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A 3-D geodynamic model of lateral crustal flow during Andean mountain building

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Abstract. Although the Andes are believed to have resulted mainly from crustal shortening, the shortening history remains debated and appears to require lateral (along-strike) crustal flow. Three-dimensional viscous flow modeling shows that, within geological uncertainties, the Andes may have been produced by either Neogene shortening alone or with significant pre-Neogene shortening. These scenarios require major along-strike crustal flow and predict significantly different histories of uplift and crustal motion.

1. Introduction

The Central Andes in northern Chile, Bolivia, and southern Peru form the widest portion (600~700 km) of Earth's second largest mountain belt with an average elevation of ~4 km and a nearly doubled crustal thickness (~50-75 km) [*Allmendinger et al.*, 1997] (Fig. 1). Most workers believe that the Andes resulted mainly from Neogene crustal shortening associated with subduction of the Nazca plate under the South American plate [*Isacks*, 1988; *Kley et al.*, 1999]. However, Neogene crustal shortening appears insufficient for building the Andes, and its large along-strike variations appear to require lateral crustal flow to produce the relatively smooth Andean topography [*Kley and Monaldi*, 1998 ; *Hindle et al.*, 2000; *McQuarrie*, 2002]. Additional mechanisms of crustal thickening have been proposed, including tectonic underplating [*Baby et al.*, 1997] and magma addition [*Reymer and Schubert*, 1984], but their contribution is difficult to constrain.

Alternatively, there could have been significant pre-Neogene crustal shortening, especially in the western Cordillera and fore-arc where the early geologic record of shortening is not well preserved [*Horton et al.*, 2001]. *McQuarrie* [2002] suggested that mountain building in the Bolivian Andes started as early as 70 Ma, and that total shortening north of 23°S may be as high as ~530 km, which would have resulted in significant southward flow of crustal material.

However, the role of lateral crustal flow has not yet been numerically assessed because it was not included in the two-dimensional models used previously. Hence we explore this issue using a three-dimensional finite strain model.

2. Crustal Shortening and Crustal Volume

To estimate the role of crustal shortening we consider a 35 km thick initial crust, similar to the present craton, 0.3 km above present sea level. Digital topographic data (NGDC Globe 1.0) yields the volume of the Andes above 0.3 km to be ~4.1×10^6 km^3 (Fig. 1). Assuming Airy isostasy and average crust and mantle densities of 2830 - 2870 and 3300 kg m^{-3}, the observed crustal volume added by mountain building is 29 - 31×10^6 km^3.

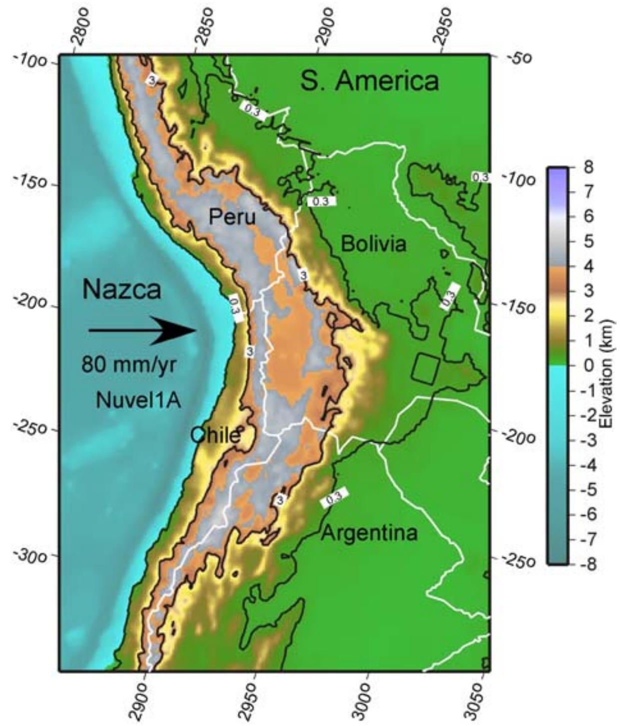

Figure 1. Topographic map of the Andes showing the 0.3 and 3 km contours. White lines are national borders.

This volume can be compared to that expected assuming a spatial pattern of shortening. Shortening estimated from Neogene deformation in the eastern thrust belts peaks around 250 km in the Central Andes [*Kley and Monaldi*, 1998]. The amount of post-Neogene shortening in the western Cordillera and the forearc is poorly known but likely limited, because the coastal regions have been dominated by extensional and/or transtensional deformation [*Hartley et al.*, 2000]. The western Cordillera, which probably uplifted to 2 km before the Oligocene [*Jordan et al.*, 1997], has been an active volcanic arc with at least some elevation attributable to magmatism [*Dorbath and Masson*, 2000]. Our upper estimate of shortening in the western Cordillera since the Neogene is <50 km, implying less than 300-km total shortening in the Central Andes since the Miocene.

