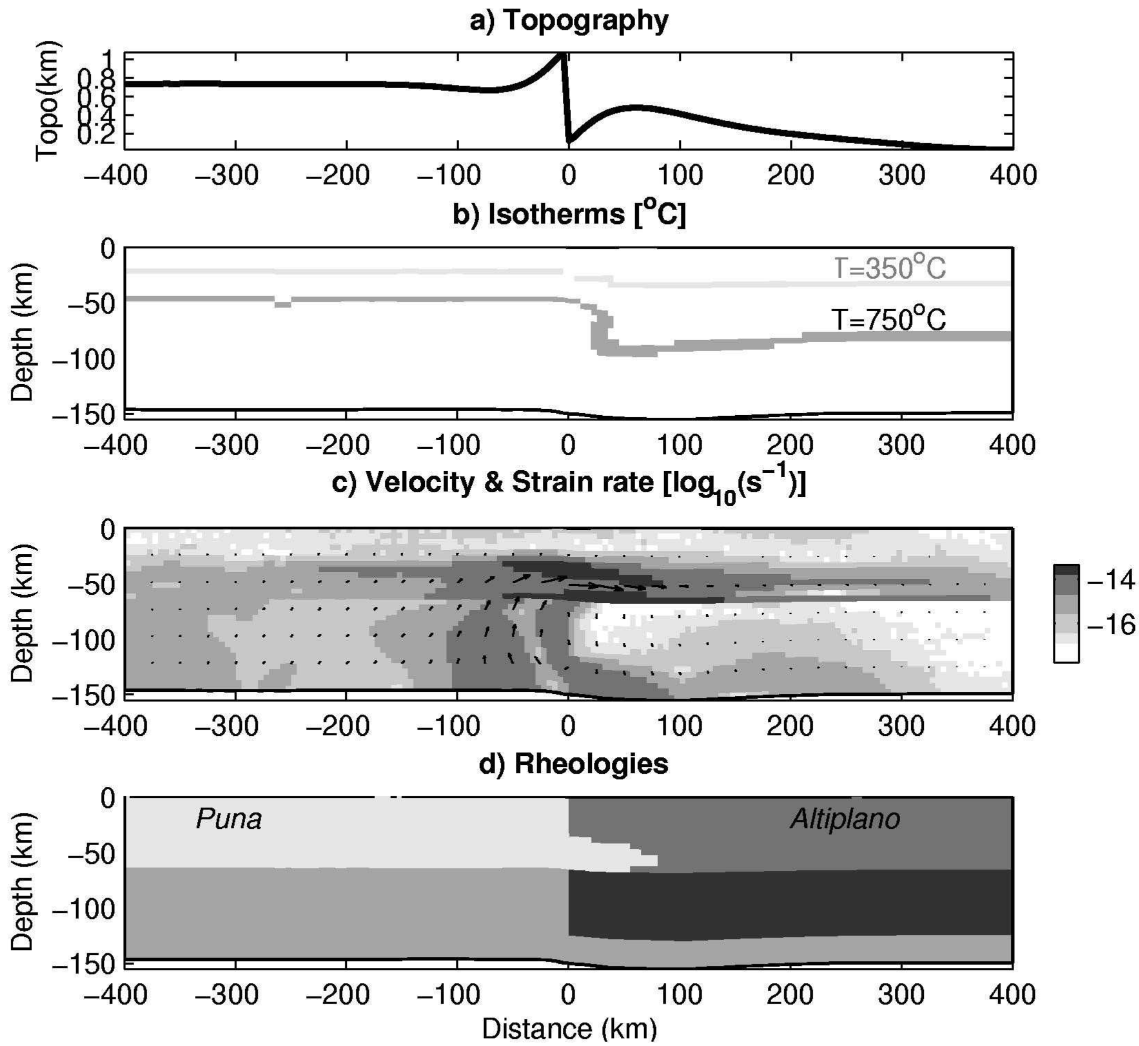

a) Topography
Topo(km)
b) Isotherms [^{o}C]
T=350^{o}C
T=750^{o}C
Depth (km)
c) Velocity & Strain rate [$\log_{10}(s^{-1})$]
−14
−16
d) Rheologies
Puna
Altiplano
Distance (km)

FIGURE 6

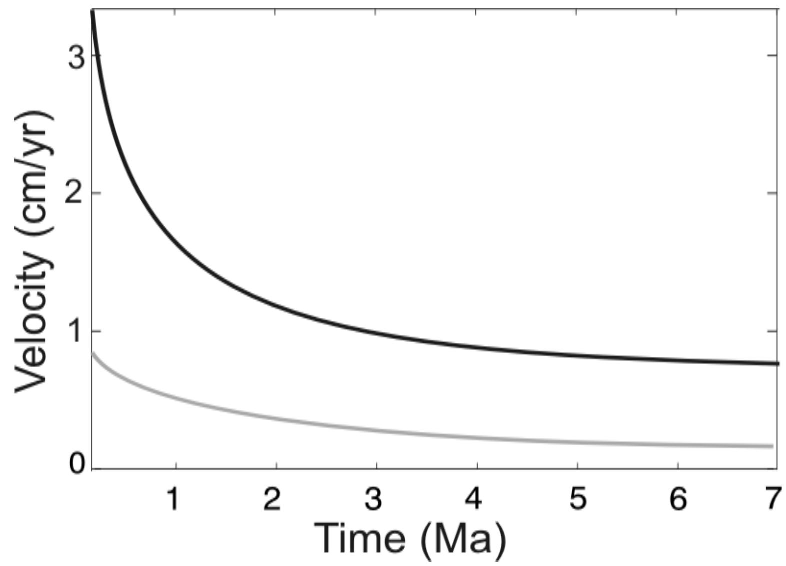


FIGURE 8

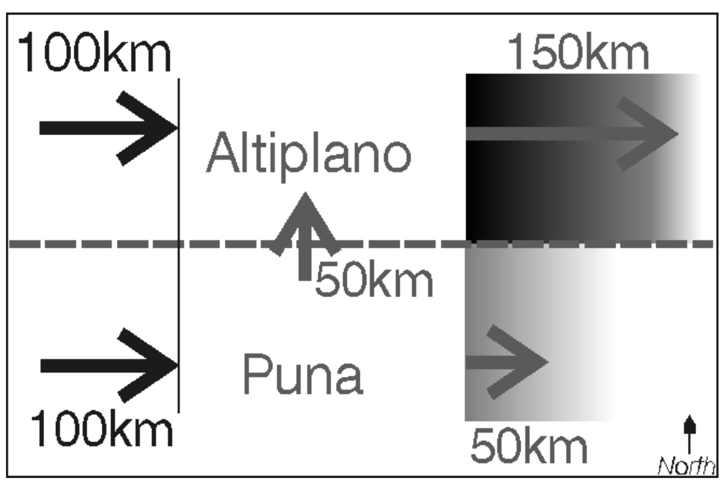


FIGURE 9