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Crustal thickness in the central Andes from teleseismically recorded depth phase precursors

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SUMMARY

Although crustal thickness is a critical constraint for geodynamic models of Andean orogenesis, relatively few measurements exist from local seismic studies. In this study, we use reflections from the underside of the Moho as illuminated by intermediate depth earthquakes to provide new estimates of crustal thickness for the central Andes (16–34°S). These reflected signals (pmP) are identified as precursors to the depth phase pP when recorded at teleseismic distances (35–85°). Although relatively small in amplitude, the pmP phase is often clear even on single seismograms. Less obvious pmP phases were enhanced by stacking traces from arrays. This method was most effective for events of $M_w > 6$ and depth > 100 km. Crustal thickness determined in this study ranges from 59 to 70 km in southern Peru, from 49 to 80 km across the Puna-Altiplano and from 50 to 60 km above the Pampean flat slab. The lack of such phases for certain events may be evidence of heterogeneity at the Moho, and some precursors appear to correspond to intracrustal (magma?) and lithospheric mantle discontinuities. The results obtained using the pmP technique combined with those from other studies show the crustal thicknesses under the Central Andes are more variable than expected from purely isostatic considerations. We infer that one or a combination of geodynamic processes involving lower crustal flow, lower crustal delamination, and mantle–crustal lithospheric coupling can explain the highly variable topography on the Moho beneath the Central Andes.

Key words: Earthquake source observations; Body waves; Subduction zone processes; South America.

INTRODUCTION

The Andes are the type example of a mountain range constructed by crustal thickening and shortening that result from a non-collisional, subduction zone orogen. Extending some 8000 km from Venezuela to Tierra del Fuego, they are the dominant morphological feature of South America (Fig. 1). The Andes include the earth's second greatest (after Tibet) continental plateau (in terms of height and areal extent), the highest active volcanoes (>6800 m), the highest peaks outside of the Himalayas (>7000 m), and some of the thickest crust reported anywhere (>80 km; e.g. Zandt *et al.* 1996; Yuan *et al.* 2002).

The Central Andes also serve as a natural laboratory for associating surface tectonics with such deep processes as changing subduction dip, lithospheric delamination and subduction erosion. The history of the Andes is the product of a complex overlay of magmatism, uplift due to contractional deformation, intervening episodes of oblique extension, collision of oceanic terranes (in the north), formation of sedimentary basins, mineralization, loss of continental crust by forearc subduction erosion and ultimately removal of the base of thickened crust by delamination (see reviews in Allmendinger *et al.* 1997; Kay *et al.* 1999). The geodynamic mechanisms responsible for this history of complex crustal reworking and thickening, and the amount, timing and fate of removed continental crust and mantle lithosphere are the focus of ongoing debate.

The crustal thickness of the Andes is amongst the greatest in the world (e.g. Beck & Zandt 2002; Yuan *et al.* 2002). Isacks (1988) argued for a relatively simple relationship in which the high crustal thickness in the Central Andes could principally be explained by east–west shortening. Assuming that Airy isostatic balance applies to most of the modern Altiplano-Puna and Eastern Cordillera, the geologic record of shortening should serve as a proxy for surface elevation through time. Kley and Monaldi (1998, 2002) argue, however, that there are anomalous regions where the estimates of shortening and thickening cannot account for the observed elevation solely based on isostasy. Various geodynamic processes have been proposed to account for such discrepancies including, but not limited to, lower crustal channel flow (Isacks 1988; Clark & Royden 2000; Gerbault *et al.* 2005), and crustal and mantle lithospheric delamination (Kay & Kay 1993). A key test of many of these models is crustal thickness, which is still poorly known in parts of the central Andes. For example, no data are yet available between the latitudes 26°S and 30°S.

