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The evolution of deformation and topography of high elevated plateaus 2. Application to the central Andes

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Abstract. The central Andes form a wide elevated plateau flanked in the west by a steep slope toward the deep Chilean Trench and in the east by a gentle slope that subsides gradually toward the Brazilian Shield. The low elevated trench topography is dynamically supported, whereas the high Andean mountain topography is mostly isostatically supported by a thick crust. The last mountain building phase, which thickened the crust and formed the present-day Andes, began 26 m.y. ago, in the late Oligocene, with the increase of the convergence rate between the Nazca and the South American plates. We investigate the time evolution of the Andean deformation and topography by applying a temperature dependent viscoplastic flow model of continental lithosphere to the South American plate. The model predicts the observed present-day topographic profile across the central Andes, from the trench across the high Altiplano plateau to the Brazilian Shield. Our numerical results, combined with observations of the spatial and temporal distribution of igneous activity in the central Andes, lead us to conclude that the Altiplano developed and extended to its present width of 400 km as a result of thermal weakening of the lithosphere since late Oligocene until present. The model also predicts the observed eastward migration of the locus of the Andean crustal deformation with time. At early stages, both the crustal and mantle loci of deformation lie in the thermally weak region, which results in crustal thickening in this finite region. At later stages, as the crust thickens, it induces increased buoyancy forces, which resist crustal thickening beyond 65 km. As a result, the locus of crustal deformation migrates eastward. The detachment of the crustal locus of deformation from that of the mantle can explain the observed change in deformation pattern from thick-skinned tectonism during early stages of the deformation to thin-skinned tectonism during the more recent stages.

Introduction

The Altiplano of the central Andes is the largest active high elevated plateau produced during ocean-continent collision. The plateau lies at an average elevation of nearly 4 km and is underlain by a thick crust, of almost twice the thickness of normal crust [James, 1971]. It is 1500 km long and 300-400 km wide and follows the shape of the South American western coast shoreline between the latitudes of 15°-27°S (Figure 1). The topography across strike is asymmetric: a steep western slope descends sharply toward the Chilean Trench and a moderate eastern slope subsides gradually toward the Brazilian Shield. The topographic transition along the western slope, from the Andean peaks at 4-6 km above sea level to the Chilean Trench at 5-7 km below sea level, which is common to the entire length of the Andean mountain belt, form one of the most pronounced topographic gradients on Earth's surface.

Recent studies suggest that the Andean crust has been mostly thickened as a result of crustal shortening [e.g., *Suarez et al.*, 1983; *Lyon-Caen et al.*, 1985; *Sheffels et al.*, 1986; *Sheffels*, 1990]. Correlations between Andean mountain building phases and variations in the Nazca-South America convergence rate support this conclusion [*Pilger*, 1984; *Prado-Casas and Molnar*, 1987; *Sempere et al.*, 1990]. Furthermore, these studies suggest that the present-day Andean topography is largely a product of an increase in the convergence rate between the two plates that occurred 26 m.y. ago.

Several geophysical models have been proposed to explain the mechanical evolution of the central Andes. *Isacks* [1988] explained the Cenozoic evolution of the central Andes using a kinematic model of crustal shortening preceded by a thermal weakening event. He suggested that the deformation occurred in two phases: a widespread and pervasive horizontal shortening and vertical thickening of the crust, followed by a concentrated deformation along the eastern side of the Andes, where the upper crust is thrusted beneath the Subandes. However, his model is qualitative and therefore cannot generate quantitative results that can be compared with the observations. *Gratton* [1989] applied a scaling law to describe the growth of mountain belts and their topographic profiles and applied his model to the Andes. His model assumes vertically averaged rheology and produces a symmetrical mountain profile, which can explain the isostatically supported portion of the Peruvian Andes but does not produce the high elevated plateau topography that characterizes the Altiplano. *Wdowinski et al.* [1989] and *Wdowinski and O'Connell* [1991] used viscous flow models to investigate the observed Andean deformation in response to the subduction of the Nazca plate beneath the South American lithosphere. Their models take into account forces induced by the subducting Nazca plate but they are time independent and are concerned only with the present-day observed deformation.

