The nature of orogenic crust in the central Andes

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[1]   The central Andes (16°–22°S) are part of an active continental margin mountain belt and the result of shortening of the weak western edge of South America between the strong lithospheres of the subducting Nazca plate and the underthrusting Brazilian shield. We have combined receiver function and surface wave dispersion results from the BANJO-SEDA project with other geophysical studies to characterize the nature of the continental crust and mantle lithospheric structure. The major results are as follows: (1) The crust supporting the high elevations is thick and has a felsic to intermediate bulk composition. (2) The relatively strong Brazilian lithosphere is underthrusting as far west (65.5°W) as the high elevations of the western part of the Eastern Cordillera (EC) but does not underthrust the entire Altiplano. (3) The subcrustal lithosphere is delaminating piecemeal under the Altiplano-EC boundary but is not completely removed beneath the central Altiplano. The Altiplano crust is characterized by a brittle upper crust decoupled from a very weak lower crust that is dominated by ductile deformation, leading to lower crustal flow and flat topography. In contrast, in the high-relief, inland-sloping regions of the EC and sub-Andean zone, the upper crust is still strongly coupled across the basal thrust of the fold-thrust belt to the underthrusting Brazilian Shield lithosphere. Subcrustal shortening between the Altiplano and Brazilian lithosphere appears to be accommodated by delamination near the Altiplano-EC boundary. Our study suggests that orogenic reworking may be an important part of the "felsification" of continental crust.   *INDEX TERMS:* 7205 Seismology: Continental crust (1242); 7215 Seismology: Earthquake parameters; 7218 Seismology: Lithosphere and upper mantle; 7230 Seismology: Seismicity and seismotectonics; 7260 Seismology: Theory and modeling; *KEYWORDS:* continental crust, lower crust, regional waveforms, central Andes, crustal structure

1.   Introduction and Tectonic Setting

[2]   A major question in Earth science is how continental crust forms and how it is modified and reworked through time. Mountain belts associated with convergent margins have long been thought to be sites of both continental growth by addition of material from the mantle and continental recycling or reworking from delamination [*Kay et al.*, 1994; *Rudnick*, 1995]. However, both of these processes are difficult to document and probably occur at different stages of mountain belt evolution. Seismological studies are beginning to give us information about the composition of the lower crust, a critical constraint in understanding the processes involved in mountain building and crustal recycling. The Nazca-South American plate boundary and associated Andean Cordillera is one of the largest mountain belts in the world with some of the thickest crust on Earth; hence it is an ideal place to study the nature of tectonically thickened crust in orogenic systems.

[3]   The central Andes (between 16°S and 21°S) in northern Chile and Bolivia are the highest and widest part of the Andean mountain belt, with deformation extending well into the back arc region, more than 800 km inland from the trench (Figure 1). To first order, the high elevations of the back arc region correspond to large regions of crust in excess of 60 km thick. The present-day relative plate convergence between the Nazca and the South American plate is 80 mm/yr [*DeMets et al.*, 1990]. Approximately 10–15 mm/yr of that convergence is taken up in a broad zone of deformation comprising the Andean Mountain belt [*Norabuena et al.*, 1998]. Near-orthogonal convergence has occurred at a rate of 50–150 mm/yr since about 50 Ma [*Somoza*, 1998; *Pardo-Casas and Molnar*, 1987].

[4]   The Andean Cordillera is often considered a modern analog to the older western North America Cordillera. The American cordilleras have become the type examples of continental margin mountain belts. Most workers agree that a large part of the crustal thickening in the back arc of the central Andes is due to tectonic shortening (for summaries, see *Isacks* [1988], *Lamb et al.* [1997], and *Allmendinger et al.* [1997]) rather than magmatic addition from the mantle. Magmatic addition may be significant locally in the volcanic arc regions [*Kono et al.*, 1989]. The geologic record of the western margin of South America indicates a change from a neutral to a compressional back arc setting during the
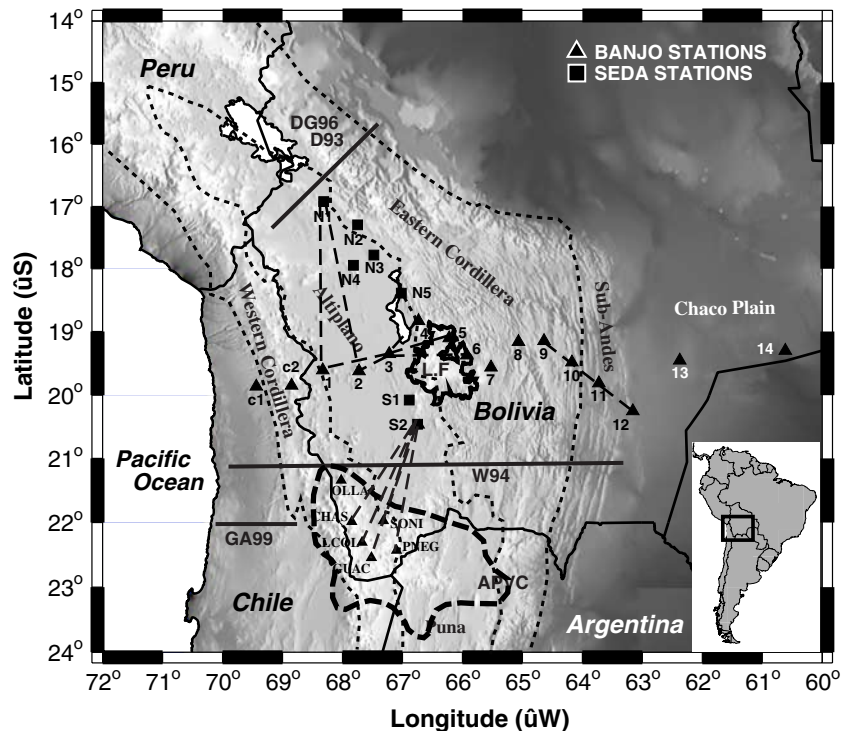
Figure 1. Map showing location of the study area and the BANJO (large triangles), SEDA (squares), and Atiplano-Puna Volcanic Complex (APVC) (small triangles) portable broadband seismic stations. Station elevations range from 100 m (station 14) to 4200 m (station 5). The morphotectonic provinces referred to in the text are shown with short dashed lines. Paths used in surface wave studies by *Baumont et al.* [2002] are shown in long dashed lines. The Los Frailes (LF) volcanic complex and the magma body in the APVC from *Chmielowski et al.* [1999] are also shown. Thin black lines show locations of other geophysical studies referred to in the text. D93, *Dorbath et al.* [1993]; DG96, *Dorbath and Granet* [1996]; W94, *Wigger et al.* [1994]; GA99, *Graber and Asch* [1999].

Cretaceous [*Coney and Evenchick*, 1994]. This is preserved in the geologic record in the back arc as a transition from platform deposits to westerly derived foreland deposits [*Coney and Evenchick*, 1994; *Sempere*, 1990; *Horton and DeCelles*, 1998]. *Coney and Evenchick* [1994] proposed that the onset of compression in the back arc of the Andes was triggered by the opening of the South Atlantic at 110–130 Ma, which initiated the westward absolute motion of the South American plate.

[5] There is still considerable debate on the details of the timing of deformation and the overall amount of crustal shortening in the central Andes. It is generally thought that increased compression in the back arc began in the Late Cretaceous, but no significant areas rose above sea level until approximately 60 Ma [*Sempere et al.*, 1997]. Many workers argue that a large part of the shortening occurred in the last 26 m.y. [*Allmendinger et al.*, 1997; *Sempere*, 1990; *Isacks*, 1988; *Sheffels*, 1990], with as much as 2 km of uplift in the last 10 m.y. [*Gregory-Wodzicki et al.*, 1998; *Gubbles et al.*, 1993; *Isacks*, 1988]. However, other studies suggest that significant tectonic shortening occurred prior to the Neogene [*Horton and DeCelles*, 1998; *Lamb et al.*, 1997; *Horton et al.*, 2001; *DeCelles and Horton*, 2002]. The minimum upper crustal shortening amounts during the Neogene determined from geologic studies range from 240 to 320 km in the vicinity of 20°S and can account for

at least 70–80% of the present crustal volume (see *Allmendinger et al.* [1997] and *Kley and Monaldi* [1998] for summaries). It is less clear how the lower crust and mantle lithosphere respond to the shortening. Even these minimum amounts of shortening, if applied uniformly to the entire lithosphere, quickly cause a space problem within the mantle wedge above the subducting slab. Processes such as lithospheric subduction, delamination and/or convective instability have all been proposed as ways to deal with the space problem. In the central Andes we can evaluate the role of the entire lithosphere in a region of active mountain building. Recent geophysical data collected in the back arc region of the central Andes have given us an unprecedented image of the lithospheric structure. This present-day view of the lithosphere provides strong constraints on the evolution of the central Andes.

[6] The central Andes in the vicinity of 20°S are composed of several north-south trending morphotectonic provinces (Figure 1). The Western Cordillera is an active magmatic arc made up of andesitic to dacitic stratovolcanoes with peaks exceeding 6000 m and associated ignimbrite deposits. However, locally a noticeable decrease in volcanic arc activity occurs between 19.2°S and 20.7°S over a north-south distance of 150 km [*Worner et al.*, 1994]. The Altiplano is a high, relatively flat, internally drained plateau with an average elevation of 4000 m. The Altiplano surface

Table 1. Earthquakes Used in the Receiver Function Analysis

Year	Julian Day	Time, UT	Latitude, °N	Longitude, °E	Depth, km
1994	119	0711:30.40	−28.302	−63.172	572
1994	143	0141:46.30	18.306	−100.522	86
1994	160	0033:16.50	−13.834	−67.563	636
1994	160	0526:30.70	−13.846	−67.336	631
1994	185	2136:44.80	14.939	−97.325	33
1994	206	0439:17.50	−56.963	−25.522	33
1994	220	0755:39.60	−13.819	−68.404	602
1994	230	0113:06.10	35.562	−0.107	8
1994	244	1515:53.20	40.406	−125.648	10
1994	344	1617:41.10	18.243	−101.354	67
1995	3	1611:58.99	−5.769	−6.822	33
1995	50	0403:15.70	40.513	−125.620	10
1995	67	0346:02.30	16.598	−59.510	15
1995	85	0216:13.30	−55.858	−28.163	76
1995	104	1315:17.00	−60.661	−19.951	10
1995	132	1512:23.10	−19.339	−63.891	601
1995	138	0006:26.80	−0.950	−21.985	10

is covered by several large salars, Quaternary basin fill, and Late Oligocene to Recent volcanic rocks [*Lamb et al.*, 1997; *Allmendinger et al.*, 1997; *Davidson and de Silva*, 1992]. Toward the south the Altiplano grades into the higher Puna plateau, which has an average elevation of 5000 m. This transition is the site of the Altiplano-Puna Volcanic Complex (APVC), a late Miocene to Recent ignimbrite "flare-up" that covers $50,000^2$ km and is one of the largest silicic volcanic fields in the world [*de Silva*, 1989; *de Silva et al.*, 1994].

