

# Crustal-thickness variations in the central Andes

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## ABSTRACT

We estimated the crustal thickness along an east-west transect across the Andes at lat 20°S and along a north-south transect along the eastern edge of the Altiplano from data recorded on two arrays of portable broadband seismic stations (BANJO and SEDA). Waveforms of deep regional events in the downgoing Nazca slab and teleseismic earthquakes were processed to isolate the P-to-S converted phases from the Moho in order to compute the crustal thickness. We found crustal-thickness variations of nearly 40 km across the Andes. Maximum crustal thicknesses of 70–74 km under the Western Cordillera and the Eastern Cordillera thin to 32–38 km 200 km east of the Andes in the Chaco Plain. The central Altiplano at 20°S has crustal thicknesses of 60 to 65 km. The crust also appears to thicken from north (16°S, 55–60 km) to south (20°S, 70–74 km) along the Eastern Cordillera. The Subandean zone crust has intermediate thicknesses of 43 to 47 km. Crustal-thickness predictions for the Andes based on Airy-type isostatic behavior show remarkable overall correlation with observed crustal thickness in the regions of high elevation. In contrast, at the boundary between the Eastern Cordillera and the Subandean zone and in the Chaco Plain, the crust is thinner than predicted, suggesting that the crust in these regions is supported in part by the flexural rigidity of a strong lithosphere. With additional constraints, we conclude that the observation of Airy-type isostasy is consistent with thickening associated with compressional shortening of a weak lithosphere squeezed between the stronger lithosphere of the subducting Nazca plate and the cratonic lithosphere of the Brazilian craton.

## INTRODUCTION AND TECTONIC SETTING

The central Andes, in western Bolivia, northern Chile, and northwestern Argentina are among the highest and widest mountain belts in the world (Fig. 1). Elevations in the central Andes reach over 6000 m, and the deformation extends nearly 1000 km inland from the trench. The central Andes offer a modern example of the mountain-building process associated with oceanic-continental convergence and provide a modern analogue for interpreting older convergent mountain belts worldwide. However, the mountain building processes of the central Andes are not well understood, partly owing to a lack of large-scale structural control. The origin of the central Andes, in particular its thick crust (James, 1971), remains enigmatic, as does its relation to the subduction process. Distributed crustal shortening and magmatic addition associated with the subduction process have been proposed as the major contributors to thickening the crust (e.g., Isacks, 1988; Kono et al., 1989).

The central Andes are divided from west to east into five major subparallel morphological and/or structural units (Fig. 1). The Western Cordillera is a high, active volcanic arc with peaks locally reaching 6000 m. The Altiplano is a high, relatively flat plateau with an average elevation of 3800 m. The Altiplano is bordered to the east by the Eastern Cordillera, a high mountain range dominated by folding and thrusting of Paleozoic through Cenozoic rocks. Farther east is the Subandean zone, an active thin-skinned fold-and-thrust belt. The Chaco Plain is a stable platform underlain by the Brazilian craton. The exact age of the deformation and uplift of the Altiplano, Western Cordillera, and Eastern Cordillera is not tightly constrained, but most workers agree that uplift occurred in the past 27 m.y. Estimates of upper-crustal shortening in the Eastern Cordillera and the Subandean zone range from 100 to 225 km (Sheffels et al., 1990; Lyon-Caen et al., 1985; Baby et al., 1990; Schmitz, 1994). In the Altiplano itself, estimates of crustal shortening are more