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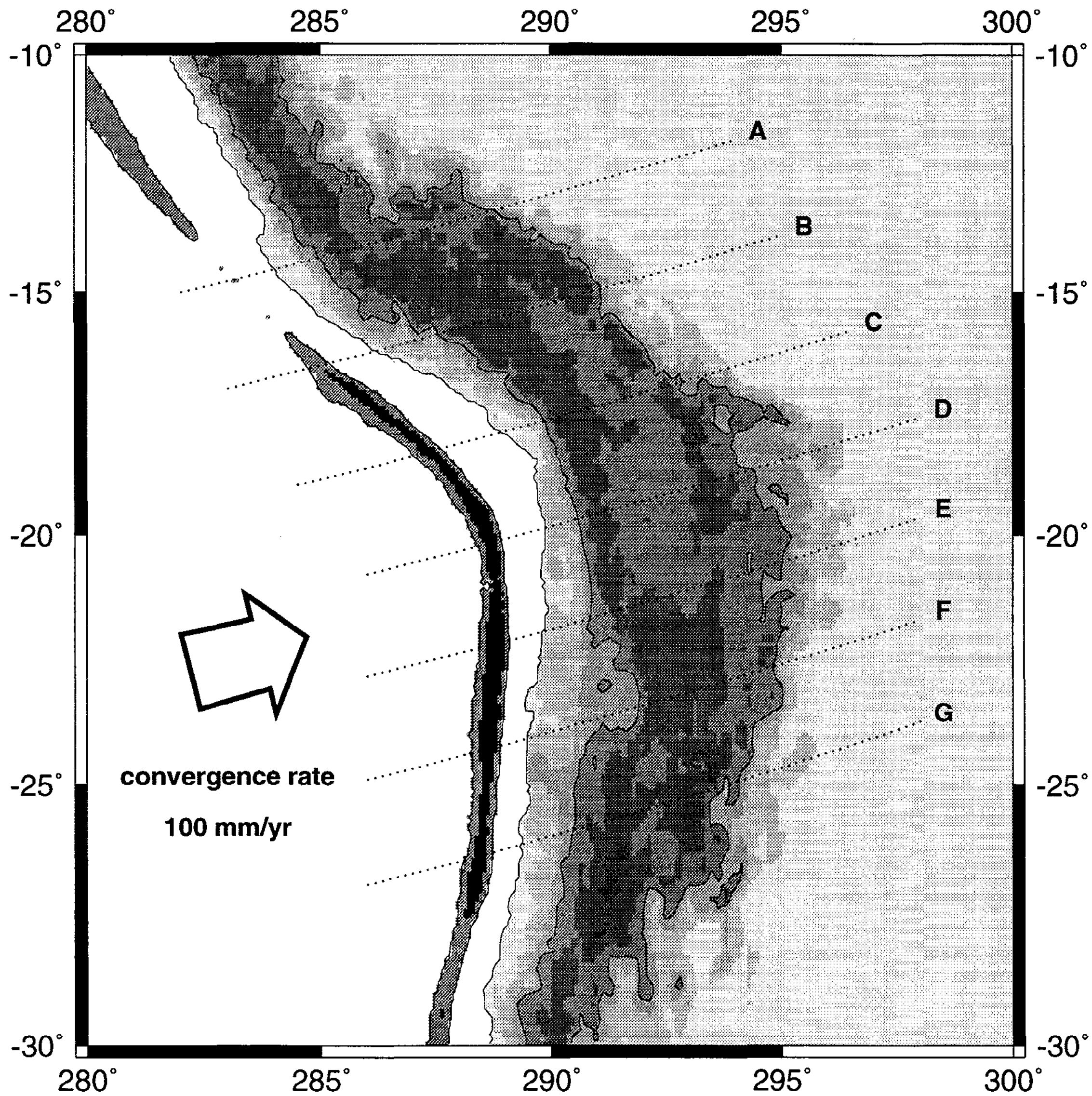

Figure 1a. Topography of the central Andes and bathymetry of the Chilean Trench, as generated from the data set ETOPO-5 [*National Geophysical Data Center (NGDC)*, 1988]. The gray scale topography shows 1-km elevation increments above sea level; the gray scale bathymetry shows 1-km elevation increments below -5 km.

In this study we investigate the time evolution of the central Andean deformation and topography by applying the temperature dependent viscoplastic flow model presented by *Wdowinski and Bock* [this issue] (hereafter referred to as paper 1) to the South American lithosphere. Our model includes both mechanical and thermal aspects of the deformation and predicts many aspects of the past and present-day observed deformation and topography. Furthermore, this is the first study that considers both the dynamically supported trench topography and the isostatically supported plateau topography, thus predicting the observed topography from the Chilean Trench to the Brazilian Shield.

Observations

The central Andean morphology is characterized by linear belts that are parallel to the South American-Pacific shoreline (Figure 1*a*). The main morphological units are: forearc region, Cordillera Occidental (magmatic arc), Altiplano, Cordillera Oriental, Subandes, and the Brazilian Shield (Figure 1*c*). The forearc region, whose topography lies mostly below sea level, extends from the Chilean trench to the foothills of the Andes. The landward topography continues to rise to the mountain tops of the Cordillera Occidental at more than 6 km above sea level. In between the Cordillera Occidental and the Cordillera Oriental lies a high elevated plateau, the Altiplano, at an average elevation of nearly 4 km. From the Cordillera Oriental the topography decreases gradually through the Subandean region to a few hundred meters above sea level at the Brazilian Shield.

Most of the high Andean topography is isostatically compensated by a thick crust [*James*, 1971], but a low-density asthenosphere may also contribute [*Froidevaux and Isacks*, 1984]. The one high region that is not in isostatic equilibrium is the Eastern Cordillera, where the topography is partly supported by a dynamic process [*Lyon-Caen et al.*, 1985;