[7] The Eastern Cordillera is an inactive fold and thrust belt dominated by Ordovician, Silurian and Devonian age rocks with both east and west verging structures [*Lamb et al.*, 1997; *Lamb and Hoke*, 1997; *McQuarrie and DeCelles*, 2001]. The Los Frailes ignimbrite complex covers 8500 km^2 on the western edge of the Eastern Cordillera between 19°S and 20°S [*de Silva and Francis*, 1991]. The Los Frailes ignimbrites have an average elevation of 4500 m and are the easternmost silicic volcanic system in the central Andes. The complex has numerous volcanics with ages ranging from 20 Ma to Recent [*de Silva and Francis*, 1991]. The sub-Andean zone is the active frontal part of the eastward-verging fold and thrust belt. Further east is the Chaco, the modern foreland basin presumably underlain by the Brazilian craton with elevations of only a few hundred meters [*Lamb et al.*, 1997; *DeCelles and Horton*, 2002].

[8] In this paper we have used teleseismic and deep regional earthquakes recorded by the BANJO and SEDA experiments [*Beck et al.*, 1994] for receiver function analysis to investigate the crustal structure in the back arc region of the central Andes between 17°S and 21°S (Figure 1). Short reports on a preliminary receiver function study and a broadband waveform study have been already published [*Beck et al.*, 1996; *Zandt et al.*, 1996]. We also use data from the Altiplano Puna Volcanic Complex (APVC) experiment (Figure 1) in the southern Altiplano [*Chmielowski et al.*, 1999]. A larger-scale study incorporating some of these data to generate a low-pass-filtered, migrated, and stacked receiver function image is recently published [*Yuan et al.*, 2001]. In this study we emphasize the individual receiver function analysis using seismic velocities constrained with published full-waveform modeling [*Swenson et al.*, 1999, 2000] and surface wave dispersion studies [*Baumont et al.*,

2002]. The new seismological results combined with other recent geophysical and geologic studies provide higher-resolution structure models and better constraints on the lithospheric evolution of the central Andes.

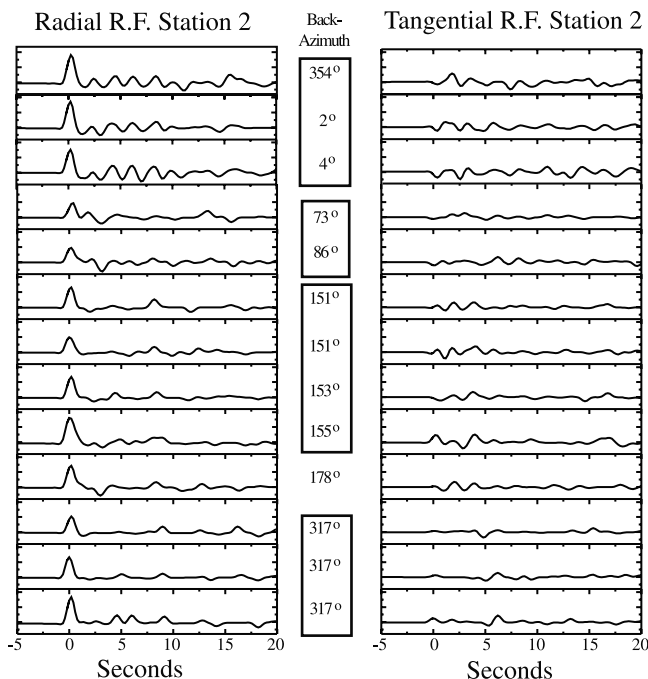## 2. Receiver Function Analysis: Data and Methods

[9] We use 12 teleseismic earthquakes and 5 deep regional earthquakes recorded on the BANJO and SEDA portable seismic networks to investigate the crustal structure in the central Andes (Table 1). The broadband portable arrays were operating during parts of 1994 and 1995 and recorded continuously with either a STS2, Guralp 40T or Guralp ESP sensor and Reftek recorder [*Beck et al.*, 1994]. The Central America, Mexico, and Scotia Arc subduction zones and the Mid-Atlantic Ridge were the best source regions for teleseismic earthquakes (Table 1). We also used several regional events in the Nazca slab, including the large, deep (640 km), M_w = 8.3, 9 June 1994 Bolivia earthquake and two of the largest aftershocks. All the regional events we used had depths greater than 570 km and occurred within 10° of the stations so that the incoming P waves arrived at a steep angle. We obtained useable receiver functions at all available sites except station 7, which was vandalized shortly after deployment, and the two stations in Chile, C1 and C2, which were not deployed long enough to record enough data for a reliable stack.

[10] Receiver function analysis is now a widely used, straightforward method of extracting information about the crust and upper mantle from three-component broadband waveforms [*Owens et al.*, 1987]. We use receiver function analysis to isolate P-to-S (Ps) conversions from discontinuities in the vicinity of each station. The Ps conversions and multiples can be isolated by deconvolving the vertical component from the radial and tangential components. We computed all of our receiver functions for teleseismic and local events using a new time domain, iterative deconvolution algorithm [*Ligorria and Ammon*, 1999]. Long-period stability is insured a priori by constructing the deconvolution as a sum of Gaussian pulses. This method has the advantage of finding the "simplest" receiver function with high-frequency content but without sidelobe artifacts or long-period instability often associated with frequency

domain and other time domain techniques. For flat layers the radial receiver function contains *Ps* conversions from discontinuities beneath the station, while the tangential receiver function should be zero (in the *P* wave time window) with no converted phases. In practice, the tangential receiver functions often contain energy because of noise, dipping structure or anisotropy. The timing and amplitude of a *Ps* conversion constrain the depth and impedance contrast across the discontinuity. However, there is a trade-off between the depth and the average seismic velocity to the discontinuity; hence additional information is needed to uniquely determine either parameter [*Ammon et al.*, 1990]. We have reliable information about the average seismic velocities in the crust from previous studies and can thus use receiver function analysis to identify discontinuities in the crust and determine their depths.

[11]   For the average seismic velocities needed to constrain the depth to the discontinuities, we have used results from recent studies by *Wigger et al.* [1994], *Zandt et al.* [1996], *Swenson et al.* [1999, 2000], and *Baumont et al.* [2002]. These studies, using seismic refraction data, regional waveform data, and surface waves, all found that average crustal velocities in the central Andes are very low compared to global averages. *Zandt et al.* [1996] and *Swenson et al.* [1999, 2000] used waveforms of intermediate-depth (100–260 km), regional earthquakes (m_b = 5.3–5.9) recorded by the BANJO and SEDA portable seismic networks to constrain the average crustal velocity and crustal Poisson's ratio. *Swenson et al.* [2000] used reflectivity synthetic seismograms and a grid search to constrain four parameters of the Altiplano litho-
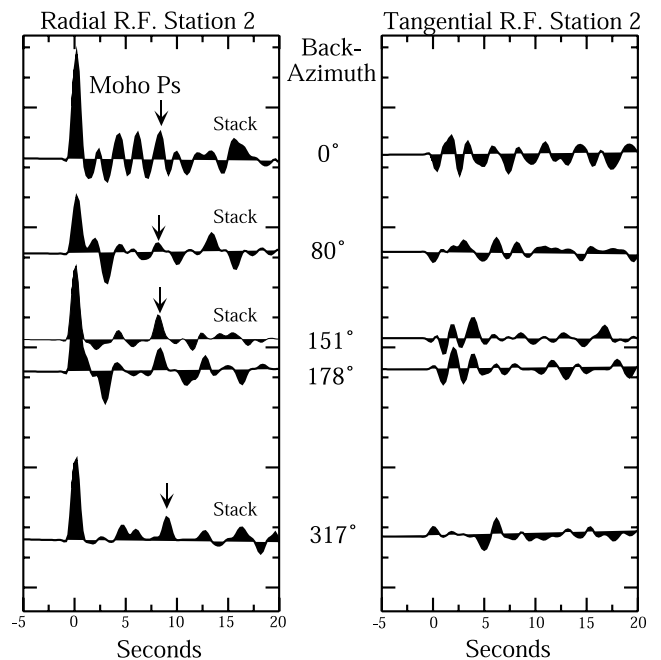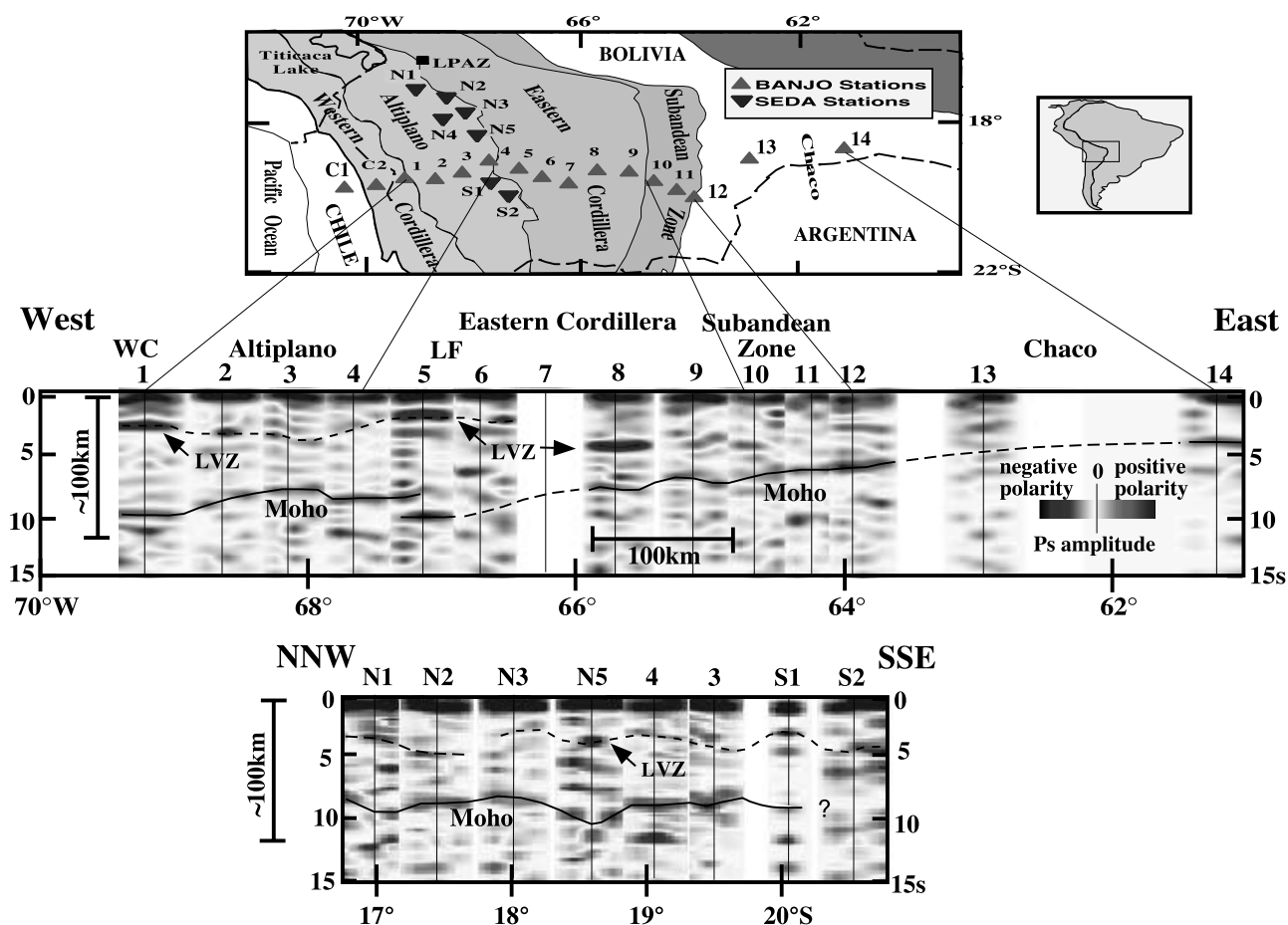

Figure 3. (left) Radial and (right) tangential receiver function stacks for station 2, located on the Altiplano, plotted as a function of backazimuth.

sphere: average crustal velocity, crustal thickness, and crust and upper mantle Poisson's ratios. *Swenson et al.* [1999, 2000] found that the models that provide the best overall fit to the data are characterized by a low average crustal *P* wave velocity (5.75–6.0 km/s), a crustal thickness of 60–65 km, a crustal Poisson's ratio of 0.25, and a mantle Poisson's ratio of 0.27–0.29. In addition, *Swenson et al.* [2000] found similar results for the high-elevation regions of the Eastern Cordillera (BANJO stations 5 and 6). The refraction study of *Wigger et al.* [1994] also found low average crustal velocities of 6.0 km/s for the lower elevation regions of the Eastern Cordillera and the sub-Andean zone in the vicinity of 21°S.