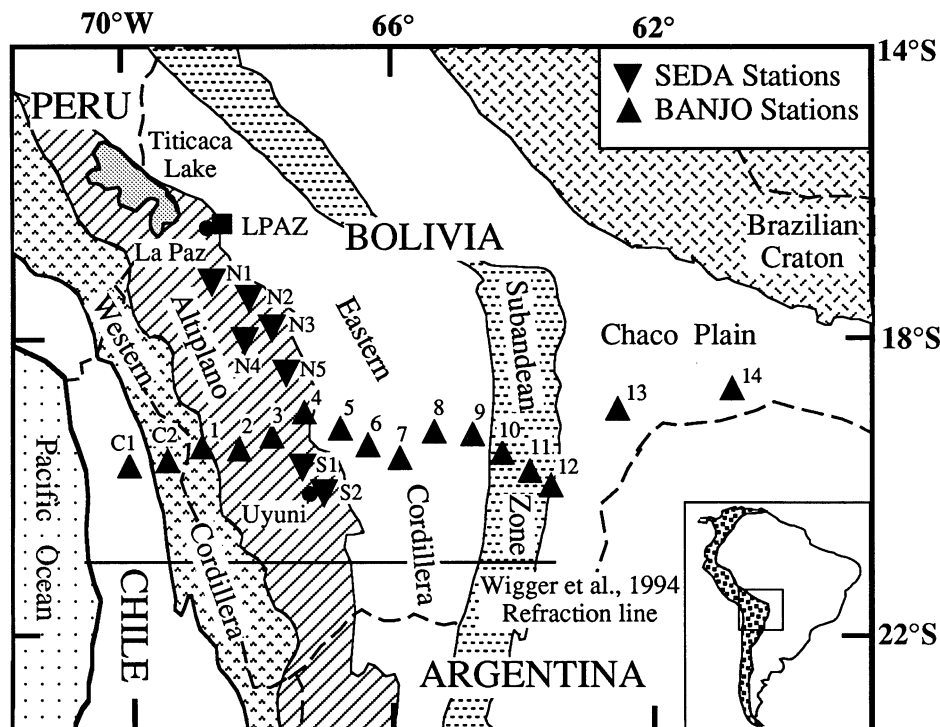


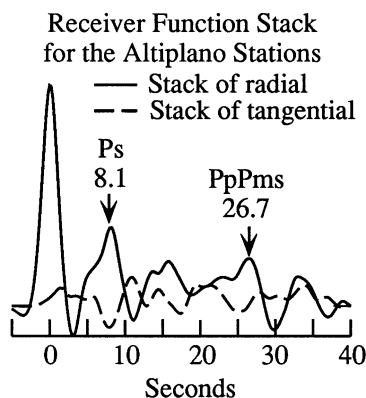
Figure 1. Map of central Andes of Bolivia and northern Chile showing morpho-tectonic units and BANJO and SEDA broadband seismic station locations. LPAZ is permanent seismic station near La Paz, Bolivia.

problematic because much of it is covered by Cenozoic sedimentary and volcanic units. However, where Paleozoic and Mesozoic rocks are exposed, they are highly deformed, indicating significant crustal shortening.

In 1994 we deployed two broadband, three-component seismic arrays with a total of 24 stations in the central Andean Cordillera of Bolivia and northern Chile (Fig. 1). Our seismic experiment consisted of an east-west transect called the BANJO (Broadband ANdean JOint) experiment and a north-south transect called the SEDA (Seismic Exploration of the Deep Altiplano) experiment (Beck et al., 1994). The BANJO experiment consisted of 16 stations along an east-west transect at 19°S to 20°S and extended for nearly 1000 km from near the coast of northern Chile to the Chaco Plain. The SEDA experiment consisted of seven stations that were deployed in a 350 km north-south transect along the eastern boundary of the Altiplano, between La Paz and Uyuni, Bolivia. To address the question of the formation of the central Andes, we have estimated crustal thicknesses along these two transects. Data from teleseismic and deep regional earthquakes recorded on these portable broadband stations were used to constrain crustal thicknesses in the vicinity of each station.

### RECEIVER FUNCTION ANALYSIS

Receiver function analysis is a technique to constrain crustal and, in some cases, lithospheric structure in the vicinity of three-component broadband seismic stations (e.g., Myers and Beck, 1994; Owens et al., 1987). A teleseismic P wave impinging on the crust generates converted phases at major discontinuities below the station. The P-to-S converted phases are isolated by a time-domain deconvolution of the vertical component from the radial component to remove common-source effects (radial receiver function) (Ammon, 1992). For a flat-layered medium, the deconvolution of the vertical component from the tangential component (tangential receiver function) should be zero in the P-wave time window and can be used to check this assumption. One of the largest impedance contrasts within the lithosphere is the crust-mantle boundary, or Moho, which often generates a prominent P-to-S conversion, Ps, that arrives after the direct P wave. The timing between the direct P and Ps arrivals is a function of both the average crustal velocities ( $V_p$  and  $V_s$ ) and crustal thickness. There is a tradeoff between average crustal velocities (both  $V_p$  and  $V_s$ ) and crustal thickness, and therefore additional information is necessary to uniquely determine the crustal thickness