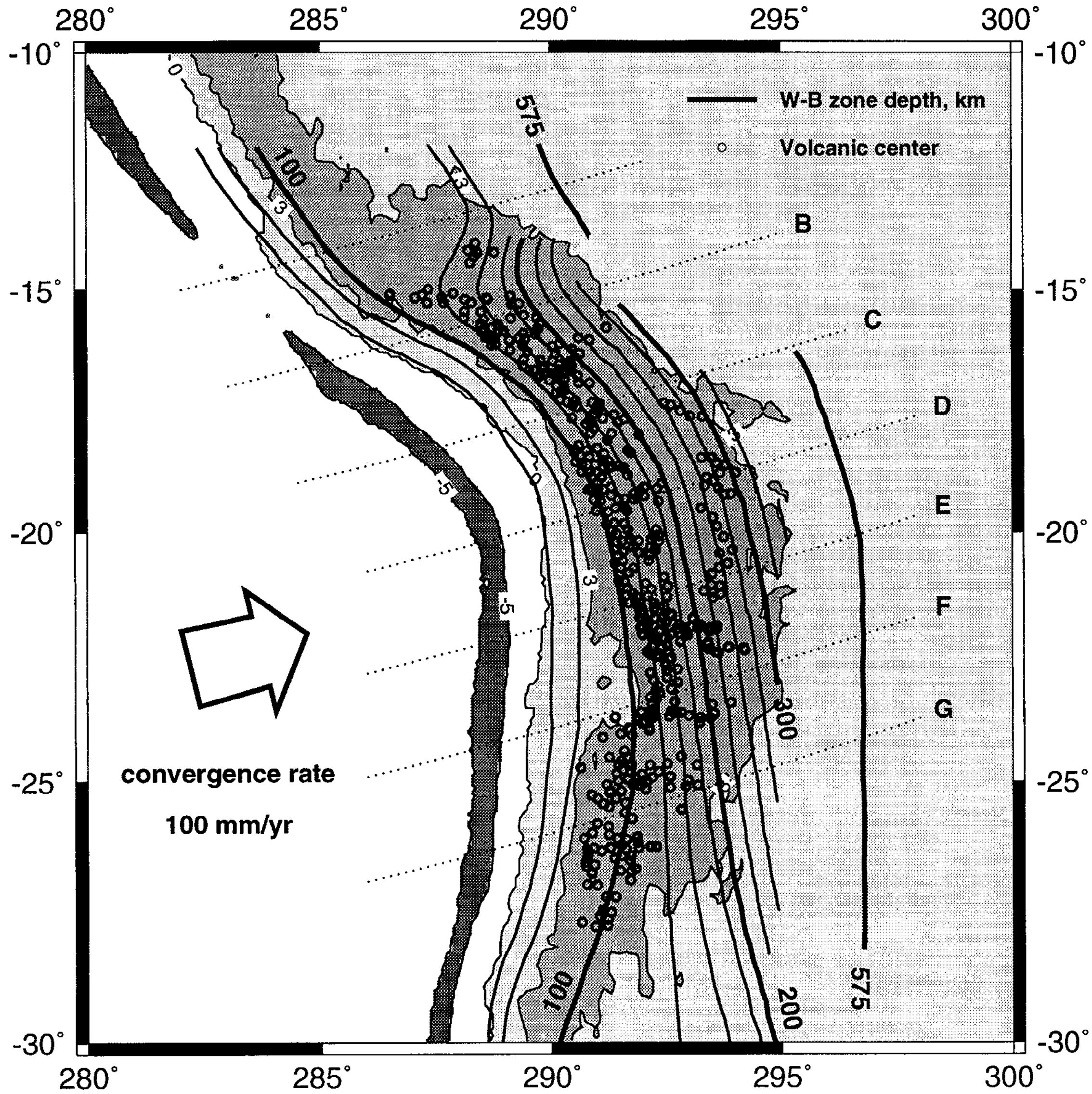

Figure 1b. Benioff-zone shape of the subducted Nazca plate, volcanic centers, area with average topography above 3 km, and area with average topography below -5 km [after *Isacks*, 1988].

Fuako et al., 1989]. Similarly, high negative free air anomalies observed in the vicinity of the trench indicate that the near-trench topography is mostly dynamically supported [*Wdowinski*, 1992]. The thickness of Andean crust has been determined by both seismic refraction and gravity surveys. Beneath the Brazilian Shield and Pacific coast the crust is 35-40 km thick, whereas beneath the Western Monocline and the Altiplano the crust is much thicker and exceeds 60-65 km [*Fuako et al.*, 1989; *Wigger et al.*, 1991].

The western margin of the South American plate is seismically very active. Most of the seismic energy is released as the subducting Nazca plate thrusts beneath the South American plate into the asthenosphere. Studies of the Benioff zone seismicity show along-strike segmentation of the subducting Nazca plate, which correlates with the morphological and tectonic segmentation of the Andes [*Stauder*, 1975; *Barazangi and Isacks*, 1976]. Beneath the central segment of the Andes (15°-27°S), seismic activity indicates that the Nazca plate is being subducted at an angle of 20° along the contact with the overriding South American plate (horizontal distance of 300 km from the trench to the 100 km contour). Beyond the 100-km contour, the Nazca plate continues to subduct at an angle of 30° to a horizontal distance of 700 km from the trench (Figure 1*b*). Only a very small portion of the observed seismicity originates within the Andean crust, and this mostly consists of thrust events in the Subandes region [*Suarez et al.*, 1983].

The present-day Andean deformation, as calculated from 17 years of seismic records, is confined to the Subandes region, which is located 700-900 km east of the trench, has a low-magnitude compressional component ($1\text{-}2\times10^{-16}$ s^{-1}) [*Suarez et al.*, 1983]. However, surface and subsurface geological structures suggest a complex pattern of deformation in the past 25 m.y. Fold and thrust structures, indicating a compressional tectonic regime, are found in the Eastern Cordillera and the Subandes regions [e.g., *Jordan et al.*, 1983; *Isacks* 1988]. Normal faults, representing an extensional tectonic regime, are found in the high Andes [e.g., *Dalmayrac and Molnar*, 1981] and in the forearc region [e.g., *von Huene*, 1988; *Sebrier et al.*, 1985]. Evidence for volcanic and magmatic