[12]   *Baumont et al.* [2002] determined Love and Rayleigh wave phase-velocity dispersion curves for the fundamental modes between approximately 15 s and 80 s using a two-station method (BANJO and SEDA data) for paths crossing individual morphotectonic regions. *Baumont et al.* [2002] then inverted the dispersion curves for suites of *S* wave velocity models that fit the observed dispersion across each morphotectonic region. They used a stochastic inversion approach from *Shapiro* [1996] to determine a family of V_S models that fit the data within the error bounds. From these solutions an average model and its standard errors were estimated. These studies using different phases provide very important constraints on average crustal velocity for modeling the receiver functions. We used an average *P* wave crustal velocity of 5.75–6.0 km/s and a crustal Poisson's ratio 0.25 to determine the depth to the Moho and other discontinuities in the crust from the receiver functions.

[13]   We determine receiver functions for individual event-station paths and then stacked receiver functions with similar backazimuths and ray parameters. Figure 2 shows an example of individual radial and tangential receiver func-

Figure 2.   Individual radial and tangential receiver functions for events recorded at station 2 on the Altiplano. The receiver functions are plotted as a function of backazimuth and have been band-passed between 1 and 50 s. Receiver functions with similar backazimuths and ray parameters (as outlined by the boxes) are used to construct the stacks shown in Figure 3.

Figure 4. Psuedo cross sections of receiver function traces along (top) W-E and (bottom) NNW-SSE profiles. For each station, three receiver function stacks corresponding to different backazimuths were projected onto the respective record sections. Positive polarity Ps conversions plot as red, and negative polarity Ps conversions plot as blue. Note the Moho Ps observed beneath most stations except where rapid changes occur in crustal thickness. Several low-velocity zones (LVZ) occur in the midcrust beneath the Altiplano and Eastern Cordillera. See color version of this figure at back of this issue.

tions, and Figure 3 shows the resultant receiver function stacks for station 2 located on the Altiplano. For most backazimuths we had from two to four events to include in the stack, and for most stations we have three to four stacks. We usually had stacks from teleseismic events with a backazimuth of 150°–165° (corresponding to source regions in the South Sandwich Islands), 310°–320° (corresponding to source regions in the Mexico and Central America subduction zones), and 70°–90°(corresponding to source regions on the Mid-Atlantic Ridge). The deep events associated with the 9 June 1994 Bolivia earthquake (depth of 645 km) had variable backazimuths depending on the station. In addition, we had numerous single event-station paths.
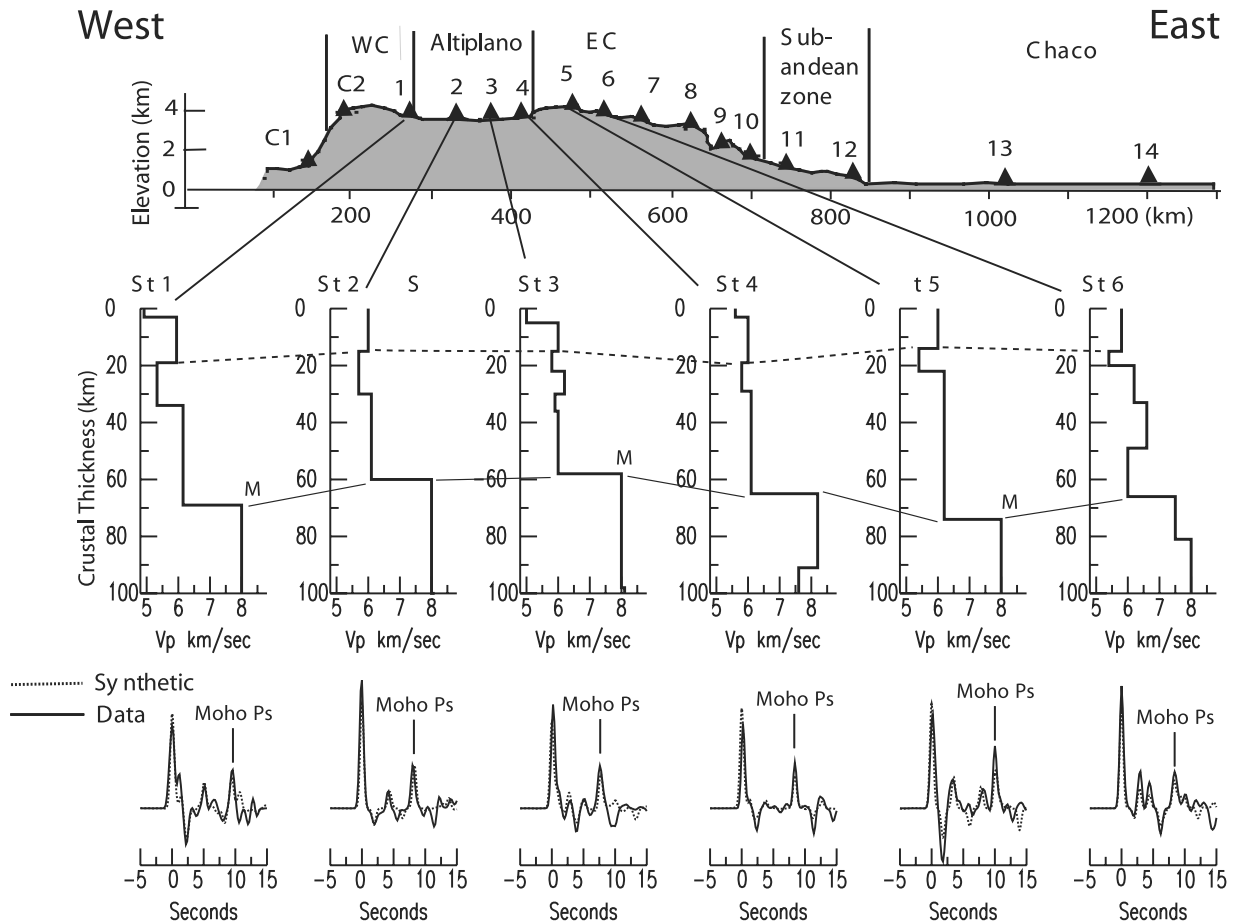
3. Results

3.1. Introduction

[14] Shown in Figure 4 are the radial receiver function stacks for each station plotted with time increasing downward as east-west and north-south psuedo cross sections. For each station three receiver functions with different backazimuths are projected into the line of the cross section.

In most cases the receiver functions are stacks, but in a few cases we include an individual event-station receiver function if a stack was not available. The red indicates a positive-polarity arrival, and the blue indicates a negative-polarity arrival. A positive polarity corresponds to a downward increase in velocity and a negative polarity corresponds to a decrease in velocity across the discontinuity. The receiver functions contain not only Ps conversions but also the multiples. The largest discontinuity within the lithosphere is the Moho, and it often produces the largest Ps conversion on the radial receiver function. For most receiver functions in our study the Ps conversion from the Moho was identifiable (Figure 4). The time delay of the Ps conversion (Ps-P) from the Moho relative to the direct P wave varies from 10 s to 4 s, corresponding to crustal thicknesses between approximately 80 and 32 km. The data show the expected decrease in crustal thickness with decreasing elevation toward the east.

[15] Another common feature observed at many stations is evidence for a midcrustal discontinuity. Because our only constraint is the average crustal velocities, there is still a trade-off between the depths of the midcrustal discontinuities and the absolute velocity within any particular layer. If

Figure 5a. Example receiver functions for BANJO stations 1 through 6 and resultant *P* wave velocity models for the crust. The solid lines are the observed and the dashed lines are the synthetic receiver functions. Note the pronounced low-velocity zone beneath stations 1 (Western Cordillera), 5 and 6 (Eastern Cordillera), and the weaker low-velocity zone across the entire Altiplano.

there is a strong discontinuity in the mid-crust, then the multiples can interfere with the Moho *Ps* conversion, and the receiver function is much more complicated.

[16] The *Ps-P* time is a function of crustal thickness and average crustal *S* and *P* wave velocities. We used forward modeling to determine the simplest velocity model (with the least number of layers) that fit the first 15–20 s of the receiver function using the surface wave and regional-waveform modeling constraints on average crustal velocities (Figures 5a and 5b). We discuss each region from west to east across the central Andes at 20°S.

3.2. Western Cordillera-Altiplano Boundary (Station 1)

[17] We have analyzed receiver function stacks at station 1 located at the Western Cordillera-Altiplano boundary and observe a *Ps* conversion at 9.8 s (Figure 5a). This corresponds to a crustal thickness of 70 km, assuming an average crustal velocity of 6.0 km/s and a V_p/V_s ratio of 1.73. We might expect higher average crustal velocities for the Western Cordillera, but *Swenson et al.* [2000] modeled regional waveforms crossing the eastern edge of the Western Cordillera and recorded at station 1 and found an average crustal *P* wave velocity of 6.0 km/s. In our receiver function stacks at station 1 we see no evidence for a higher-velocity lower

crust. This may be a slightly mixed path (Western Cordillera-Altiplano transition) and is not completely representative of the main volcanic arc. In a local tomography study, *Graber and Asch* [1999] found a two-layer crust with an upper crustal *P* wave velocity of approximately 6.0 km/s (45-km-thick layer) and a lower crustal velocity of approximately 7 km/s (20- to 25-km-thick layer). This tomography study sampled the western edge of the volcanic arc further south at 22°–23°S. We might expect a more mafic high-velocity lower crust under the active arc due to magmatic underplating. Surface wave dispersion paths across the Western Cordillera constrained with a crustal thicknesses of 70 km indicate a significantly higher midcrustal V_s (3.6 km/s) than in the central Altiplano (3.3 km/s) [*Baumont et al.*, 2002]. In addition, *Pn* studies crossing the western Cordillera suggest a smaller crustal thicknesses of about 60 km [*Baumont et al.*, 2001]. Further work needs to be done to determine if the crustal root beneath station 1 is localized or if it is representative of the volcanic arc in this region.