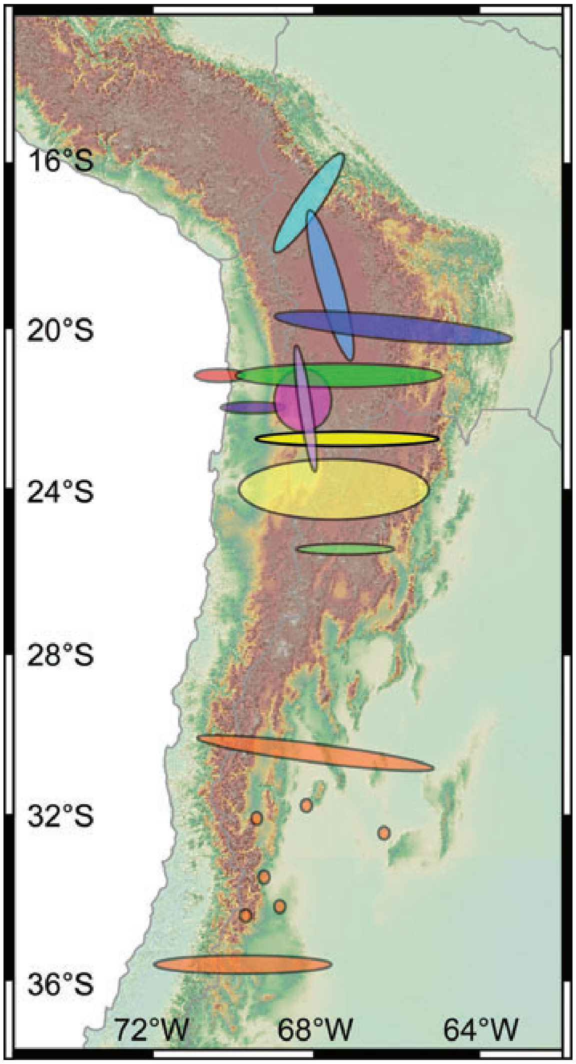

Figure 1. Previous seismic investigations in our study area: Dorbath *et al.* 1993 (turquoise); SEDA (light blue); BANJO (dark blue); ANCORP (green); CINCA (red); CALAMA (purple); PAMB (Pink); PISCO (light purple); REFUCA (light green); PUNA (yellow) and CHARGE (Orange).

Defining crustal structure, especially thickness, has been the goal of a number of geophysical surveys in the Andes. These studies (Fig. 1) include the SEDA and BANJO experiments (Zandt *et al.* 1996; Swenson *et al.* 2000; Beck & Zandt 2002), the PISCO array (Graeber & Asch 1999; Schmitz *et al.* 1999), the CALAMA experiment (Graeber & Asch 1999), the ANCORP line (ANCORP Working Group 1999, 2003), the CHARGE study (Fromm *et al.* 2004), the PAMB array (Chmielowski *et al.* 1999, Zandt *et al.* 2003), the PUNA arrays (Yuan *et al.* 2000, 2002), a study of the northern Altiplano (Dorbath *et al.* 1993) and the REFUCA experiment (Heit *et al.* 2006). These experiments have involved a variety of active (controlled source) and passive (receiver function) seismic methods. Two significant findings to date are that: (i) Moho topography is surprisingly variable with depths ranging from ∼60 to 80 km beneath the Puna-Altiplano to ∼50 to 60 km above the Chilean flat-slab crust and (ii) the Andean crust is unexpectedly felsic in composition (Zandt *et al.* 1996; Swenson *et al.* 2000; Yuan *et al.* 2002). Although field experiments in the Andes will undoubtedly continue provide additional coverage, here we use the extensive existing teleseismic recordings to substantially expand the number of crustal thickness measurements in the central Andes, without the expense of a local field campaign.

METHOD

Our study builds upon a method pioneered by Zhang & Lay (1993), which was subsequently reconfigured and applied to the Andes by Zandt *et al.* (1994). The method is based upon the identification of reflections from the underside of the Moho of seismic waves originating from intermediate to deep (sub-Moho) earthquakes (Fig. 2). Zhang & Lay (1993) focused on *S*-wave reflections (specifically SH). Zandt *et al.* (1994) used both *P* and *S* waves. In this study, we concentrate on the analysis of *P* waves because their higher frequencies yield greater resolution and their arrival times can be more accurately measured than those of the later *S* waves. We began by extracting the portion of the seismic recording that spans the onset of P and pP, as estimated by the seismic traveltime program TauPtime using the ak135 earth model (Crotwell *et al.* 1999). This program generates arrival times for the main phases expected from an earthquake, including P, pP, PcP, for given focal depths and distances. Distinct energy arrivals between P and pP that could not be

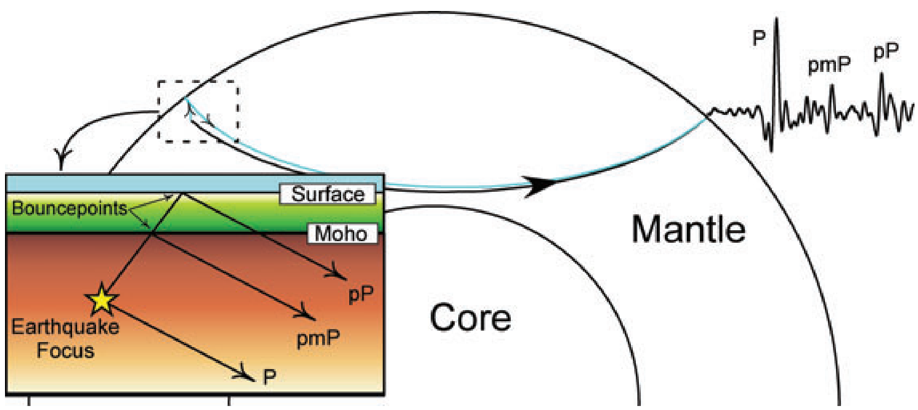

Figure 2. Illustration of reflection of upgoing *P* wave reflecting from the underside of the Moho (pmP) and surface (pP). Note that the ray paths for all three are virtually identical away from the source/reflecting area, eliminating receiver and distal paths effects as an explanation for the differential characteristics of these phases.

associated with conventional phases such as PcP were provisionally interpreted as reflections from the Moho (pmP). Although such arrivals are often clear on individual traces (e.g. Fig. 2; Kind & Seidl 1982) signal stacking (e.g. Figs 5 and 6; Zandt *et al.* 1994) was utilized to enhance weaker arrivals.