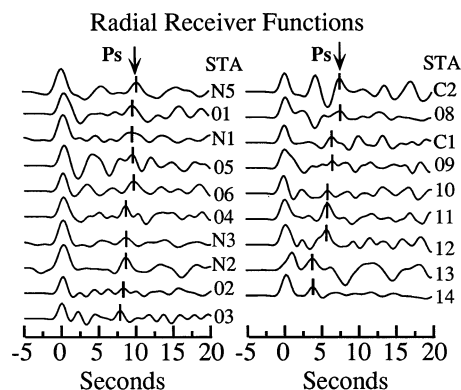


**Figure 2.** Radial (solid line) and tangential (dashed line) receiver function stacks for stations 2, 3, and 4 (Altiplano). A total of eight event-station pairs were used in stacks. Receiver functions are low-pass filtered with a 0.3 Hz corner. Notice the prominent Moho Ps conversion at 8.1 s and the crustal multiple, PpPms, at 26.6 s after the direct P arrival.

(Ammon et al., 1990). In some cases, the crustal multiple PpPms (P wave that reflects from the free surface and then converts to an S wave at the Moho beneath the station) can be identified in the radial receiver function. The time difference between the direct P wave, the Ps conversion at the Moho, and the crustal multiple (PpPms) can be used to determine the  $V_p/V_s$  ratio independently (i.e., Poisson's ratio) (Zandt et al., 1995).

Teleseismic events and deep regional events occurring within the Nazca slab were used as sources. The large, deep June 9, 1994, Bolivia earthquake (depth 640 km) sequence and events in the Mexico and South Sandwich Islands subduction zones provided the best sources. Only events with good signal-to-noise ratios and with reasonably small tangential components in the P-wave time window were utilized in our analysis. The Moho is sampled ~20 to 30 km (depending on the ray parameter) horizontally from the receiver in the direction of the seismic source. By using events from different azimuths we can sample the Moho in several regions around the station (Myers and Beck, 1994). For displaying our results, we group together the events located at northerly and southerly azimuths, respectively, from the stations.

We determined the radial and tangential receiver functions for each station-event pair for which data were available. Receiver functions from events with the same azimuth and similar ray parameters were stacked to enhance the signal-to-noise ratio. To confidently identify crustal multiples, we also summed the station stacks at stations 2,



**Figure 3.** Radial receiver function stacks for individual stations on BANJO and SEDA transects. All receiver functions are low-pass filtered with a 0.3 Hz corner. Short, heavy thick marks indicate the Ps conversion from the Moho.

3, and 4, which are all located on the Altiplano and had similar long-period receiver functions. Other stations showed more individual variation; hence, we could not stack them. Radial and tangential receiver functions low-pass filtered at 0.3 Hz for the Altiplano stack (Fig. 2) show the Ps conversion from the Moho and the PpPms arrival at 8 and 26 s after the direct P-wave arrival. This differential traveltime yields a Poisson's ratio of 0.25 for the Altiplano crust according to the formulation of Zandt et al. (1995). We calculated crustal-thickness values of 63 and 65 km by using a Poisson's ratio of 0.25 and an average crustal P-wave velocity of 6.0 and 6.25 km/s, respectively. These results are consistent with the independent estimates by Zandt et al. (1996), Schuessler (1994), and Wigger et al. (1994) based on independent seismic data and techniques. The study of Zandt et al. (1996) is the most relevant to our study because they used regional seismic data from common stations on the Altiplano. They confirmed that the best average parameters for the central Altiplano crust are a mean crustal  $V_p$  of 6.0 km/s, an average Poisson's ratio of 0.25, and a crustal thickness of 65 km.

We examined each individual station to determine the crustal thickness in the vicinity of that station (Fig. 3). We measured the timing of the Ps conversion from each receiver function stack at each station and then calculated the crustal thickness by using assumed values of average crustal velocity and Poisson's ratio based on previous work and our Altiplano stack. The vertical lengths of the bars in Figure 4 show the range of calculated thicknesses for each station. For stations C1 and C2 in Chile and stations 1 to 10 in Bolivia, we use a range in average crustal P-wave velocities from 6.0 to

6.25 km/s and a Poisson's ratio of 0.25. This range in average crustal velocity results in an uncertainty in crustal thickness of ~3 km for each station.

Poisson's ratio has a large effect on calculations of crustal thickness from receiver functions, and outside the Altiplano this parameter is not tightly constrained. For the Subandean zone (stations 11 and 12) and the Chaco Plain (stations 13 and 14), we estimate crustal-thickness values for a larger range of average crustal  $V_p$  (6.0 to 6.5 km/s) and a range in Poisson's ratio (0.25 to 0.27). A Poisson's ratio of 0.27 is the global crustal average reported by Zandt and Ammon (1995). A higher Poisson's ratio and faster  $V_p$  may be more appropriate for the Chaco Plain, where the surface layer is probably underlain by the Brazilian shield. The larger range in crustal parameters results in an uncertainty of ~6 km in crustal thickness.