[18] Within the crust we find a prominent negative-polarity arrival at approximately 2.5 s that corresponds to the top of a low-velocity zone at a depth of approximately 20 km. This low-velocity zone is observable on all three receiver function stacks at station 1 (Figures 4 and 5a). This
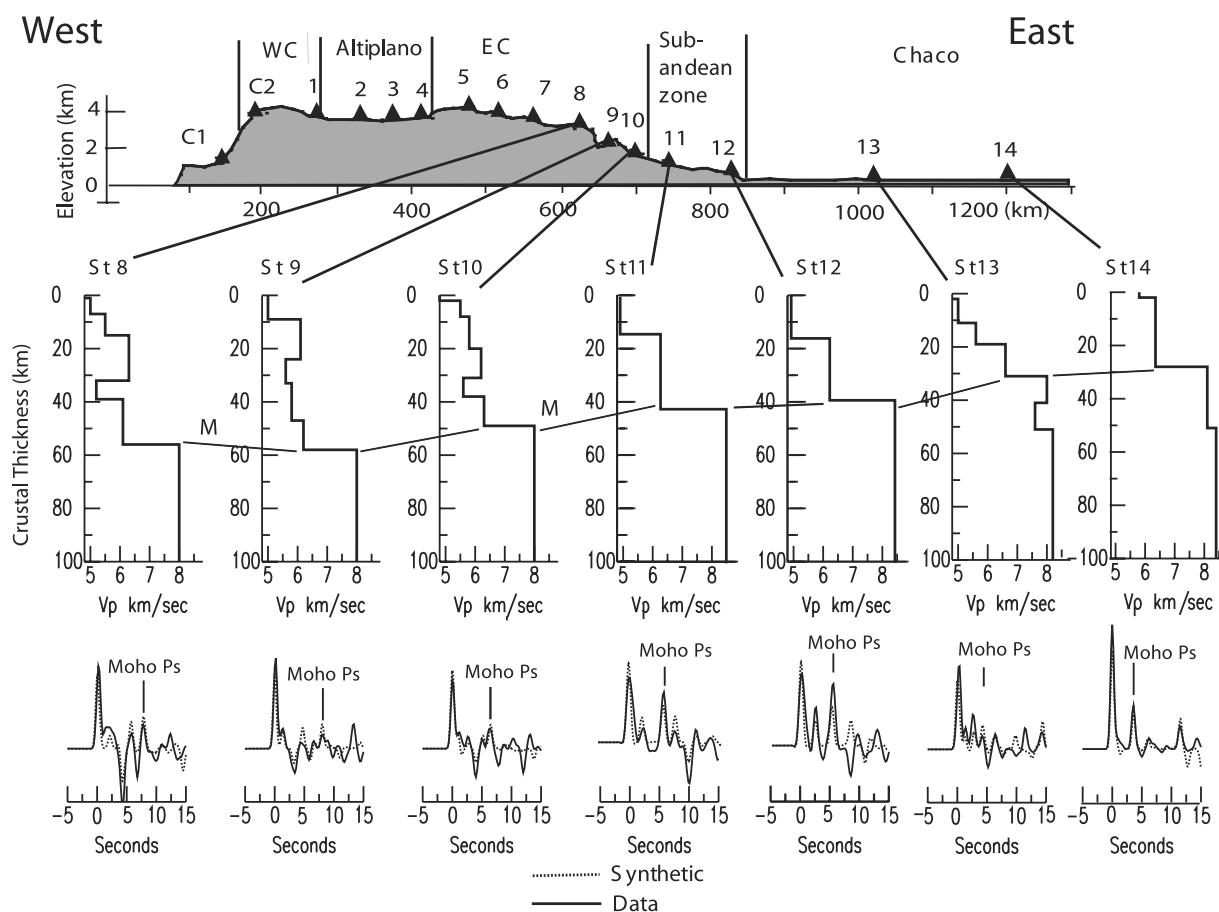
Figure 5b. *P* wave velocity models for the crust determined from receiver functions for stations in the lower elevations of the Eastern Cordillera, sub-Andean zone, and Chaco.

midcrustal low-velocity zone beneath the eastern edge of the Western Cordillera may be a region of partial melt related to the active volcanic arc. This is also the depth at which *Chmielowski et al.* [1999] found a thin, sill-like, regional low-velocity zone beneath the Altiplano/Puna volcanic complex at 22°S.

3.3. Central Altiplano (Stations 2–4)

[19] We obtain our best results from receiver functions determined at stations on the Altiplano. We also have the best independent constraints based on waveform modeling and surface wave dispersion studies on average crustal velocity and Poisson's ratio for the central Altiplano [*Zandt et al.*, 1996; *Baumont et al.*, 2002; *Swenson et al.*, 2000]. Our east-west profile shows crustal thicknesses of 59–64 km for the central part of the Altiplano, assuming an average crustal V_p of 5.8–6.0 km/s and a Poisson's ratio of 0.25 (Figure 5a). We do not find any evidence for a higher-velocity lower crust as is typical in many continental regions [*Christensen and Mooney*, 1995; *Rudnick and Fountain*, 1995]. This result is consistent with both the full-waveform modeling study [*Swenson et al.*, 2000] and the surface wave results (Figure 6), although a thin, <10 km thick, mafic layer at the base of the crust cannot be ruled out on the basis of the available data.

[20] We observe a small negative-polarity arrival in many of the receiver functions at 2.5–3.0 s. This corresponds to the top of a midcrustal low-velocity zone that can be traced across the entire width of the Altiplano at a depth of 14–20 km (Figures 4 and 5a). We do not have very reliable constraints on the thickness of the low-velocity zone because it depends on the velocities of each layer. This weak midcrustal low-velocity zone may mark the decoupling between the brittle upper crust and more ductile lower crust.

3.4. Eastern Cordillera (Los Frailes Area, Stations 5 and 6)

[21] The receiver functions across the entire Eastern Cordillera show large variations between stations. To first order, the crustal thickness decreases with the eastward decline in surface elevation. The *Ps* conversions from the Moho occur more than 8–10 s after the direct *P* wave arrival at stations 5 and 6 in the Los Frailes area but only 6 s later at station 10 near the boundary with the sub-Andean zone. At 21°S *Wigger et al.* [1994] carried out a refraction survey and found average *P* wave crustal velocities of 6.0–6.1 km/s for most of the Eastern Cordillera. Similarly, *Swenson et al.* [2000] modeled regional waveforms with event-station paths across the high-elevation regions of the Eastern Cordillera and found that the average crustal velocities are also relatively slow (5.75–6.0 km/s).

[22] Stations 5 and 6 are situated in the highest elevations of the Eastern Cordillera, within and adjacent to the Los
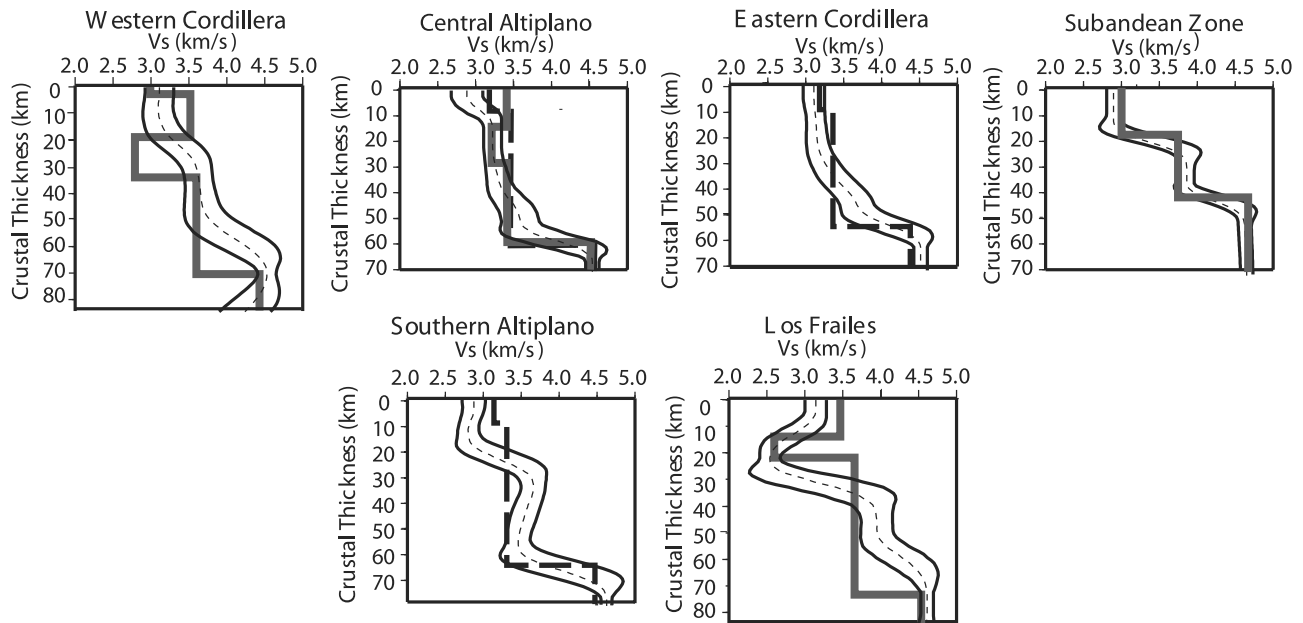
Figure 6. Comparison of crustal *S* wave velocity models from receiver functions studies (solid lines, this study), from surface wave dispersion studies (thin black lines with dashed line in the center, *Baumont et al.* [2002]), and regional waveform modeling studies (thick dashed lines, *Swenson et al.* [2000]) for the central Andes.

Frailes volcanic field. Although the receiver functions for station 5 are complicated, they can be modeled with a very simple structure that has a low-velocity layer at 15–20 km and a crustal thickness of 74 km (Figure 5a). All three stacks (corresponding to different backazimuths) consistently show a large, negative trough at 2 s, which corresponds to the top of a low-velocity zone at a depth of approximately 14 km. The surface wave dispersion models also show this strong midcrustal low-velocity zone although at a slightly greater depth of 25 km (Figure 6). The low-velocity zone beneath station 5 in the Eastern Cordillera corresponds to the location of the large Los Frailes ignimbrite field. This region also corresponds to an area of high attenuation for *Lg* propagation [*Baumont et al.*, 1999]. The surface wave study suggests the existence of a high-velocity lower crust ($V_s = 3.9$ km/s) localized beneath the Los Frailes volcanic field. We observe no indication of such a layer in the receiver functions and we are uncertain about its robustness in the surface wave models.

[23] The receiver functions determined for station 6 vary as a function of the event backazimuth. Receiver functions from events with southeast azimuths show a converted arrival from the top of a small low-velocity zone between 16 and 20 km, similar to receiver functions at station 5 (Figure 5a). They also indicate an apparent crustal thickness of 65 km. Receiver functions from northwest azimuths have a similar low-velocity zone at 15–20 km but also have a prominent negative arrival at 6.5–7.0 s, corresponding to a low-velocity zone in the lower crust at approximately 50 km depth. These receiver functions have a *Ps* conversion corresponding to an apparent crustal thickness of 75 km. The events with different backazimuths are sampling the Moho in different places; hence this could represent topography on the Moho in the vicinity of station 6. We

interpret this as a localized crustal root associated with the Los Frailes volcanic region. The overthickened crust in the western portion of the eastern Cordillera is corroborated by the *Pn* travel time study [*Baumont et al.*, 2001].

3.5. Eastern Cordillera (Eastern Half, Stations 8–10)

[24] Receiver functions from stations 8–10 indicate a crustal thickness of 56, 59, and 50 km, respectively (Figure 5b). Station 8 has a strong low-velocity zone at a midcrustal depth of approximately 30 km. This is substantially deeper than the low-velocity zones associated with stations 5 and 6 and the Los Frailes ignimbrite field. Stations 9 and 10 show variability with event backazimuth and have the least reliable receiver functions along the east-west line (Figures 4 and 5b). Both stations 9 and 10 have indications of crustal low-velocity zones, but the depths are not consistent, although stations 8 and 10 have one at 30 km depth. This is also a region where elevations are dropping rapidly, and we would expect an isostatic crust to be thinning rapidly, creating a west dipping Moho. This expectation is consistent with the large-amplitude tangential receiver functions associated with these stations. The surface dispersion for paths crossing the eastern Cordillera indicate an approximately 60-km-thick crust with lower crustal velocities similar to those observed in the central Altiplano ($V_s = 3.8$ km/s) [*Baumont et al.*, 2002].