The measured delay time between P and pP (t_{pP} and t_{pmP}) was converted to crustal thickness using the formula:

$$t_{pP} - t_{pmP} \approx 2h\sqrt{V_{Pc}^{-2} - P^2},$$

where h is the thickness of the crust, V_{Pc} is the velocity of the P wave in the crust and P is the approximate ray parameter for the phases. Like Zhang & Lay (1993) and Zandt *et al.* (1994), we assumed that the ray parameter for the incumbent pP phase is identical to that of the pmP phase. The crustal velocities used for conversion were based upon the limited measurements from field experiments reported in the literature (see below) and our own analysis of Pg traveltimes from the ISC database (Fig. 7).

The method takes advantage of the recent proliferation of digital seismic networks and single stations over the globe, but especially in North America. Earthquakes of sufficient depth and magnitude are needed to provide an energetic upgoing P wave. A minimum magnitude of 6 (M_w) was found to be desirable, although some useful results were obtained from events as small as $M_w = 5.8$, depending upon the epicentral distance to the recording station. A minimum depth of 100 km is needed to allow enough time separation to distinguish the seismic phases P, pmP and pP from each other and their respective codas. We used the vertical component of seismograms from stations between 30° and 85° from the source to avoid triplication effects and possible interactions of arrivals with the D″ layer at the core–mantle boundary. Ray paths for pP and PmP were assumed to be sufficiently similar that propagation delays away from the region of reflection could be ignored.

The points where upward travelling *pP* waves reflect off the Earth's surface or Moho are referred to below as bounce points. The bounce points that correspond to the measurements in this study are shown with their accompanying calculated depth in Fig. 10. Over 8000 individual seismic traces originating from 57 events spanning the Andes from 16°S to 40°S were examined (e.g. Fig. 3). The reflection (bounce point) locations, projected to the surface, for the pmP phases were calculated using the relationship between the ray parameter P (calculated using the TauPtime program), the angle of incidence I and the mantle velocity V.

$$P = \sin\ I/V.$$

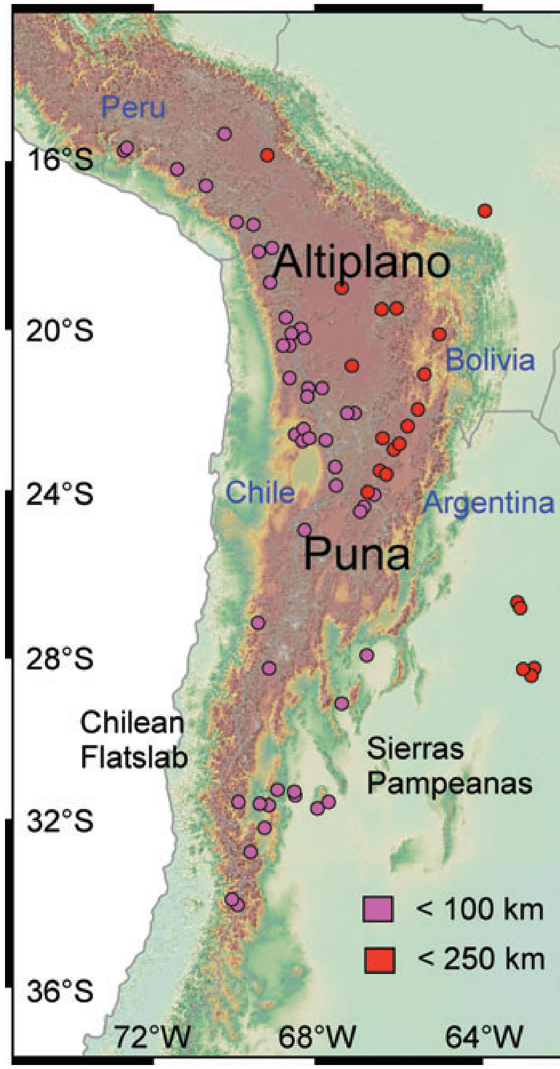

Figure 3. Epicentral locations of the 69 intermediate to deep earthquakes used in this study. Events are all greater than M_w 5.8.

Stacking and phase moveout

The key to this study is the identification of pmP. Although distinct pmP waveforms were evident on a number of individual station records (e.g. Fig. 4), the pmP phase is usually weak. Therefore, confident identification often requires signal enhancement before picking Stacking is a common means of enhancing minor precursor signals (e.g. Shearer 1991, 1993; Zhang & Lay 1993; Zandt *et al.* 1994; Flanagan & Shearer 1998). The Terrascope/Caltech network in southern California is particularly useful for signal stacking. This network has operated almost continuously for the past 15 yr and has had a varying, but always significant, number of stations situated sufficiently close together to produce robust stacks. Ray paths for particular earthquakes at differing stations are so similar that the pmP reflection bounce points can be assumed to be from the same crustal area and thus are justifiably stacked and summed. We also took advantage of several temporary seismic deployments in the western United States with relatively small interstation separations which 'caught' suitable events from the Andes.