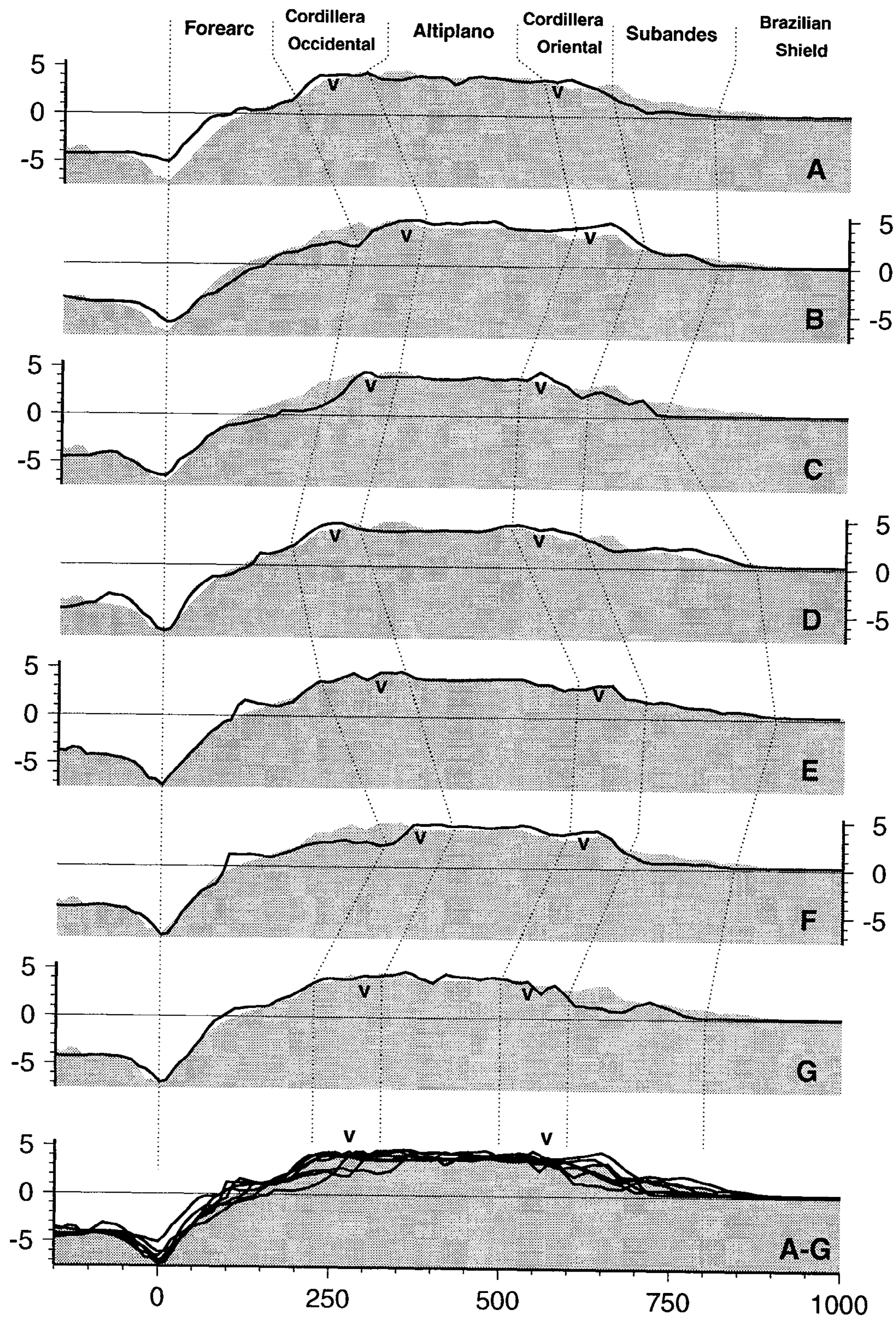

Figure 1c. Topographic profiles across the central Andes (solid lines) that are divided into the characteristic morphological units. Each profile is compared with the representative profile (shaded) to illustrate along-strike topographic variations. The symbol V represents the easternmost and westernmost location of volcanic centers. The geographic location of the profiles is shown in Figures 1a and 1b.

activity is found throughout the central Andes, from the Cordillera Occidental through the Altiplano to the Cordillera Oriental (Figures 1b and 1c). However, the present-day volcanic activity is concentrated in the Cordillera Occidental. Geochemical studies of these volcanic rocks show a strong crustal signature, suggesting that a significant portion of the magma was generated within the crust or interacted extensively with the crust [e.g., *Thorpe et al.*, 1981].

The last mountain building phase, which formed the modern Andes, started during the late Oligocene (about 26 Ma) [*Isacks*, 1988; *Sempere et al.*, 1990]. Structural relations suggest that the deformation started in the Cordillera Occidental and the locus of the deformation migrated eastward to the Eastern Cordillera and to the Subandes during the past 26 m.y. (Figure 2 after *Jordan and Gardeweg* [1989]). The early deformation stage, the "Quechua" phase, is characterized by shortening and crustal thickening of the Cordillera Occidental and the Altiplano regions [*Jordan and Gardeweg*, 1989], possibly by ductile deformation of the lower crust [*Isacks*, 1988]. The post-Quechua stage is characterized by a thin-skinned deformation style that produced the fold and thrust belt observed in the Cordillera Oriental and Subandes regions [*Jordan et al.*, 1983; *Jordan and Gardeweg*, 1990]. Numerous studies estimated the total amount of shortening to be in the range of 80-330 km [e.g., *Allmendinger et al.*, 1983; *Lyon-Caen et al.*, 1985]. Recently, *Sheffels* [1990] calculated a minimum 210 km shortening across the Bolivian fold and thrust belt from balanced cross sections.

The extension in the forearc region and in the high Andes is attributed to two different driving forces. In the forearc region the extension is explained by tectonic erosion, i.e., shearing and undercutting of the overriding plate by the subducting plate [*Coulbourn*, 1981]. In the high Andes the extension is attributed to buoyancy forces, which tend to relax the high Andean topography [*Dalmayrac and Molnar*, 1981]. However, the central segment of the high Andes (15°-27°S) does not appear to have been affected significantly by extension [*Jordan and Gardeweg*, 1989].