3.6. Sub-Andean Zone (Stations 11 and 12)

[25] The receiver functions for stations 11 and 12 show very large amplitude *Ps* conversions from the Moho (Figures 4 and 5b). The *Ps* conversions occur at approximately 6 s after the direct *P* wave. We have less information about average crustal velocities or crustal Poisson's ratio for the sub-Andean region, but *Wigger et al.* [1994] found low

average crustal velocities of 5.9 km/s in the sub-Andean zone at 21.15°S. This low average velocity is in part the result of low-velocity Mesozoic and Paleozoic sedimentary rocks in the upper 10 km of the fold and thrust belt. *Baumont et al.* [2002] modeled surface wave dispersion curves and found V_s models that contained a slow upper crust and a faster lower crust. Using an average crustal velocity of 5.9 km/s to calculate crustal thickness gives values of 42 and 40 km for stations 11 and 12, respectively. The other consistent feature of the receiver functions for stations 11 and 12 is a converted phase at 2.5–3.0 s, with a positive polarity that corresponds to a midcrustal discontinuity with an increase in velocity.

[26]   However, we cannot match the large amplitude of the *Ps* conversion from the Moho using a flat-layered crustal structure. We have calculated synthetic receiver functions for a dipping Moho (15° dip to the west) and find that although it does help to increase the amplitude of the *Ps* conversion from the Moho, it still does not completely explain the azimuthal pattern. Our preferred model for stations 11 and 12 are shown in Figure 5b for the sub-Andean zone. Both the receiver functions and the surface wave dispersion can be fit with a simple two-layer crust (Figure 6).

3.7. Chaco (Stations 13 and 14)

[27]   Stations 13 and 14 are located in the Chaco and were operating for less than a year; hence we have much less data than at the other stations. Receiver functions determined using data from station 14 are very simple, with a *Ps* conversion from the Moho at 4 s. If we assume an average crustal V_p of 6.0 km/s and a Poisson's ratio of 0.25, the result is a crustal thickness of 29 km. In contrast, receiver functions determined for station 13 are complicated, perhaps in part due to near-surface structure. They do not contain a strong *Ps* conversion from the Moho. In Figure 5b we show the fit of a synthetic receiver function to an observed receiver function for a crustal thickness of 30 km with some midcrustal structure. Although we cannot fully explain the receiver functions at station 13, they are consistent with station 14 data in that they suggest very thin crust. This is surprisingly thin crust for a region often characterized as having Brazilian craton at depth. We have few constraints on the average crustal velocity or Poisson's ratio for the Chaco region where we have sites. At 21°S, *Wigger et al.* [1994] found average crustal velocities of 5.9 km/s for the sub-Andean-Chaco boundary. If we assume that the crust is faster than the 6.0 km/s, this would result in thicker crust. *Snoke and James* [1997] analyzed surface wave group velocities and also found thin crust (32 km) in the Chaco further east in southern Brazil.

3.8. North-South Profile (SEDA Stations)

[28]   In a north-south section along the Altiplano-Eastern Cordillera border, the crustal thickness varies from 60–70 km in the northern and central Altiplano to as thin as ~50 km south of 22°S (Figure 4) [*Zandt et al.*, 2002; *Yuan et al.*, 2001]. The azimuthally variable Moho *Ps* at station N5 may be sampling the thickening of the crust in the Eastern Cordillera. In conjunction with this general southward crustal thinning, the low bulk crustal seismic velocity of the central Altiplano also decreases [*Swenson et al.*, 2000; *Baumont et al.*, 2002]. An undulatory midcrustal low-velocity layer appears to extend the entire length of the Altiplano (Figure 4). This observation strengthens the conclusion that a regionally prevalent crustal low-velocity layer (LVL) characterizes the whole Altiplano, the western flank of the Eastern Cordillera, and the Puna. The top of the LVL appears to link with the basal detachment of the fold-thrust belt under the Eastern Cordillera [*Yuan et al.*, 2001]. This depth may represent the brittle/ductile transition and be the decoupling depth between upper crustal imbrication and lower crustal ductile flow. In the northern Puna, the Altiplano-Puna magma body is apparently localized at this boundary [*Chmielowski et al.*, 1999].

4.   Discussion
4.1.   Introduction

[29]   The large number of geophysical studies in the central Andes during the last decade have provided a vastly improved image of the present lithospheric structure that gives us a snapshot into the mountain-building process. *Whitman et al.* [1992, 1996] found low attenuation of regional *P* phases in the upper mantle beneath the central Altiplano and interpreted it as lithospheric mantle. Upper mantle results summarized in Figure 7 support this general conclusion but reveal some important details. Seismic tomography studies show large lateral variations in the upper mantle beneath the central Andes [*Myers et al.*, 1998; *Dorbath et al.*, 1993; *Dorbath and Granet*, 1996]. Seismic tomography between 18°S and 20°S by *Myers et al.* [1998] shows high *P* wave velocities relative to the IASPEI91 model beneath the central Altiplano and very low velocities beneath the Altiplano-Eastern Cordillera transition (Figure 7). The low seismic velocities beneath the Altiplano-Eastern Cordillera transition near 20°S suggest that mantle lithosphere is locally removed or delaminated [*Myers et al.*, 1998]. Seismic tomography between 16°S and 18°S by *Dorbath et al.* [1993] and *Dorbath and Granet* [1996] found lateral variations in the upper mantle, with high *P* wave velocities beneath the sub-Andean zone and low *P* wave velocities beneath the Altiplano-Eastern Cordillera transition (Figure 7).

[30]   Measurements of shear wave splitting have become a standard technique in the study of crustal and mantle anisotropy. A number of these studies have been in subduction zones and utilized the abundance of both crustal and subcrustal earthquakes to constrain the depth of the anisotropy [*Kaneshima and Ando*, 1989]. Recent studies have been published on the large-scale anisotropy structure of the central Andes [*Polet et al.*, 2000] and more specifically under northern Chile [*Bock et al.*, 1998]. Both studies agree that most of the observed anisotropy west of 67°W is due to mantle flow beneath the Nazca plate and that the delay time due to anisotropy above the subducting plate is less than 0.3 s. East of 66°W *Polet et al.* [2000] identified anisotropy with an east-west fast direction that they interpreted as resulting from Brazilian lithosphere.

[31]   What have we learned from these recent studies on the lithospheric structure in the central Andes? Overall the geophysical results are consistent with the Andean mountains forming as a result of weak crust deforming between the relatively strong Nazca plate and the Brazilian lithosphere. Hence our results support tectonic shortening as the primary mechanism of crustal thickening in the back arc region
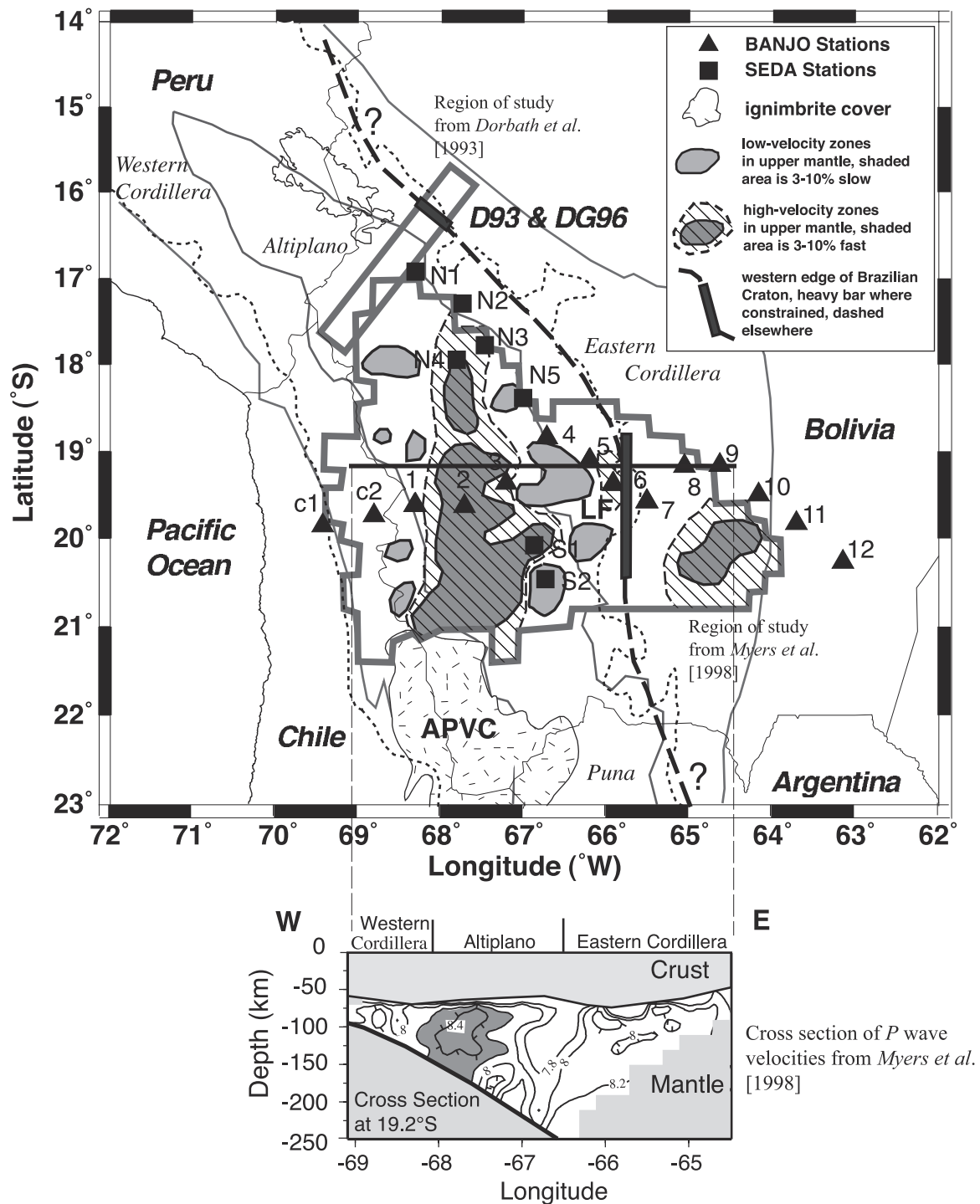
Figure 7. Map showing the upper mantle *P* wave velocity variations at a depth of 90 km from the travel time tomography study of *Myers et al.* [1998]. Western edge of the Brazilian craton determined by *Myers et al.* [1998], *Dorbath et al.* [1993] (D93), and *Dorbath and Granet* [1996] (DG96) indicated by dark gray bars. Dashed line is our interpretation of the edge of the Brazilian craton. The 3500-m elevation line is contoured with dotted lines. Also shown are the political and morphotectonic boundaries and the approximate outlines of the Altiplano-Puna Volcanic Complex (APVC). LF indicates approximate location of the Los Frailes volcanic field.
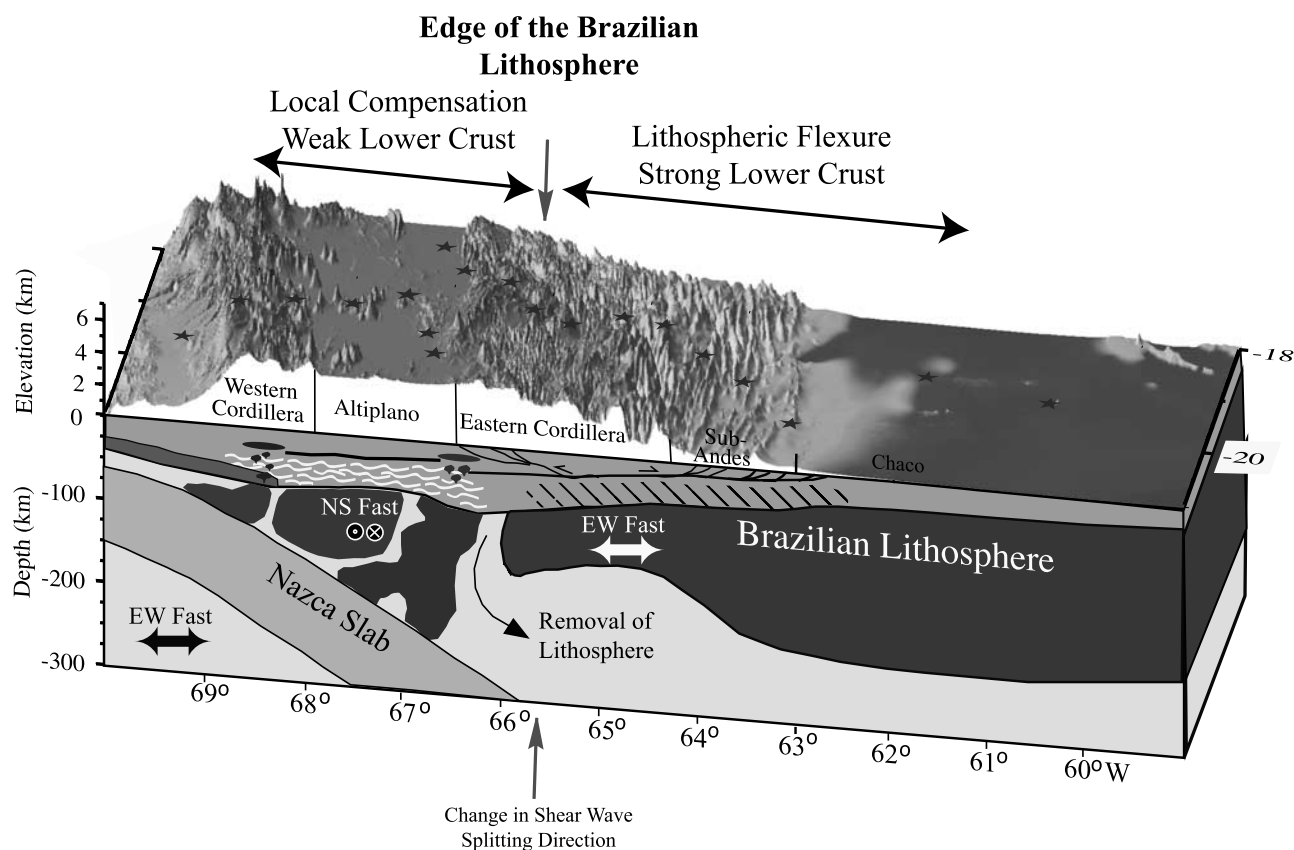
Figure 8. Schematic cross section showing our interpretation of the lithospheric structure of the central Andes from geophysical and geological studies. Red and blue indicate upper mantle *P* wave velocities that are slower and faster, respectively, than the reference IASPEI-91 model [*Myers et al.*, 1998]. See color version of this figure at back of this issue.