Effective stacking requires alignment of the seismic arrivals of interest. Zandt *et al.* (1994) used slant-stacking techniques to correct for linear moveout effects. We calculated the estimated arrival times of pP for the various Terrascope stations using the program TauPtime and found the time differential between the closest and furthest bounce points to be only 0.2 s. Assuming that pP is a suitable proxy for pmP in this context, such differentials are insignificant compared with the signal periods of over 2 s, and thus were neglected.

The data recorded by Terrascope from a M_w 6.3 earthquake that occurred on 1993 June 8 at 31.56°S, 69.24°W at a depth of 112 km is shown in Fig. 5. Although this event shows clear pmP energy on single traces at various backazimuths, the amplitude of the phase can be greatly enhanced by stacking. Assuming an average P velocity of 6.5 km s^{-1}, a crustal thickness of 57 km was calculated at the bounce point corresponding to this event. The assumed P velocity is close to a Pg velocity reported from the nearby CHARGE experiment (Alvarado *et al.* 2005).

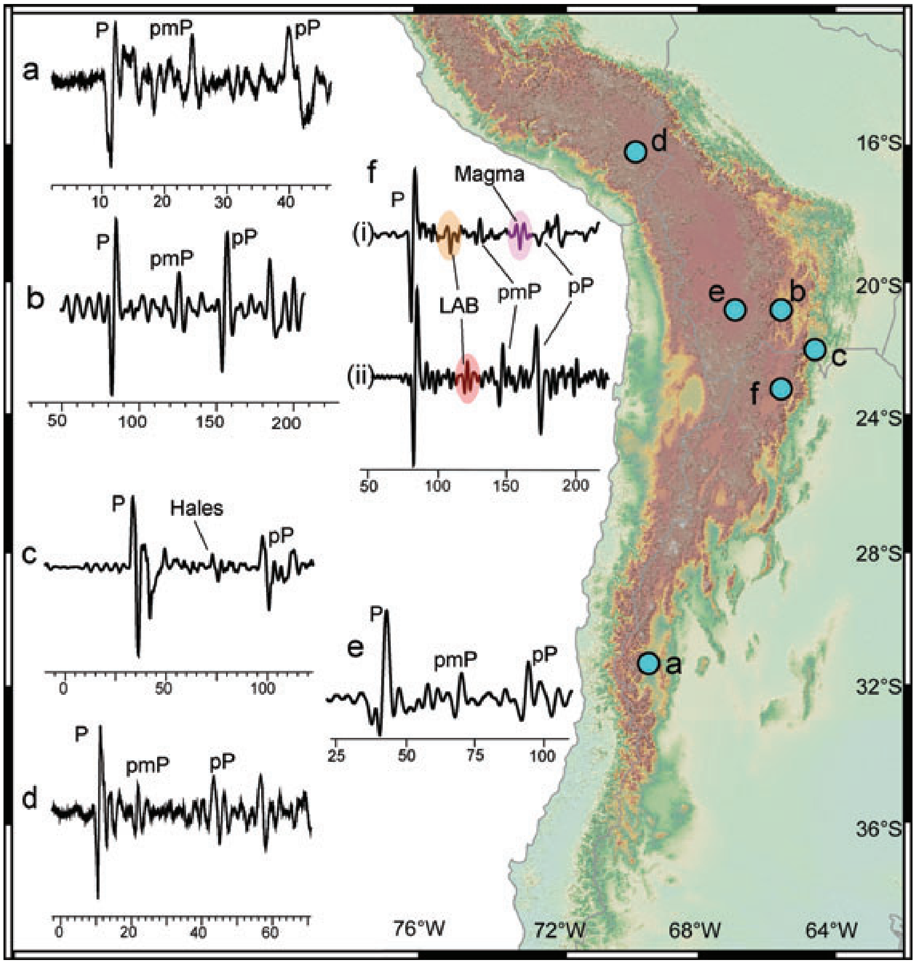

Figure 4. Seismograms illustrating examples of precursor phases. These events occurred on: (A) 1993 June 8, (B) 2004 March 17, (C) 1997 January 23, (D) 1998 October 8, (E) 1999 September 5 and (F) 1997 July 20. Examples (A), (B), (D) and (E) all show clear Moho reflection reflections (pmP) as well as surface reflections (pP). Example (C) contains an early arrival (indicated) interpreted to come from a deeper interface (the Hales discontinuity?). The two traces in example (F) are from the same event recorded at two different stations. Both show arrivals from a deeper interface (indicated in orange and red) that could mark the lithosphere–asthenosphere boundary. The top trace (i) also shows an intracrustal reflector (pink) which we suggest could mark a magma layer.