The rate of crustal shortening, which is defined by the relative velocity between the Brazilian shield and the trench, is estimated to be in the range 2-10 mm/yr. The lower estimate was calculated from the seismic deformation observed in the Subandes [*Suarez et al.*, 1983], whereas the higher estimate is the geologically averaged rate of shortening (250 km/25 m.y.) [*Isacks*, 1988]. The discrepancy between the two estimates may be due to several factors: (1) seismic deformation accounts for only 20% of the total deformation, (2) the present-day deformation is fully represented by the seismicity but is significantly lower than the geological average, or (3) these estimates are inaccurate due to lack of data. In the near future, accurate geodetic measurements will be taken in the central Andes and will provide us with direct observations of the present-day rate of shortening.

Comparison of Model's Results With Observations

In this section we compare the results of calculations from the model presented in the companion paper with available observations from the central Andes. The most reliable source of observations that can be compared with the calculations is the present-day topography. The present-day crustal structure and seismically derived strain rate field may also serve to constrain the model. Unfortunately, the past topography, crustal structure, and strain rate field are unknown, and therefore no quantitative comparison can constrain the calculated evolution of the central Andes. However, structural studies provide us with information regarding the deformation history of the central Andes, which can be qualitatively compared with the model's results.

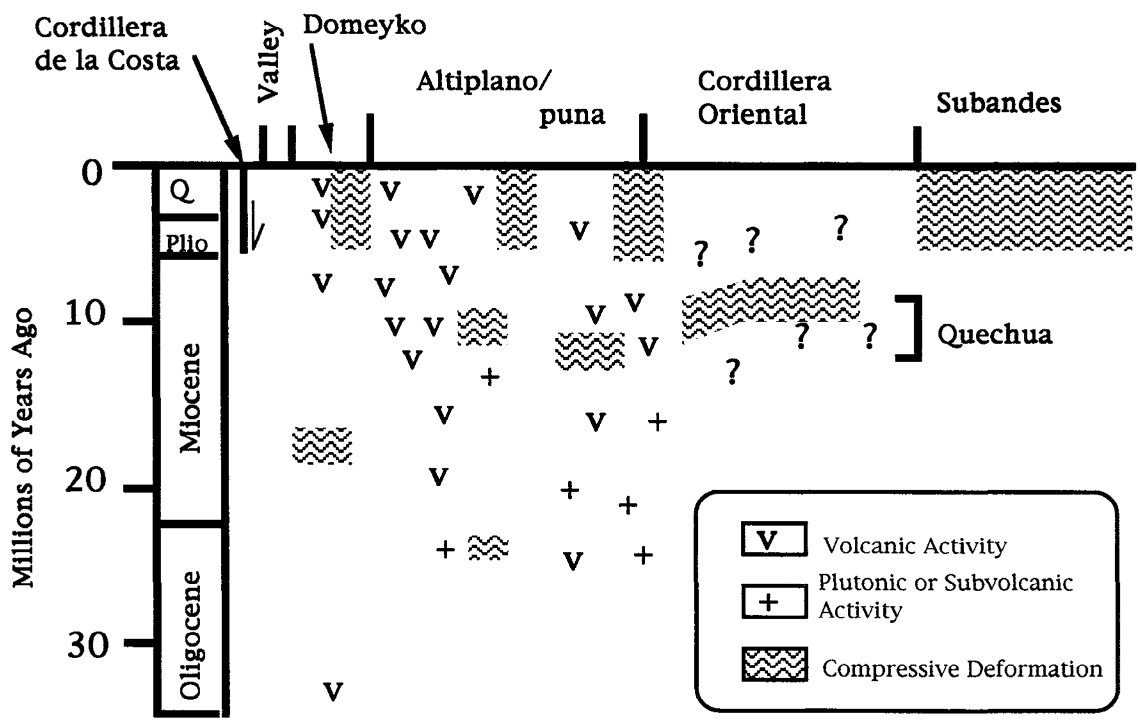

Figure 2. Spatial and temporal distribution of deformation and magmatic activity in the Central Andes [after *Jordan and Gardeweg*, 1989].

The model uses a temperature dependent viscoplastic rheology to investigate the evolution of deformation and topography of high elevated plateaus (see Figure 1 of paper 1). In order to avoid full three-dimensional computations, which are prohibitively complex, we assume that the lithosphere is infinitely long and that no flow variations occur along strike. This allows us to reduce the mathematical problem to two dimensions and to use a vertical plane strain formulation. One of the main conclusions of our modeling is that a finite region of localized deformation, a thick crust, and high topography develops only if the model includes a horizontal thermal perturbation or a finite region with an initially thick crust. However, only the thermally perturbed lithosphere generates a plateau topography. The shape and size of the plateau depend on the wavelength of the thermal perturbation, the Grashof number, and the density contrast between the crust and mantle. However, the topography of the plateau is insensitive to the amplitude of the perturbation, as long as it is larger than 5% (Figure 4c of paper 1); typically, we use a perturbed geotherm with 10% increase relative to the normal geotherm.

The elongated shape of the central Andes (15°-27°S, Figure 1a) and the relative insignificance of the along-strike topographic variations (Figure 1c) justify our use of the two-dimensional vertical plane strain model. However, the calculated topography is for the general case of the central Andes and therefore must be compared to a representative topographic profile and not to each of the profiles presented in Figure 1c. We choose to use a representative profile instead of an average profile because an average profile tends to smooth the topography and to erases sharp topographic transitions between the various domains. The representative profile (shaded in Figure 1c), which is basically profile E with minor modifications (see profile E in Figure 1c), reflects the sharp transitions between the morphologic units, although the transitions may be horizontally offset by up to 100 km relative to the actual profiles.

Figure 3 compares the observed characteristic topography (shaded) with three sets of calculated topographies at various time steps as the topography evolves with time. Two models based on a thermal perturbation (Figures 3a and 3b) predict the observed topography: below sea level trench topography, high elevated plateau with steep slope toward the trench and gradual slope inland, and close to sea level topography far away from the trench. However, the two models predict the present-day topography at two different dimensionless time units, which represent different amounts of indentation: 300 km for the initially uniform crust model, and 200 km for the initially thick crust model. As already shown in paper 1, the calculation for a third model with an initially thick crust without thermal perturbation predicts a monocline topography and not the observed high elevated plateau.