[*Isacks*, 1988]. Figure 8 is a schematic east-west cross section of the central Andes between 18° and 20°S showing our interpretation of the geophysical results from recent studies. The major results are as follows: (1) Overall the crust supporting the high elevations is thick and has a felsic (quartz-rich) composition; hence it is very weak. However, large north-south variations in crustal thickness and amounts of partial melt occur along the strike of the Altiplano. (2) The relatively strong Brazilian lithosphere is underthrusting as far west as the high elevations of the western part of the Eastern Cordillera (65.5°W) but does not underthrust the entire Altiplano. (3) As the Brazilian craton underthrusts the subcrustal lithosphere (and possibly the mafic lower crust), the subcrustal lithosphere and the lower crust are decoupling from the upper crust and delaminating or subducting into the mantle. The subcrustal lithosphere is delaminating piecemeal and is not completely removed beneath the central Altiplano.

4.2. Isostasy

[32] There is excellent agreement between studies using different seismic techniques that the thick crust has abnormally low seismic velocities and a low to normal Poisson's ratio in the high elevations of the central Andes [this study, *Baumont et al.*, 2001; *Swenson et al.*, 2000; *Zandt et al.*, 1996; *Wigger et al.*, 1994]. The central Altiplano has crustal thicknesses of between 59 and 64 km. We observe a local crustal welt of 75 km near the Western Cordillera-Altiplano border. The high elevations of the Eastern Cordillera have a

crustal thickness of 65–75 km and correspond to the location of the Los Frailes ignimbrite complex. The crustal thickness decreases to approximately 40 km under the sub-Andean zone and to 30 km in the Chaco. A surprising and important result is that the southern Altiplano crustal thickness is only ∼50 km in the vicinity of Uyuni, Bolivia (20.5°S). In addition, both the surface wave dispersion study [*Baumont et al.*, 2002] and the regional waveform modeling study [*Swenson et al.*, 2000] found that average seismic crustal velocities decrease southward (Figure 6). This north-south crustal thickness variation is corroborated in part by the study of *Yuan et al.* [2001]. In that study about 500 receiver functions were averaged from seismic stations operated between 1994 and 1997 in the central Andes (including the BANJO and SEDA stations). When this data set is separated into subsets north and south of 23°S the location of the Moho east of 67.5°W shifts by 10–15 km shallower in the southern subset.

[33] The east-west crustal thickness variations appear to be consistent with simple Airy isostasy [*Beck et al.*, 1996]. However, the along-strike crustal thickness variations clearly indicate that on a regional scale, upper mantle density variations or dynamic processes must be involved in the isostatic balance.

4.3. Crustal Shortening

[34] Although many workers now agree that most of the crustal thickening in the back arc region of the central

Andes is from tectonic shortening, the range in the estimates of crustal shortening determined from surface geology and balanced cross sections is still large. Many workers suggest that there is still a discrepancy between observed tectonic shortening and the amount needed to account for the entire crustal thickness [*Allmendinger et al.*, 1997; *Lamb et al.*, 1997; *Kley and Monaldi*, 1998; *Kley et al.*, 1997; *Schmitz*, 1994]. Between 18° and 20°S the minimum crustal shortening amounts during the Neogene range from 240–320 km and can account for at least 70-80% of the present crustal volume, assuming a 35- to 40-km-thick initial crust (see *Allmendinger et al.* [1997], *Kley and Monaldi* [1998], and *Lamb et al.* [1997] for summaries). These studies are primarily focused on regions east of the Altiplano because of the limited exposures on the Altiplano and Western Cordillera and the lack of subsurface information. Recently, *DeCelles and Horton* [2002], *Horton et al.* [2001], and *McQuarrie and DeCelles* [2001] have suggested that as much as 500 km of shortening across the entire mountain belt may have occurred, with a significant amount of the deformation being pre-Neogene in age.

[35] Given that the observations of tectonic shortening are minimum amounts and that uncertainty still remains about the geometry of major faults at depth, it is likely that further work will result in increased estimates of shortening. We suggest that within the uncertainties of each observation (amount of shortening and crustal thickness), no discrepancies exist, and hence there is no reason to call upon other mechanisms to thicken the crust. An important question is what happened to the mafic lower crust and the subcrustal lithosphere that most likely was associated with the upper crust, which was shortened 250 to 500 km.

4.4. Composition, Partial Melt and Rheology of the Crust

[36] Figure 9 shows a comparison of the crustal structure in the central Altiplano back arc, and the Western Cordillera arc [*Graeber and Asch*, 1999] with the global averages of *Christensen and Mooney* [1995] for similar tectonic environments. It is interesting to note that the Western Cordillera arc column from the central Andes shows thicker crust but overall similar velocities to other continental arcs including a high-velocity lower crust. In contrast, the back arc crustal column from the Altiplano shows no high-velocity lower crust and appears to be composed of thickened material with seismic velocities of upper crust when compared to other orogenic belts.

[37] The low average crustal velocities observed for the Altiplano are interpreted as an indication of a felsic, quartz-rich composition, with any mafic component restricted to, at most, a very thin layer at the base of the crust. [*Swenson et al.*, 2000; *Zandt et al.*, 1996]. The average P wave crustal velocities (5.8–6.0 km/s), the normal to low Poisson's ratio (0.25), and the lack of a high-velocity lower crust are quite different from global averages for cratonic crust [*Christensen and Mooney*, 1995; *Rudnick and Fountain*, 1995; *Zandt and Ammon*, 1995]. Figure 10 compares the compressional seismic velocity structure of the central Altiplano to the isothermal pressure-velocity gradients in the lower crust for a felsic rock (granite) at temperatures of 700°, 800° and 900°C and for an intermediate rock (diorite) at 1000°C. We show the velocity structure from regional waveform model-
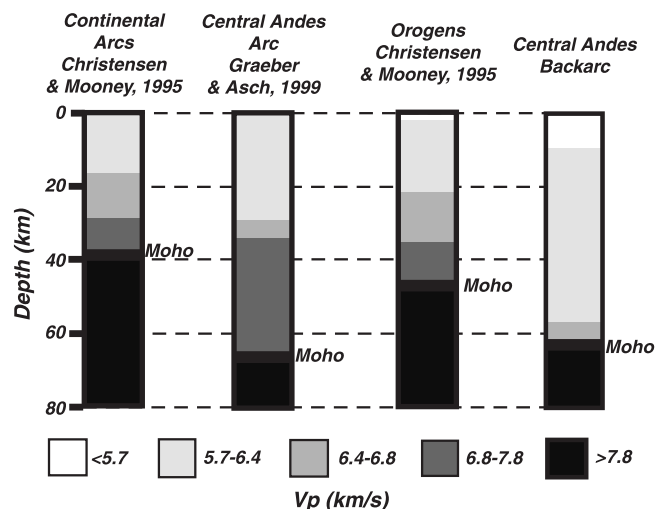

Figure 9. Comparison of the P wave crustal velocity model for the central Andes back arc (this study) and Western Cordillera from *Graeber and Asch* [1999] with the global averages of continental arcs and orogens from *Christensen and Mooney* [1995].

ing with a gradient at the base of the crust [*Zandt et al.*, 1996] and the average and standard deviation of the velocity structure determined from surface wave dispersion [*Baumont et al.*, 2002]. In felsic rocks the increase in seismic velocity is due to the transition from plagioclase-bearing assemblages to clinopyroxene-bearing assemblages. The observed velocities are most consistent with a felsic composition and temperatures of ~800°C at the base of the crust. This suggests the crustal thickening did not occur by crustal stacking of cratonic crust but resulted from shortening of material with properties typical of the middle to upper crust. The felsic crust could be an inherited property and formed long before the Andes, but more likely it is a result of back arc compression and the crustal shortening. We suggest that original mafic lower crust was depressed deep enough to undergo a phase transformation to eclogite and that it either delaminated or, if still present, has mantle seismic properties [*Giese et al.*, 1999; *Peacock*, 1996; *Christensen*, 1996]. This implies that the depth-versus-strength profile of the crust has changed through time. The end result is a very weak lower crust. Although the average seismic velocities (V_s = 3.4 km/s) and Poisson's ratio suggest weak crust, widespread melt (less than a few percent) does not occur in the lower crust of the central Altiplano [*Swenson et al.*, 2000]. Recent studies of lower crustal xenoliths from northern Tibet show that anhydrous lower crust need not be molten in spite of extreme temperatures [*Hacker et al.*, 2000].