Another example is detailed in Fig. 6 which shows the results from the analysis of an M_w 7.2 earthquake that occurred on 2000 May 12 at 23.5°S, 66.4°W at a depth of 225 km. This event was recorded by five different arrays, including the Terrascope/Caltech network, the Ristra deployment, North and South Colorado (CDROM) and Montana. Data from each array were stacked and Pp-PmP times measured. The times were converted to depth using an average crustal velocity of 6 km s^{-1}, which has been widely reported as the average crustal velocity for the Altiplano plateau (Zandt *et al.* 1996; Schurr *et al.* 1999; Beck & Zandt 2002). In spite of the diverse distribution of recording arrays, the corresponding bounce points cluster within 30 km of each other, and the resulting crustal thickness estimates are similar, only varying from 59 to 62 km (Fig. 6). The consistency among the results from these arrays, despite the variations in signal strength of pmP, shows that the technique is relatively robust.

Crustal velocity

The primary ambiguity in converting pP-pmP time delays to crustal thickness lies in assuming a conversion velocity for the crustal travel path for an event. Many authors have argued that the Altiplano-Puna plateau is anomalously felsic in composition, and various studies have reported average crustal *P*-wave velocities as low as ∼6 km s^{-1} (Zandt *et al.* 1996; Schurr *et al.* 1999; Beck & Zandt 2002). Crustal velocity studies for the remainder of the Andes are limited, but recent work by Alvarado *et al.* (2005) above the Chilean flatslab between 28°S and 33°S indicates a Pg ∼ of 6.3 km s^{-1} in that region. As an independent guide, time–travel plots were created based on first-arrivals from low magnitude local earthquakes listed in the ISC bulletin in two areas where earthquakes were used in this study: 20–25°S and 28–33°S. An example of this study is detailed in Fig. 7. Although broad in scope and lacking in azimuthal coverage, these gross traveltime curves suggest a Pg velocity of ∼6 km s^{-1} in the northern zone and ∼6.5 km s^{-1} in the southern zone. These values are generally consistent with previously published results and were used for depth conversions in the two regions although it should be noted that using the crustal velocity reported by Alvarado *et al.* (2005) for events around the Pampean flatslab will result in a reduction of crustal thickness on the order of ∼1 km.

Cross-calibration

In order to assess the consistency of the results obtained here with those from other techniques, we compiled a comparison for all areas where independent crustal thickness estimates were available. For

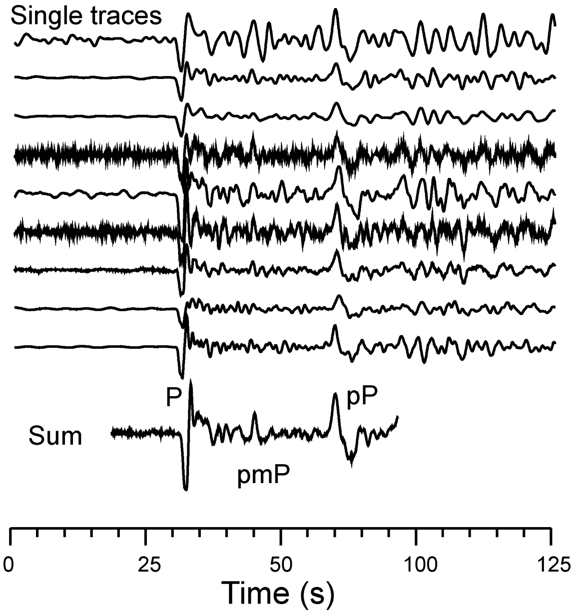

Figure 5. Top: Terrascope array recordings from an earthquake occurring on 8 June 1993, M_w 6.3, in the shallow subduction zone beneath the Calingasta valley, Argentina (31.6°S, 69.2°W). Bottom: Straight stack of array seismograms, showing substantial enhancement of reflected Moho phase.

the comparison, our estimates of crustal thickness were averaged over the same 1° squares as those for the receiver function and wide angle reflection estimates reported by Yuan *et al.* (2002). As seen in Fig. 8(i) the results from these different methods are

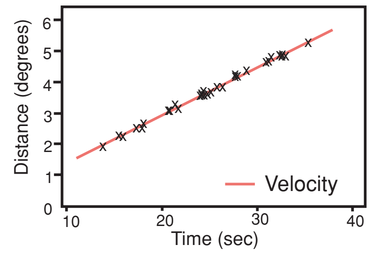

Figure 7. Plot of Pg traveltimes versus distance for stations between 28 and 33°S from the ISC bulletin. The slope of this plot provides a crude estimate of crustal velocity (6.2 km s^{-1}), which we have used to convert measured pP-pmP traveltimes to depth.

very similar This agreement is even more impressive given that the receiver function thicknesses are strongly dependent on estimates of average shear wave velocity, whereas the results reported here are dependent on assumed average compressional wave velocity. Thus the differences in thicknesses that do exist could be attributable to lateral variations in Poisson's ratio.