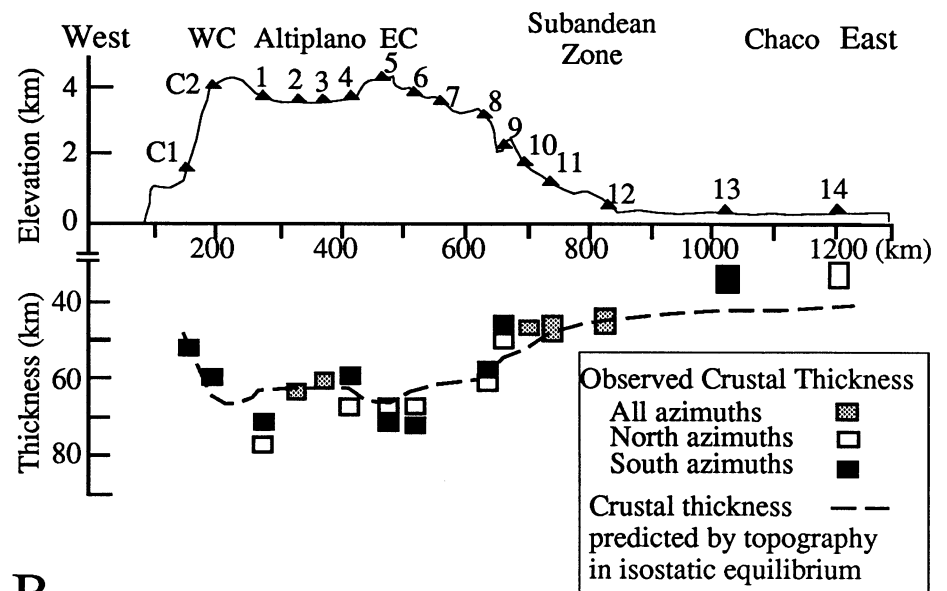
### CRUSTAL-THICKNESS VARIATIONS

We found a nearly 40 km variation in crustal thickness across the Andes. Crustal thicknesses range from 70 to 74 km under the Western Cordillera and Eastern Cordillera and thin to 32–38 km under the Chaco Plain. The central Altiplano at 20°S has crustal thicknesses of 60 to 65 km. The Eastern Cordillera stations show thinner crust to the north near La Paz (56–60 km) and thicker crust in central Bolivia (70–74 km). The Subandean zone crust thins to 42 to 48 km. Although this paper focuses on crustal thickness, unfiltered receiver functions indicate the presence of significant mid-crustal structure beneath many of the seismic stations. The higher-frequency signal in the receiver functions did not appear to correlate between adjacent stations, and the regional extent of the structures generating the signals must await more detailed modeling and comparisons of data from individual stations.

An east-west seismic refraction profile (Wigger et al., 1994) near lat 21.2°S (150 km south of our transect) also found thick crust and low average P-wave velocities ( $V_p = 5.9\text{--}6.1$ ), but sampling of the Moho beneath the Western Cordillera and Altiplano (between long 66.7°W and 68.5°W) was limited by high crustal attenuation of the explosion seismic sources. Our results in general are similar to the results obtained by Wigger et al. (1994) for the Pre-Cordillera, most of the Eastern Cordillera, and the Subandean zone; however, we identify a clear Moho beneath the Western Cordillera and Altiplano.

We calculated the crustal thickness predicted from the topography of the Andes by assuming that the crust is isostatically compensated at a depth of 100 km and using an