The amount of shortening generally decreased from the Central Andes to the north and south, although there are large along-strike variations. Taking an upper estimate of 200-km shortening at 30°S [*Kley and Monaldi*, 1998], assuming symmetry at the northern end of the mountain belt, and smoothing along-strike variations, we derive a simplified

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initial geometry of the western margin of South America at ~25 Ma (model A in Fig. 2). In this model the area telescoped since 25 Ma is 730,000 km^2. This times the initial crustal thickness, taken as 38-42 km (i.e., slightly thicker than that for stable South America due to possible pre-Neogene thickening), yields an increased crustal volume about 28-31 x10^6 km^3, 90-107 % of that needed to build the present Andes. Hence within the uncertainties of geological data, Neogene crustal shortening may suffice to produce the present Andes.

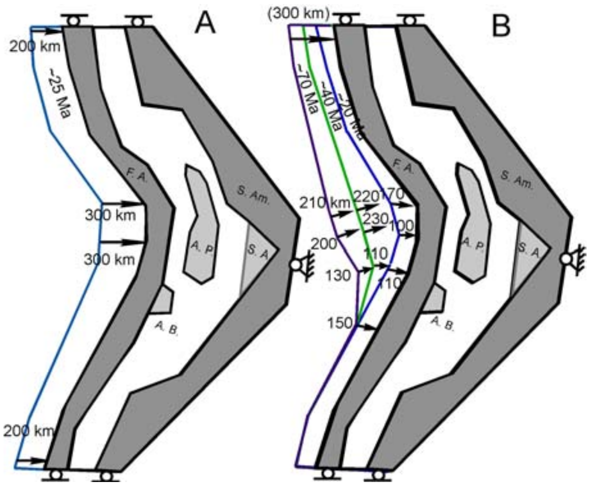

Figure 2. Model geometry (map view) and boundary conditions of the 3-D finite element model for two scenarios of crustal shortening and the associated initial geometry of the western margin of South America. In model A mountain building occurred mainly since Miocene. Model B includes significant pre-Miocene shortening. S. Am.: a transition zone between the Andean orogen and stable South America; F. A.: fore-arc; A. P.: Altiplano; A. B.: Atacama Basin; S.A.: sub-Andes.

McQuarrie [2002] proposed a longer shortening history that started ~ 70 Ma, with ~500-km of total shortening in the Bolivian Andes and ~150 km of shortening south of 25°S. However, the pre-Neogene shortening history is not well constrained. Based on her reconstructed shortening history we consider an alternative model (model B in Fig. 2) in which the area telescoped by crustal shortening is 890,000 km^2. Assuming an initial crustal thickness close to that of the present stable South America (32 - 35 km), the increased volume from shortening is 29-31x10^6 km^3, 94-107% of that required for the present Andes.

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3. Finite-strain model of Andean crustal shortening

Both scenarios imply significant along-strike crustal flow [*Kley and Monaldi*, 1998; *McQuarrie*, 2002]. We explore this issue using three-dimensional finite element models which include vertical variation of crustal rheology because most of the along-strike motion likely occurred in the ductile lower crust.