In Fig. 8(ii), the estimates of crustal thickness derived from the pmP method are compared against those of the ANCORP group (2003) in a crustal cross-section. The ANCORP study calculated crustal thicknesses based on receiver function and deep reflection profiling. Except for two outliers, the pmP-derived estimates agree well with those on the ANCORP profile. The outliers raise the issue of whether the Moho as delineated by pmP marks the same interface 'seen' by surface source seismic or receiver function techniques. There is room for ambiguity between the methods given the very different seismic frequencies involved between the deep

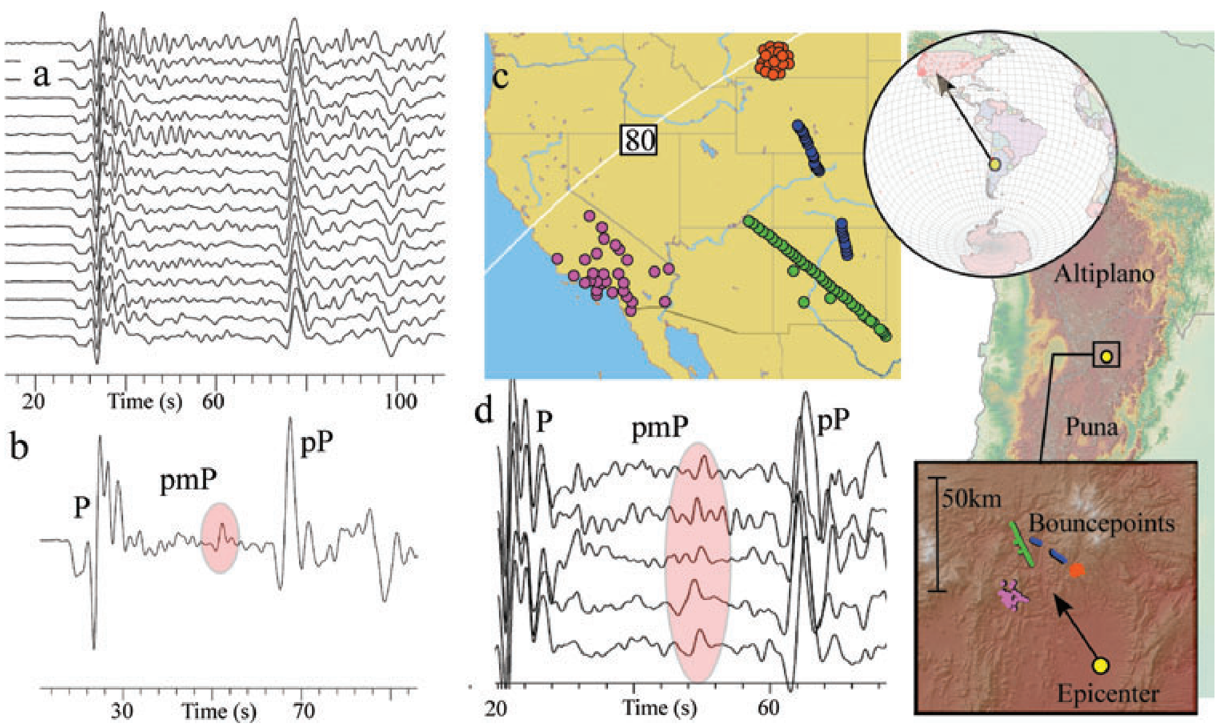

Figure 6. Fortuitous recordings of a M_w 7.2 earthquake occurring on 2000 May 12 by temporary arrays in New Mexico, northern and southern Colorado and Montana, as well as Terrascope, all about 80° from the epicentre (C). Individual traces (A) from Terrascope, summed to yield the composite in (B). Comparison of stacked composites (D) from the four arrays (top to bottom: Montana, California, New Mexico, northern Colorado and southern Colorado). The map at right shows the epicentral locations, the azimuth to the arrays and the bouncepoints for the Moho reflections for the various arrays.

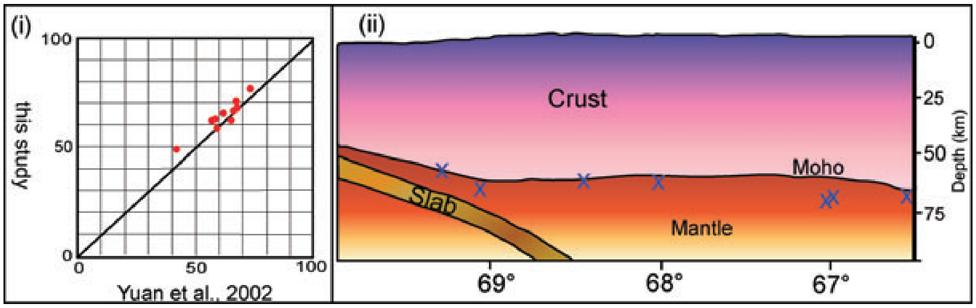

Figure 8. (A) Comparison of crustal thickness estimates from this study with those from the receiver function study of Yuan *et al.* (2002), demonstrating the general agreement between the two independent data sets. (B) A Cross-section through the central Andes showing Moho topography inferred from receiver functions (modified from ANCORP 2003). Crustal thicknesses from this study represented by crosses, again exhibiting general agreement.