A remarkable result of the model is its prediction of the change in deformation pattern in the central Andes with time. During the early stages, the loci of the crustal and mantle deformation coincide with one another and are concentrated in the weak region that is thermally perturbed (Plates 1a and 1b). This deformation pattern simulates thick skinned tectonism, in which the crust uniformly thickens due to horizontal compression and as a result the topography is steadily uplifted (Figure 3a, time units 1, 2, and 3). At later stages, the locus of the crustal deformation migrates inland, whereas the locus of the mantle deformation remains near the wedge tip (Plates 1c and 1d). Detachment of the loci of compressional deformation accompanies a significant shear deformation that is concentrated in the weak lower crust. This deformation pattern simulates the thin skinned tectonism that characterized the formation of the fold and thrust belt in the Subandes during the post-Quechua phase. The model also predicts the present-day crustal structure and a low-magnitude compressional strain rate that deforms the Subandes region (Plate 1d). The magnitude of the calculated strain rate (1-4x10-16 s^{-1}) agrees well with the seismically derived strain rate calculated by *Suarez et al.* [1983].

Discussion

The modeling indicates that the high elevated plateau forms over the region subjected to a higher than normal geothermal gradient, and its width corresponds to the width of this region. In order to model the 400-km-wide Altiplano (Figure 3), we a priori impose a 400-km-wide thermal perturbation between points coinciding with the Cordillera Occidental and the Cordillera Oriental. An abnormally high geotherm beneath this region is evident from the widespread volcanic and plutonic activity that started in the late Oligocene and continued until present times (Figures 1 and 2). The spatial and temporal distribution of the igneous activity supports our assumption of a 400-km-wide region with a perturbed geotherm.

Isacks [1988] used a kinematic model of crustal shortening to explain the Cenozoic evolution of the central Andes and arrived at a similar conclusion. However, in his model, the thermal activity preceded the horizontal shortening and extended initially over a region of 600 km. As the lithosphere shortened, the width of the thermally weak zone decreased to 400 km. He suggested that the 600-km-wide region weakened as a consequence of the upper plate thinning during an episode of low-angle subduction, which later steepened to the present-day subduction geometry. We also attribute the thermal activity that weakened the crust beneath the central Andes to heat released by subduction zone processes. However, our model does not require a change in the subduction zone geometry because the width of the thermally perturbed region remains constant throughout the duration of the last mountain building phase.

Figure 3 shows two calculations that fit well with the characteristic topographic profile of the central Andes (solid line in Figure 3a and dash-dotted line in Figure 3b). Both calculations include a 400-km-wide thermal perturbation but differ in their initial crustal configuration. The first one (Figure 3a) assumes an initially uniform crust throughout the South American lithosphere and predicts the present-day topography after three time units (30 m.y.), which corresponds to 300 km of shortening. This result agrees well with the upper bound of crustal shortening calculated by *Lyon-Caen et al.* [1985]. The second calculation (Figure 3b) assumes an initially thick crust located above the wedge tip, similar to the present-day topography in other segments of the Andes, and predicts the present-day topography after two time units (20 m.y.), corresponding to 200 km of shortening. This result, which agrees with the recent shortening estimate of *Sheffels* [1990], represents a more realistic scenario for the central Andes topography and crustal structure during the late Oligocene, considering that this region was influenced by subduction processes since Jurassic time. An average shortening rate, which is the characteristic velocity (u_0, see