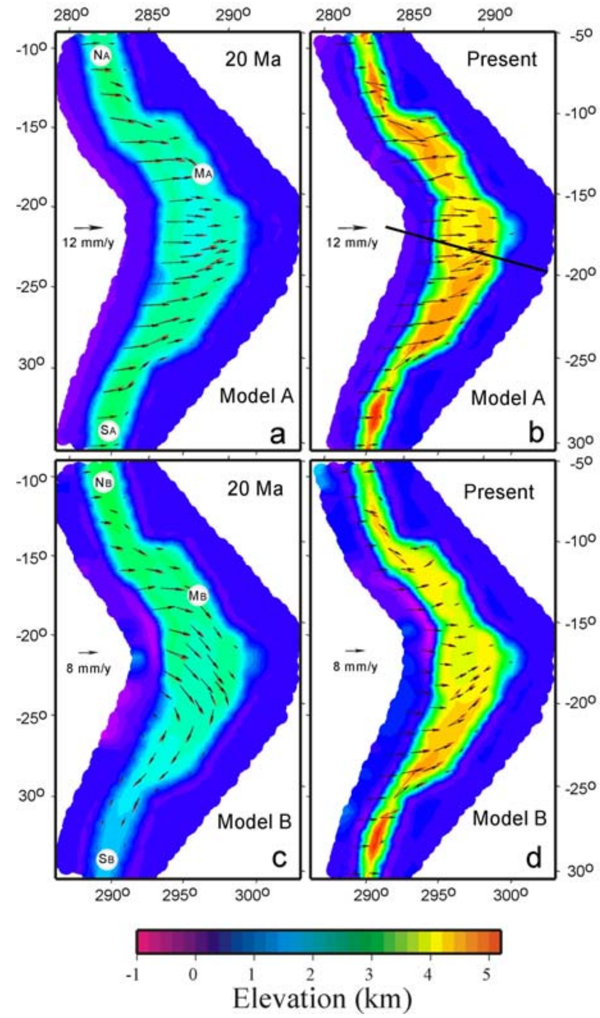

Figure 3. Snapshots of the predicted elevation (color background) and velocity field at the surface (black arrows) and the bottom (red arrows). (a-b) are for shortening model A (Fig. 2), and (c-d) are for shortening model B. Solid line shows the location of the profile in Fig. 4, and white circles with letters are locations whose uplift history is shown in Fig. 5.

The models are based on the shortening histories in Fig. 2. The model includes a 10-km think upper crust and two layers representing the middle and lower crust whose thickness changes with time. Winkler foundations are used on the surface and the bottom of the crust to simulate the load caused by change in elevation and the buoyant force arising from the crust root [*Williams and Richardson*, 1991]. The model assumes a viscous rheology (Newtonian) because elastic effects are negligible in long-term crustal deformation. The Andes are assumed to have developed in a rheological weak zone [*Isacks*, 1988; *Wdowinski and Bock*, 1994], with upper and middle-lower crustal viscosities of 1x10^{22} and 1x10^{20} Pa s. In some cases a stiffer rheology was used for the Altiplano Plateau and the Atacama Basin. For the fore-arc and the transition zone between the Andes and the stable South America, upper and middle-lower crustal viscosities are taken to be 1x10^{25} and 3x10^{22} Pa s, respectively. Stable South America is taken as rigid. Other rheologic structures were explored and their effects are discussed later. Model A assumes a constant shortening rate over the past 25 Myr. In model B shortening is divided into three periods [*McQuarrie*, 2002]; in which we assume a constant rate of shortening. Crust and the mantle densities are taken as 2800-2900 and 3300 kg m^{-3}. We calculated the finite strain using a numerical model built upon the commercial finite element package FEPG. Other model details are discussed in *Liu et al.* [2002].

Fig. 3 shows the predicted surface and lower crustal velocity and elevation. Model A (Fig. 3a and 3b) shows significant crustal motion from the north and south toward the Central Andes, although shortening was fastest there. This is because the initial width of the orogenic belt at the ends is much narrower than in the Central Andes, resulting in a higher strain rate there. Contributions from such along-strike crustal flow could explain the formation of the broad Central Andes with only moderate (300 km) crustal shortening there. Most of the along-strike crustal flow occurred in the middle-lower crust because of its low viscosity. The predicted surface velocity in the past 25 Myr, on the other hand, is roughly subparallel to the direction of plate convergence, consistent with the GPS velocities [*Kendrick et al.*, 2001; *Norabuena et al.*, 1998] and the geologically averaged surface velocity in the past 10 Ma [*Hindle et al.*, 2002].

In model B, large crustal shortening in the Bolivian Andes before 20 Ma causes significant southward crustal flow (Fig. 3c). The flow direction in the southern Andes reversed in the past 10 Myr (Fig. 3d) because in this model, crustal shortening in the south in the past 20 Myr (150 km) exceeds that in the Central Andes (~110 km; Fig. 2). The predicted present elevation and velocity fields are similar to model A.

Fig. 4 compares the predicted and observed elevation and crustal thickness in the Central Andes. A 40-km thick Andean crust at ~25 Ma produced a satisfactory fit to the observations (Fig. 4), as does a 35-km crust at ~70 Ma in model B. Because our models focus on the crustal volume and its distribution during the mountain building, they do not include the detailed crustal structures and erosion/deposition processes necessary to reflect some important aspects of Andean mountain building, including eastward migration of the deformation and its concentration in the sub-Andean belt [*Gregory-Wodzicki*, 2000; *Horton et al.*, 2001].

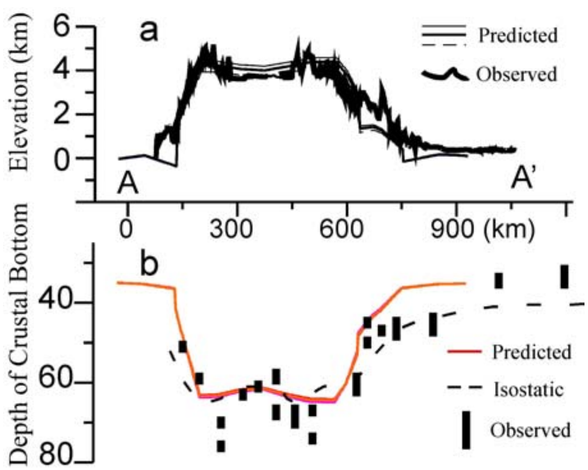

Figure 4. (a) Comparison of the predicted and observed long-wavelength elevation at 20ºS (location in Fig. 3a). Thin dashed, solid and thin solid lines are for crustal densities of 2890, 2870, and 2840 kg m^{-3}, respectively. Lower crustal density predicts higher elevation. (b) Comparison of predicated and observed crustal thickness [*Beck et al.*, 1996]. Differences due to crustal density are indistinguishable at this scale.