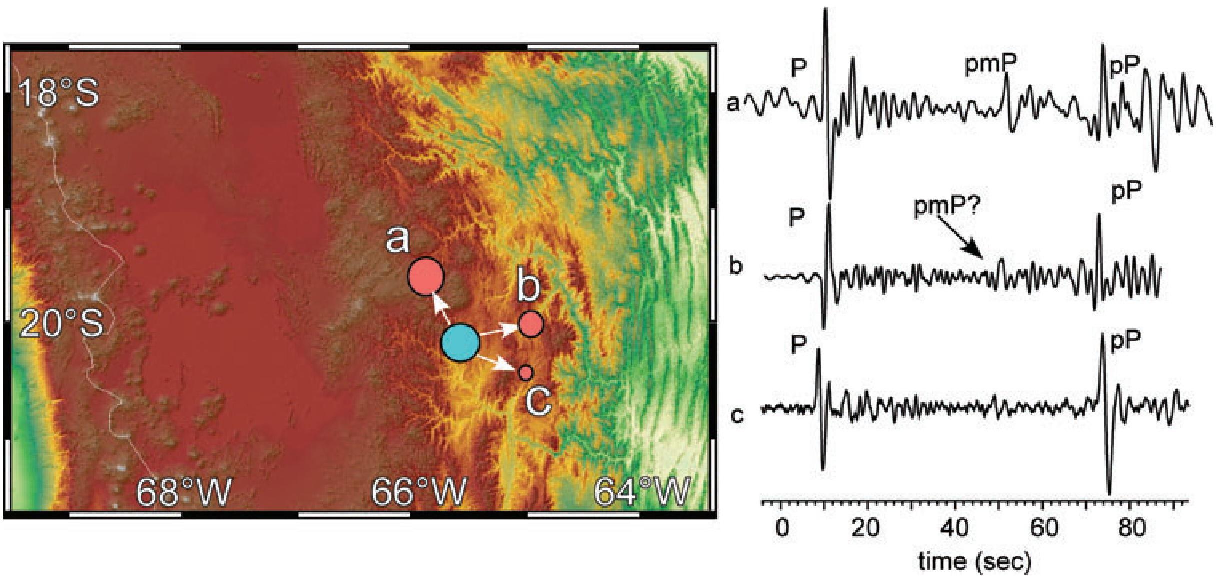

Figure 9. Recordings for an event on 2001 June 29 (blue circle) by stations a) PAS – Pasadena, California; (b) TAM – Tamanrasset, Algeria; (c) TSUM – Tsumub Namibia. The bouncepoints (pink circles) for the pmP phase sample distinctly different areas of the Moho (the Fresnel zone for pmP phase is approximately 1 km in each case). The variation in the strength of the Moho reflection either implies azimuthal anisotropy or lateral heterogeneity at the Moho. We suggest that the latter is the most likely explanation for the full range of variations in Moho reflectivity observed in this study.

reflection profiling and this technique. For this study, it was observed that the best underside Moho reflections were obtained using frequencies between 1 and 0.2 Hz. Thus there exists the possibility of a gradational velocity variation across the Moho preferentially reflecting these differing wavelengths.

Non-reflective (?) Moho

Although the method used in this study was successful in extracting estimates of crustal thickness at a number of locations in the Central Andes, a large fraction of the seismic records examined showed no discernible precursor phase. Compare, for example, the seismogram in Fig. 9(A) with those in Figs 9(B) and (C). Only the former is sufficiently clear to be reliably incorporated into this study. Since both P and pP are equally strong in all three cases, we suggest that the amplitude difference in pmP are most likely due to variability in the reflectivity of the Moho itself. The reflection character of the Moho has long been recognized as being highly variable from surface controlled source experiments (e.g. Oliver *et al.* 1983; Prussen 1989; Cook 2002). There have also been reports of 'weak' Moho reflections above subduction zones in western North America from passive source receiver function studies (Bostock *et al.* 2002; Zandt *et al.* 2004). In the receiver function studies, the lack of a Moho was interpreted to be caused, respectively by (i) serpentinization of the mantle forearc mantle due to dewatering of the subducted oceanic crust and (ii) an episode of crustal delamination. The lack of a clear Moho signal from the pmP technique may also be indicative of significant mineralogic variations in the lower crust or upper mantle—especially when contrasted with adjacent highly impulsive reflections from the same source. However, quantification of these variations is beyond the scope of this study.