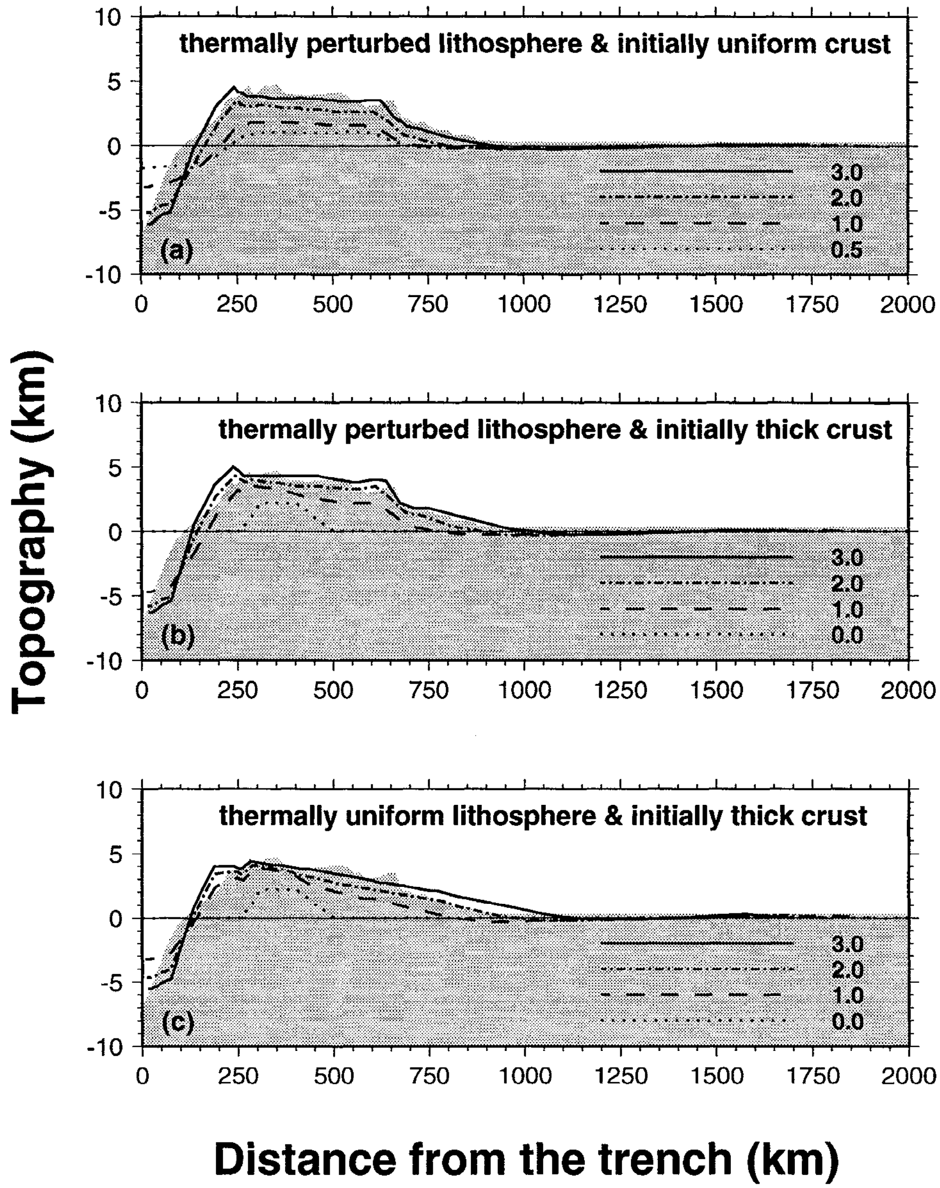

Figure 3. A comparison between the characteristic observed topographic profile (shaded) and calculated topographic profiles (obtained with Gr = 500 and $\tau_s = 6.5$). (a) Evolution of a thermally perturbed lithosphere that initially has a 35-km-thick uniform crust. (b) Evolution of a thermally perturbed lithosphere that initially has 50-km-thick crust above the wedge tip (300-400 km from the trench). (c) Evolution of a lithosphere with initially thick crust (same as in Figure 3b) without thermal perturbation. The details of the various models are presented in paper 1.

Table 1 of paper 1), can be estimated by dividing the total shortening by the duration of the last mountain building phase, which started 26 m.y. ago. We calculate an average shortening rate of 11.5 mm yr^{-1} for the case of initially uniform crust and 7.5 mm yr^{-1} for the case of an initially thick crust.

The central segment of the Andes (15°-27°S) is the only segment that contains a high elevated plateau, and therefore the model can be applied only to it. However, the other segments exhibit similar forearc topography, without the wide elevated plateau and the thick crust beneath it. Our model suggests that the forearc topography is controlled by a tectonic force that shears the base of the lithosphere parallel to the subduction direction, whereas the formation of the mountain topography is controlled by a horizontal component of the tectonic force. The similarity of the forearc topography all along the Andean mountain belt suggests that the parallel component is similar everywhere. However, the differences in the width of the Andean mountain belt reflect along-strike variations in the horizontal component.

According to the theory of plate tectonics, the convergence at subduction plate boundaries occurs due to subduction of only one plate, whereas the overriding plate remains undeformed and does not contribute any material to the asthenosphere. However, the diffuse deformation observed in the Central Andes suggests that the tectonics of Nazca-South American subduction zone diverges from the rules of the theory of plate tectonics. *Lyon-Caen et al.* [1985] treated the South American