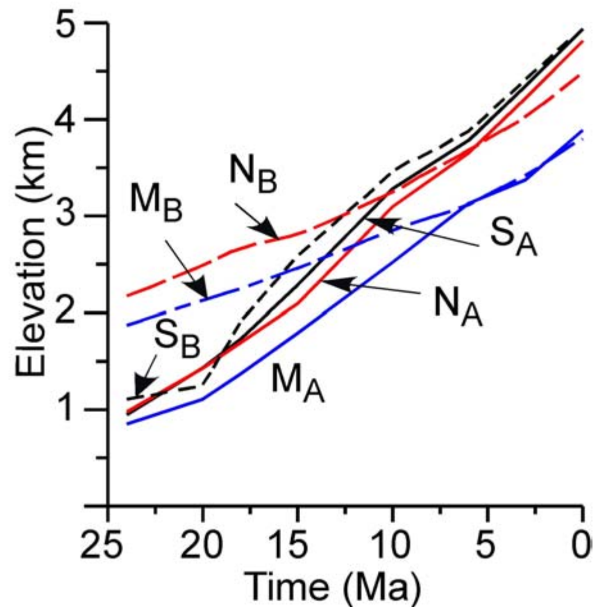

Figure 5. Predicted uplift history at the northern and southern limbs and the Central Andes (locations in Fig. 3a and 3c) for shortening models A (solid lines) and B (dashed lines).

4. Discussion

We find that, within the uncertainties of geological data, the present Andes could have been built by either Neogene crustal shortening or with significant pre-Neogene shortening. However, these scenarios predict significantly different histories of mountain building (Fig. 5). In model A, the southern and northern limbs uplifted simultaneously. In model B, the northern limb rose to >2 km by 25 Ma while the southern limb was <1 km then, but rapidly uplifted in the past 20 Myr. However, the current paleo-elevation data are too sparse and their uncertainties too large [*Anders et al.*, 2002] for rigorous tests of these models.

In either model crustal shortening results in significant along-strike crustal flow. Similar crustal flow has been suggested for building the Tibetan Plateau [*Clark and Royden*, 2000]. The magnitude of the predicted crustal flow depends on the assumed crustal rheology. Matching the present topography requires the effective viscosity of the Andean lower crust to be around 10^{20} Pa s. The fit may be improved by assuming a relatively stiff Altiplano block and Atacama basin (Figs. 3-4).

Effective crustal flow requires a thermally weakened lower crust as assumed here. The Altiplano-Puna plateau is one of the most extensive ignimbrite provinces in the world, with high average heat flow (84 mWm^{-2}) [*Henry and Pollack*, 1988]. Seismic studies reveal low velocity and Poisson's ratio beneath the plateau, indicating a felsic crust [*Swenson et al.*, 2000]. Seismic wave attenuation is consistent with partial melting in the lower crust [*Yuan et al.*, 2000].

Although most crustal flow would occur in the lower crust and thus be difficult to observe, some along-strike motion may have been accommodated by strike-slip faults within the brittle upper crust. In the southern limb, the NNE-SSW striking faults are generally dextral in late Cenozoic, such as the Atacama fault zone [*Beck*, 1998] and the Liquine-Ofqui fault zone [*Cembrano et al.*, 2000]. In the northern limb, strike-slip faults such as the Cordillera Blanca of central Peru show predominately sinistral shear [*McNulty et* *al.*, 1998]. These observations are consistent with the results of both models of crustal shortening in the Late Cenozoic (Fig. 3). Strike-slip history for the Early Cenozoic, which may be helpful for testing the two shortening models, is not well constrained.

The model results may be compared with the available paleomagnetic data. Miocene rotation in the fore-arc is small (a few degrees), although some larger rotation, possibly related to deformation of local blocks, has been reported [*Roperch et al.*, 2000]. In model A the predicted rotation in the fore-arc is < 6º in the past 25 Myr. Model B predicts up to 40º clockwise rotation between 20ºS - 25ºS and no rotation south of 25ºS in the past 20 Myr. Further paleomagnetic studies can test these models of crustal shortening and mountain building.

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