[38] In contrast, an abrupt ~15 km decrease in crustal thickness south of 22°S corresponds to the Altiplano-Puna volcanic center, a mid-crustal magma body, and an ~1 km increase on average elevation. [*Zandt et al.*, 2002; *Yuan et al.*, 2001]. The southern Altiplano is a transition zone to the Puna, where the average elevation is nearly 1 km higher than the central Altiplano. This would suggest that some of the difference in crustal structure is probably the result of a higher-temperature upper mantle beneath the Puna [*Whitman et al.*, 1996].
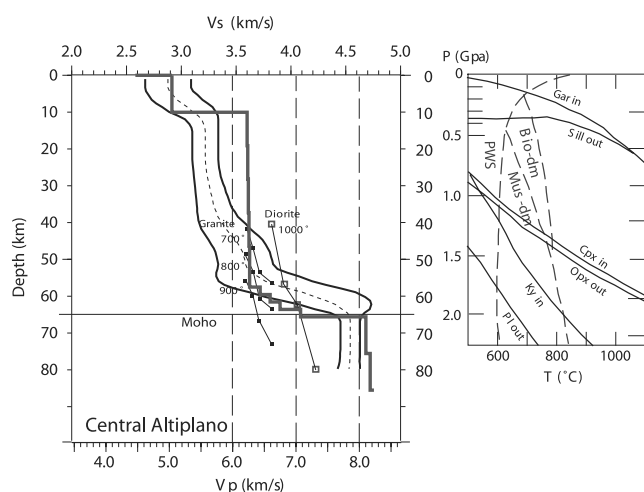
Figure 10. Seismic velocity structure determined from regional waveform modeling (thick line, *Zandt et al.* [1996]) and surface waves (thin lines, *Baumont et al.* [2002]) for the central Altiplano compared to isothermal pressure-velocity gradients for a felsic rock (granite) for temperatures of 700°, 800°, and 900°C, and an intermediate rock (diorite) at 1000°C. On the right is the theoretical phase diagram for a granite bulk chemical composition, which forms the basis for the calculation of the elastic wave velocity gradients following *Sobolev and Babeyko* [1994]. Superposed on the phase diagram are the melting curves (PWS, pelitic wet solidus; Mus-dm, muscovite dehydration melting; Bio-dm, biotite dehydration melting) for a metapelite [*Thompson and Connolly*, 1995]. In felsic rocks the increase in velocities are due to the transition from plagioclase-bearing assemblages to clinopyroxene-bearing assemblages (e.g., albite-to-jadeite). Both the initiation depth and depth range (width) of the transition are dependent on composition, temperature, and pressure. The observed velocities are most consistent with a felsic composition and temperatures between 700° and 800°C near the base of the crust. The lack of significant melt in the lower crust of the central Altiplano (see text) suggests an anhydrous lower crust, but the presence of hydrous minerals is uncertain. Gar, garnet; Sill, sillminite; Cpx, clinopyroxene; Opx, orthopyrexene; Ky, kyanite; Pl, plagioclase.

[39] There is still uncertainty about the crustal thickness and the lower crustal composition between 21° and 22°S. *Wigger et al.* [1994] also found low crustal velocities at 21°s using data from a refraction survey but they were unable to image the Moho. *Wigger et al.* [1994] and *Schmitz et al.* [1997] interpreted the low velocities in the lower crust as a mafic composition with a significant amount of partial melt (approximately 16%). Our seismic data are north of 21°S or near 22°S, so we cannot make a direct comparison with the receiver function data in the vicinity of the *Wigger et al.* [1994] refraction line (Figure 1). It is possible that the lower mafic crust has not delaminated in this vicinity, but it seems more likely that the lower crust is similar to the central Altiplano and more felsic in composition with a much smaller percentage (<5–10%) of partial melt.

[40] If the crust has predominantly a felsic composition, then we would expect it to be extremely weak, and we

would expect a significant amount of lower crustal flow [*Royden*, 1996]. The heat flow in the Altiplano and high elevations of the Eastern Cordillera is high, averaging 80 mW/m², although the scatter in the measurements is very large [*Springer and Forster*, 1997; *Henry and Pollack*, 1988]. In contrast, the heat flow in the sub-Andean zone and the Chaco is 40 mW/m², consistent with continental cratons [*Springer and Forster*, 1997]. The high heat flow in the Altiplano could be the result of asthenosphere at shallow levels and/or increased heat production as result of crustal thickening [*Springer and Forster*, 1997]. Crustal thickening initially produces cold crust and then increased heat production if the composition is appropriate. *Le Pichon et al.* [1997] determined lithospheric thermal models through time for doubling the thickness of continental crust. They showed that if you increase the crustal thickness to 60–70 km and use an intermediate lower crustal composition with the lithosphere attached, it requires more than 30 m.y. and closer to 40 m.y. for the mid to lower crust to heat up to the point of melting. This, combined with at least some mantle lithosphere beneath the central Altiplano, could explain why the mid to lower crust does not contain widespread melt but does suggest that the crust should be warm and very weak. The southern Altiplano is a transition region between the Altiplano and the Puna, where we expect the mantle lithosphere to be thinned based on work by *Whitman et al.* [1996]. Some partial melt in the lower crust in the southern Altiplano may in part result from the asthenosphere being closer to the base of the crust.

[41] At the Altiplano-Eastern Cordillera transition (near 19°–20°S) we also observe indications of at least some localized partial melt in the midcrust. We see very low velocities at 14–20 km depth, high *Lg* attenuation, and young volcanic rocks at the surface (Los Frailes ignimbrite complex) derived in large part from crustal melts [*de Silva and Francis*, 1991]. We also observe very low upper mantle seismic velocities under the Altiplano-Eastern Cordillera transition, suggesting asthenosphere near the base of the crust [*Myers et al.*, 1998]. All of this suggests that some localized partial melt occurs in the midcrust under the high elevations of the Eastern Cordillera.

4.5. Underthrusting of the Brazilian Craton

[42] The mechanically strong Brazilian lithosphere has underthrust to approximately 65.5°W beneath the high elevations of the Eastern Cordillera and does not extend all the way across the Altiplano. Several types of data suggest such an interpretation. First, the *Myers et al.*'s [1998] seismic tomography results suggest low *P* wave velocities beneath the Altiplano-Eastern Cordillera boundary, which are not consistent with cratonic lithosphere. In the vicinity of 16°S, *Dorbath et al.* [1993] used local seismic tomography and found a strong lateral change in upper mantle *P* wave velocities beneath the Eastern Cordillera, with high velocities to the east and low velocities to the west under the Altiplano. Second, the shear wave anisotropy results of *Polet et al.* [2000] show a change in the fast direction from north-south to east-west at 65.5°W, consistent with the edge of the Brazilian lithosphere. Third, studies of the Bouguer gravity anomalies in the vicinity of 20°S show that flexure of the lithosphere is important in supporting the sub-Andean zone and parts of the Eastern Cordillera

[*Watts et al.*, 1995; *Lyon-Caen et al.*, 1985; *Whitman et al.*, 1996]. The best overall fit to the gravity profile is for an effective elastic thickness of 75 km and a fractured plate beneath the western part of the Eastern Cordillera [*Watts et al.*, 1995]. Finally, Pb isotope studies of volcanic rocks in the central Andes show distinct isotopic basement domains [*Aitcheson et al.*, 1995] between the Altiplano and the high elevations of the Eastern Cordillera. All of these studies indicate that the strong Brazilian lithosphere extends beneath the Andes to approximately 65°–66°W but does not underlie the Altiplano.

[43] From these structural interpretations we infer two different mechanisms of deformation across the central Andes. East of 65.5°W, the crust and mantle are coupled and moving west, compressing the high elevations. The shortening in the upper crust has been taken up on discrete thrust faults that sole into the top of an underthrusting cold, strong Brazilian lithosphere [*Lamb et al.*, 1997; *Baby et al.*, 1997]. The Altiplano and high-elevation regions of the Eastern Cordillera have an upper crust and mantle that are decoupled from one another by a weak, felsic lower crust. The thick Altiplano lower crust is dominated by distributed ductile deformational mechanisms as a result of the Brazilian craton moving into it. Lower crustal flow contributes to the subdued surface topography in the central Altiplano. This is consistent with models by *Royden* [1996] in which detachment of the crust from the mantle is essential to the formation high, flat topography.

4.6. Eclogitization and Delamination

[44] Mountain belts associated with convergent margins have long been thought to be sites of continental recycling or reworking from delamination driven by density contrasts [*Kay et al.*, 1994; *Rudnick*, 1995]. However, delamination is a difficult process to document and probably occurs at different stages of mountain belt evolution depending on the thermal history and composition. For example, farther south near 24°–27°S geochemical and geophysical data suggest that lithospheric delamination occurred at 2–3 Ma [*Kay and Kay*, 1993; *Kay et al.*, 1994; *Whitman et al.*, 1996]. We speculate that in the central Andes we are seeing localized delamination driven by density instabilities due to the eclogitization of mafic lower crust. The lack of a high-velocity lower crust (mafic composition in the granulite facies) implies that the mafic lower crust is now eclogite and has seismic velocities similar to those of the mantle or that it has been delaminated.

[45] One location where we would argue for delamination is beneath the Altiplano-Eastern Cordillera transition at 20°S, where we observe thick crust (65–75 km). The large amount of localized shortening depressed the lower crust into the eclogite facies and transformed it to a higher density. With increased shortening this caused the lower crust and upper mantle lithosphere to become unstable and delaminate, resulting in warm asthenosphere near the base of the crust as observed in the tomography [*Myers et al.*, 1998]. We envision this as a piecemeal process that was localized in the region with the largest amount of shortening. The increased heat flux from the mantle and the depression of upper crustal material to deeper levels can account for the crustal melts in the midcrust that produced the Los Frailes ignimbrites.

[46] In the Altiplano we see a north-south variation in crustal thickness that could indicate varying amounts of delamination. The relatively high mantle seismic velocities beneath the central Altiplano crust are consistent with mantle lithosphere, but the high-velocity material may also include some lower mafic crust that has been depressed to depths where it has transformed to eclogite. The seismic velocities are similar for both eclogite and mantle lithosphere [*Christensen*, 1996; *Giese et al.*, 1999]. Hence the seismic Moho that we have mapped could be the old upper crust-lower crust boundary, implying that the entire 60 km of crust has been built from thickening of the mid to upper crust. This would explain why the crustal velocities are so low and why we do not see a high-velocity lower crust. In the southern Altiplano the delamination process may have already occurred. This would account for the thinner crust, more partial melt in the crust, and higher elevations. The lithospheric thinning would have contributed to the uplift as asthenosphere replaced higher-density mantle lithosphere.

[47] Several studies suggest that the elevations of the Altiplano and Eastern Cordillera were only 1–2 km at 10 Ma [*Gregory-Wodzicki et al.*, 1998; *Kennan et al.*, 1997; *Gubbels et al.*, 1993]. If correct, this suggests that at least 2 km of uplift has occurred in the last 10 m.y. *Le Pichon et al.* [1997] suggested that an intermediate-composition lower crust could initially transform to the eclogite facies (if enough shortening occurred to depress it deep enough), and then as it warmed up it could transform back to granulite facies. Felsic as well as mafic compositions can undergo transformation with corresponding density changes in the ecologite facies. The decrease in density with the eclogite to granulite transformation could contribute to the uplift of the Altiplano in the last 10 m.y. [*Le Pichon et al.*, 1997]. It is difficult to determine the thermal history of the Andes, but phase transformations should be considered when interpreting their structural history.