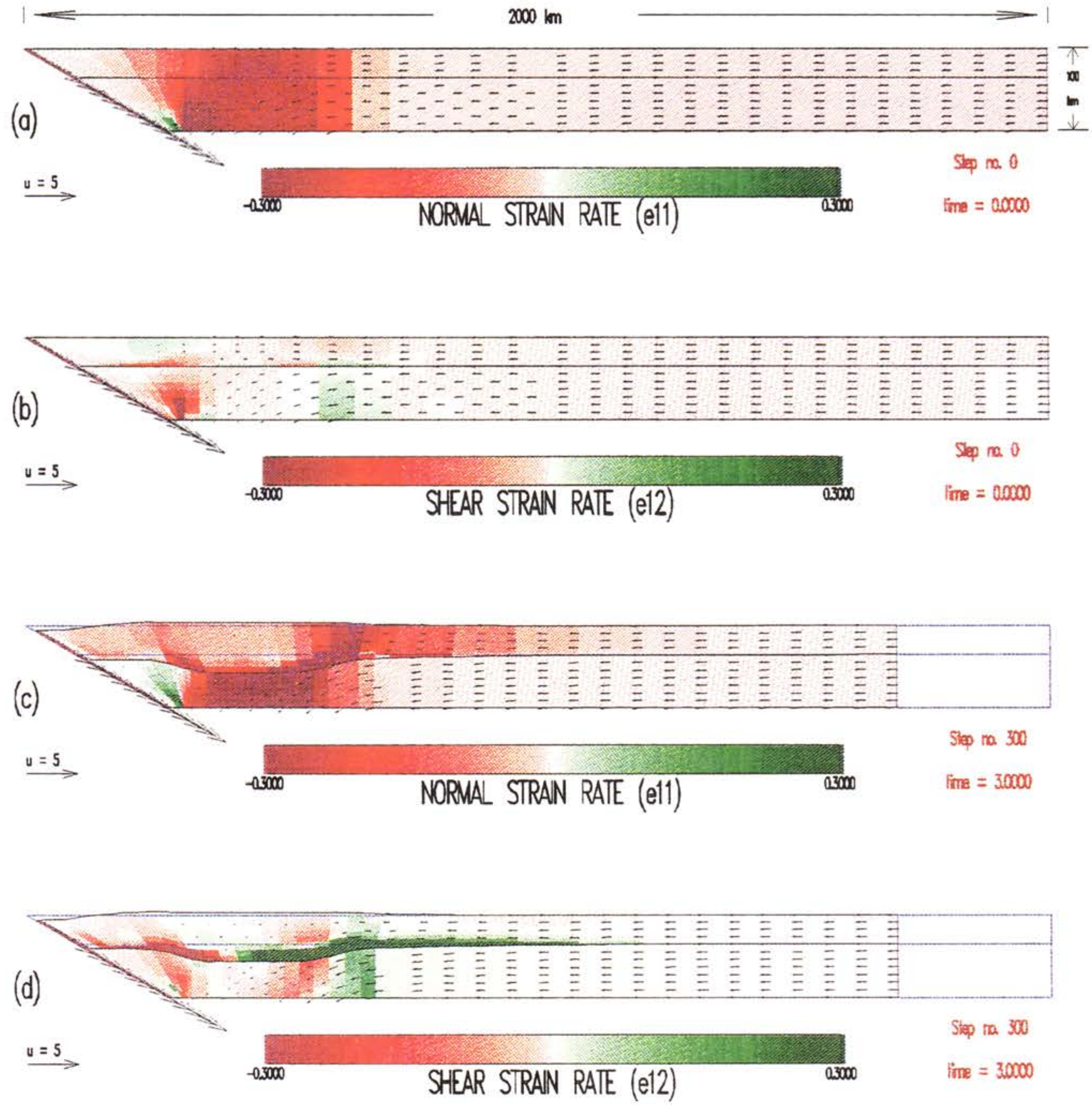

Plate 1. Calculated topography, crustal structure, normal strain rate (e11), and shear strain rate (e12) during the initial and final stages of the central Andes mountain building phase. These calculations consider an initially uniform crust and are obtained by using Gr = 500 and $\tau_s = 6.5$ (same as Figure 3a). The locus of compressional deformation (red) of the mantle lithosphere is localized near the wedge tip at all time steps. However, the compressional deformation of the crust diffuses with time and its locus migrates inland and is accompanied by a significant shear component (green) concentrated in the weak lower crust (details in paper 1).

lithosphere as being underthrusted beneath the Andes and assumed that the South American lithosphere behaves as an elastic plate. Recently, *Tao and O'Connell* [1992] treated subduction zones as viscous media and suggested that both the Nazca and the South American plates are subducting at a different rate. They called this phenomenon "ablative subduction". Our results also indicate that the South American lithosphere is partially lost to the asthenosphere. This partial subduction serves to mechanically differentiate between crust and mantle and plays an important role in conservation of crustal material near Earth's surface, while the continental mantle can be recycled.

The good fit between the observed and calculated topography, crustal structure, strain rate field, and style of deformation suggests that the temperature dependent viscoplastic flow model simulates well the evolution of the central Andes. However, the model uses some assumptions that need further revision in the future, e.g., two dimensionality, constant velocity boundary conditions, conservation of crustal material, mantle delamination, and imposed thermal structure. The assumption of two-dimensional flow allowed us to avoid full three-dimensional computations, but as a result, we cannot obtain along-strike variations of topography and deformation. Similarly, the model uses constant velocity boundary conditions and therefore cannot account for variations in the rate of the Andean crustal shortening, as suggested by *Wdowinski and O'Connell* [1991]. By assuming conservation of crustal material, the model neglects various processes that may have affected the evolution of the central Andes, such as weathering, tectonic erosion, and addition of mantle material to the crust (underplating). These processes affect some areas locally, e.g., tectonic erosion near the trench, weathering in the Cordillera Oriental and Subandes, and underplating in magmatic arc. In fact, some of the along-strike topographic variations may be explained by some of these processes. Another assumption that needs further attention is the thermal structure and thermal processes such as mantle delamination or the effect of magmatism and partial melt in the lithosphere. These issues will be addressed in future work.

Conclusions

The time evolution of the central Andean topography and deformation is investigated by applying a temperature dependent viscoplastic model developed in paper 1 to the South American lithosphere, which deforms in response to tectonic and buoyancy forces. The tectonic forces are induced by subduction of the Nazca plate, which shears the overriding South American plate along the contact between the two plates and horizontally indents the South American lithosphere inland. The buoyancy forces are induced by horizontal density variations due to topographic relief and crustal thickening. The model predicts the observed present-day topographic profile across the central Andes, which consists of dynamically supported near-trench topography and isostatically supported high elevated plateau topography. The 400-km width of the Altiplano is a result of thermal activity that heated and weakened the lithosphere between the present Cordillera Occidental and Cordillera Oriental. The spatial and temporal distributions of volcanic and plutonic activity in the central Andes indicate that such thermal activity has affected the region since late Oligocene until present.

The model also predicts the observed eastward migration of the locus of the Andean crustal deformation and explains the change in the deformation pattern from thick skin to thin skin tectonism. During the early stages of deformation, the crustal and mantle loci of deformation lie in the region between Cordillera Occidental and Cordillera Oriental, where the geotherm is perturbed. The deformation is characterized by a large compressional component and insignificant shear component, leading to thickening of the crust and steady plateau uplift. This stage corresponds to the thick skin deformation of the Altiplano region. At later stages, as the crust thickens, buoyancy forces of larger magnitude resist crustal thickening beyond 65 km. As a result, the locus of the crustal compressional deformation migrates eastward to the Subandes region, whereas the locus of the mantle deformation remains at the weak thermally perturbed region beneath the Altiplano. The detachment of the crustal locus of deformation from that of the mantle produces a significant shear deformation component that is centered in the weak lower crust. This deformation, dominated by a strong shear component, corresponds to the thin skin tectonism that produced the fold and thrust belt of the Subandean region.

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