## A Crustal Thickness in the Central Andes



## B NNW Eastern Cordillera Altiplano SSE

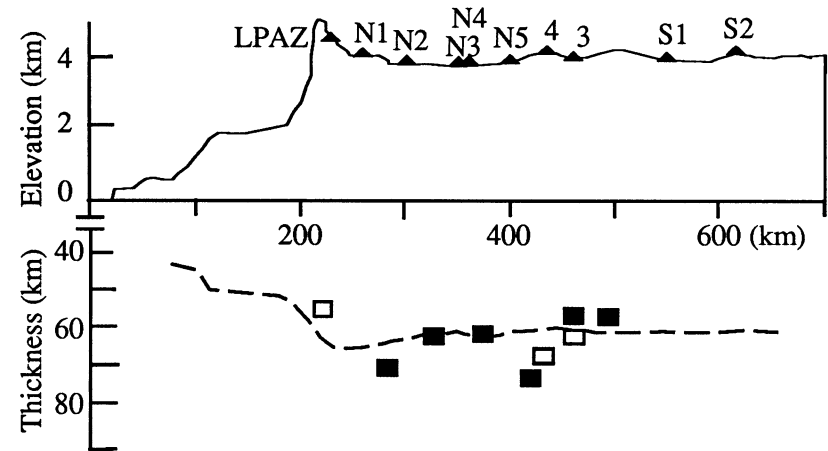


Figure 4. A. East-west cross section showing topography (solid line), observed crustal thickness (shaded, solid, and open symbols), and predicted crustal thickness based on the topography assuming an Airy-type behavior (dashed line). The shaded bars are crustal thicknesses estimated where all events gave similar results. For stations where events sampling the Moho south of the station and events sampling north of the station gave different values, solid and open symbols are shown, respectively. WC = Western Cordillera; EC = Eastern Cordillera. B. North-south cross section with symbols as in A.

average crustal density of 2.7 g/cm<sup>3</sup> and a mantle density of 3.3 g/cm<sup>3</sup> (Fig. 4). Overall, the correlation between our observed crustal thickness and the predicted values is remarkable (Fig. 4). However, we note several discrepancies. At station 1 in the Western Cordillera of Bolivia, our observations indicate crust 6–10 km thicker than predicted. Station 6 in the Eastern Cordillera also shows crust 4–6 km thicker than predicted. The change in crustal thickness between stations 8 and 9 is very abrupt and indicates an average westward dip on the Moho of 7°. The elevations at stations 9 and 10 and at stations 13 and 14 in the Chaco

Plain predict thicker crust than we observe. In these regions the topography may be held up in part by the flexural rigidity of a stronger lithosphere (Lyon-Caen et al., 1985; Fan et al., 1995).

### IMPLICATIONS AND CONCLUSIONS

The close agreement of the crustal thickness with topography suggests that to a first-order approximation, the crust in the Altiplano is in Airy-type isostatic balance, and the high topography is not supported by the flexural rigidity of the lithosphere mantle. The tectonic implications of Airy-type isostatic behavior depend on the vertical dis-

tribution of strength in the lithosphere (Zuber, 1994). If an average weak rheology is assumed for the entire lithosphere, the crustal-thickness changes would reflect variations in the vertical strain rate or in lateral strength (England and Houseman, 1988). A weak lithosphere will accommodate shortening by a pure-shear thickening of the entire lithosphere. A rheology with a weak lower crust sandwiched between a strong upper crust and strong upper mantle can warp under compression into large-scale folding generating long-wavelength Moho topography (Jin et al., 1994). A "jelly sandwich" rheology is likely to accommodate significant shortening by decoupling of the crust and mantle and simple-shear backarc underthrusting (subduction) of the mantle lithosphere (Willett and Beaumont, 1994). Assuming a similar rheological layering, Bird (1991) showed that an overthickened crust is a likely locality for lower-crustal flow that may either flatten or roughen the Moho, depending on the degree of heterogeneity of the adjacent mantle lithosphere. We conclude that the observation of Airy-type isostasy does not, by itself, discriminate between these different models of continental deformation.

There are a number of observations that suggest the lithosphere in the Altiplano and Cordilleras is weak compared to the subducting oceanic Nazca plate to the west and the cratonic lithosphere of the Brazilian shield to the east. The low average velocities and low Poisson's ratio of the crust are consistent with a quartz-rich, felsic bulk composition and constrain any mafic lower-crustal layer to a thickness of <10 km (Zandt et al., 1996; Schuessler, 1994). The heat flow in the Altiplano and Cordilleras is high, ~70 and 110 mW/m<sup>2</sup> (Henry and Pollack, 1988), suggesting a high geothermal gradient. At depths of 50 to 70 km, lower crust of felsic to intermediate composition should be near its solidus and contribute to a weak overall lithosphere (Kuszniir and Park, 1986). With these additional constraints, the observation of Airy-type isostasy is consistent with thickening associated with compressional shortening of a weak lithosphere. No matter the mode of shortening, for significant amounts of shortening, a mechanism must be acting to remove some portion of the shortened upper mantle beneath the Altiplano. Whether this mechanism is a form of delamination (England and Houseman, 1988) or underthrusting and rollback (Willett and Beaumont, 1994) remains unresolved.

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