[48] Overall the geophysical results are consistent with *Isacks*'s [1988] and *Lamb et al.*'s [1997] two-stage deformation model, with some modification. *Isacks* [1988] proposed that lithospheric thinning weakened the crust, allowing widespread horizontal shortening and vertical thickening of the crust. Deformation along the western edge of the South American continent started at least 60 m.y. ago near the Western Cordillera and propagated toward the east [*Horton and DeCelles*, 1998; *Horton et al.*, 2001; *DeCelles and Horton*, 2002]. As thickening continued and the mafic lower crust reached a depth of 45–50 km, it transformed to eclogite. As the mafic lower crust transformed to eclogite, the upper and lower crust were decoupled and thickened, resulting in a much weaker crust with a felsic "lower crust" that deformed. The rapid uplift in the last 10 m.y. may be in part due to (1) the transformation of some of the deepest felsic lower crust from eclogite facies back to granulite facies with increasing temperature, (2) lower crustal flow, or (3) both. The Brazilian craton east of 65.5°W compresses this weak lower crust under the Altiplano, causing more thickening and rapid uplift with very little surface deformation. The corresponding surface deformation is transferred to the east as the upper crust thrusts over the Brazilian craton [*Isacks*, 1988; *Lamb et al.*, 1997]. The subcrustal eclogite and mantle lithosphere are in the process of delaminating under the Altiplano-Eastern Cordillera boun-

dary and the southern Altiplano, but they have not yet delaminated completely beneath the central Altiplano. In fact, based on the amount of upper crustal shortening (up to 500 km), it seems likely that a significant amount of lower crust and mantle lithosphere has been recycled into the mantle, resulting in a very felsic, quartz-rich crust.

5. Conclusions

[49] We have used receiver function analysis combined with other geophysical studies to characterize the nature of the continental crust and mantle lithosphere in the central Andes. The major results are as follows:

1. Crust underlying high elevations is thick and has a felsic (quartz-rich) composition; hence it is very weak. In the high elevations along an east-west profile at 18°–20°S, the crustal thickness and elevations (average of 3.8 km) indicate compensation by Airy isostasy. The seismic velocities and Poisson's ratio indicate a felsic- to inter-mediate-composition crust with little or no mafic lower crust. This suggests that any mafic lower crust has been delaminated or transformed to eclogite during the orogenic process.

2. The relatively strong Brazilian lithosphere under-thrusts as far west as the high elevations of the western part of the Eastern Cordillera (65.5°W), but it does not underthrust the entire Altiplano.

3. The subcrustal lithosphere is delaminating piecemeal under the Altiplano Eastern-Cordillera boundary, but it is not completely removed except under the central Altiplano. The large amounts of mid- and upper crustal shortening suggest that delamination near the Altiplano-Eastern Cordillera boundary accommodates a significant amount of subcrustal shortening between the Altiplano and Brazilian lithosphere.

[50] From our study we suggest that the back arc in the central Andes has undergone a process that has decoupled the mid and upper crust from the mafic lower crust and mantle lithosphere through time. This orogenic reworking may be an important process in the "felsification" of continental crust.

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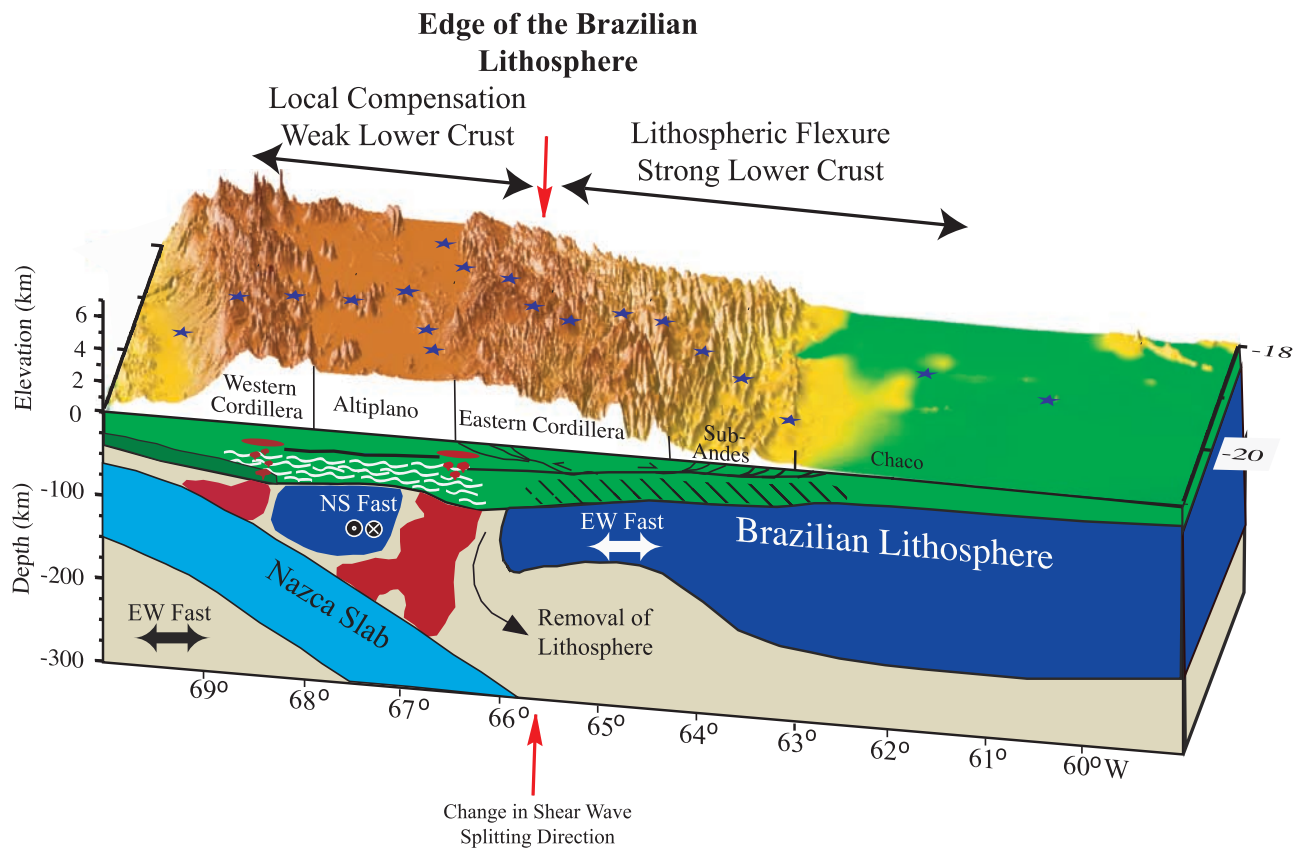
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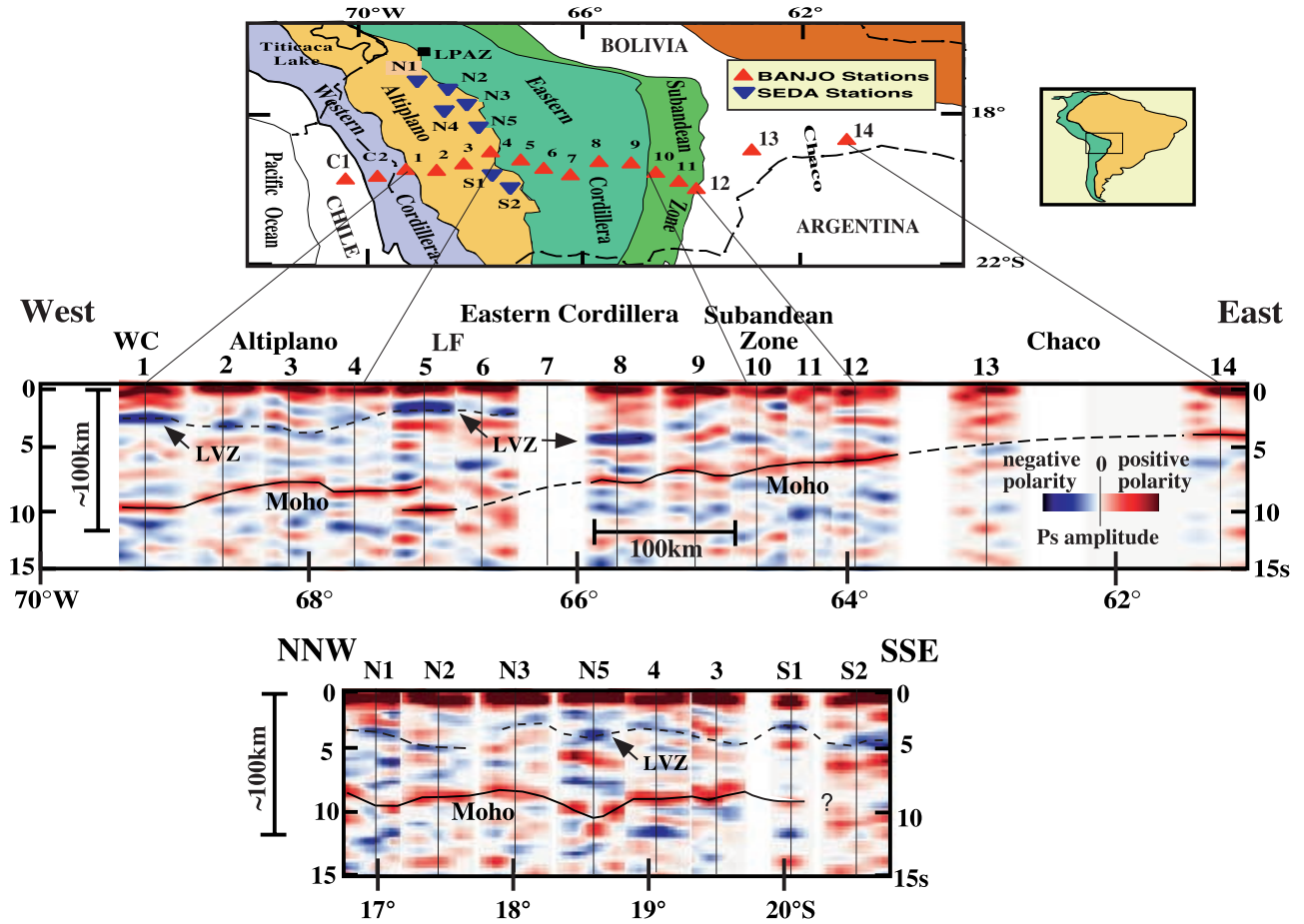
Figure 4. Psuedo cross sections of receiver function traces along (top) W-E and (bottom) NNW-SSE profiles. For each station, three receiver function stacks corresponding to different backazimuths were projected onto the respective record sections. Positive polarity *Ps* conversions plot as red, and negative polarity *Ps* conversions plot as blue. Note the Moho *Ps* observed beneath most stations except where rapid changes occur in crustal thickness. Several low-velocity zones (LVZ) occur in the midcrust beneath the Altiplano and Eastern Cordillera.

Figure 8. Schematic cross section showing our interpretation of the lithospheric structure of the central Andes from geophysical and geological studies. Red and blue indicate upper mantle P wave velocities that are slower and faster, respectively, than the reference IASPEI-91 model [*Myers et al.*, 1998].