A new crustal thickness compilation for the central Andes: tectonic implications

Summaries of the new crustal thickness estimates reported here along with those previously reported from receiver function technique and controlled-source seismic refraction studies (Yuan *et al.* 2000, 2002; ANCORP Working Group 2003; Fromm *et al.* 2004) are shown on a Digital Elevation Model (DEM) of the Andes in Fig. 10. The new results greatly expand the area thus far sampled for crustal thickness. For example, no previous measurements were

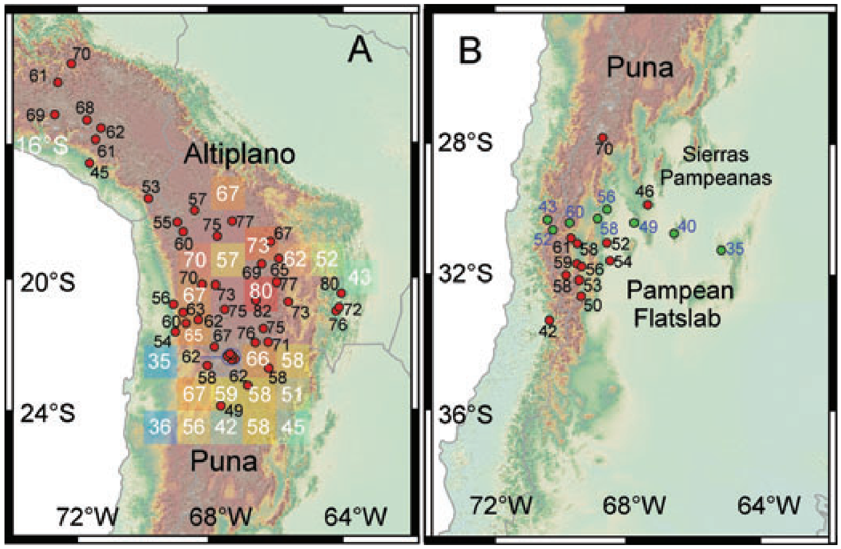

Figure 10. Crustal thickness, in kilometres, calculated from this study (red dots), compared with those from Yuan *et al.* (2002) (A) and Fromm *et al.* (2004), (green dots in B). The results from Yuan *et al.* (2002) are presented as averages within 1 deg^2 blocks.

available between 26°S and 30°S where a crustal thickness of 70 km was obtained near the Incapillo ignimbrite complex (28°S, 69°W). Likewise, a new crustal thickness estimate of 45 km is reported at the western margin of the Sierras Pampeanas. New crustal thickness estimates of 60–70 km in Fig. 7 are the first to be reported for the area east of the southern margin of the Peruvian flat-slab (18–14°S)

In the northern Puna and southern Altiplano region, our crustal thickness estimates of 49–66 km using the pmP method agree well with the crustal thicknesses of 42–67 km reported by Yuan *et al.* (2002). Under the main Altiplano plateau, Yuan *et al.* (2002) obtained a crustal thickness of 80 km compared with 82 km from the pmP method. Further south, strong pmP reflections from the Calingasta Valley in the Precordillera over the Chilean flat slab region yield Moho depths of 55–60 km that are in good agreement with nearby crustal thicknesses determined from receiver functions (Fromm et al 2004) in the CHARGE experiment.

One of the most interesting results of this study is the corroboration of the results of Yuan *et al.* (2002) that a marked topography exists on the Moho beneath the relatively flat Altiplano-Puna plateau, with depths varying from >80 km beneath the Altiplano to ∼45 km beneath the northern Puna. Such a variation seems inconsistent with a relatively uniform surface topography, which has an average elevation of 3.7 km (Isacks 1988). Given isostatic equilibrium, a relatively flat plateau suggests that the Moho should be at an average depth of 60–65 km throughout. The fact that this is not the case raises some interesting questions about the nature of the Moho. One is that in the Andes, and very likely other active-margin orogenic belts, isostasy is not always an accurate method for gauging crustal thickness. A combination of geodynamic processes allow for perturbation in the natural isostatic response making isostatic based estimates unreliable.

Alternatively, the various seismic techniques utilized in the Andes may be detecting a seismic discontinuity, but that is not the conventional crust–mantle boundary. Authors like Griffin and O'Reilly (1987) have highlighted the existence of a difference between the continental lithologic and seismic Moho. This difference can be greatly exacerbated by orogenic events where the creation of thickened crust causes eclogitization of the lower portion. Given the fast mantle-like seismic velocities of eclogites, an impedance contrast reflection for the Moho would occur at much shallower depths than the lithologic crust–mantle boundary. Thus varying levels of eclogitization in the crust can account for changes in Moho depth. The presence of eclogite in the Andean crustal root has been postulated in a large number of studies (e.g. Kay *et al.* 1994; James & Sacks 1999; Beck & Zandt 2002).

However, while eclogitization may explain limited discrepancies in crustal depth, it seems unlikely to explain large-scale differences observed between the central Altiplano and the northern Puna. Fig. 10 displays the prominent crustal maxima and minima: a high value of >80 km in the Altiplano and a low value of ∼45 km in the northern Puna. As previously stated, crustal seismic velocities in the crust of the Altiplano are anomalously felsic, thus a deep crust with mantle-like *P*-wave velocities in unlikely. It seems likely, therefore, that to explain these differences a significant portion of the Moho topography must have been physically modified by geodynamic processes. The two primary candidates are lithospheric delamination and lower crustal flow. The details of the possible effects of these processes follow.

Puna-Altiplano Plateau

Estimates of crustal shortening rates in the central Andes provide one straightforward explanation for the large crustal thickness in this region. Calculations of over 350 km of total shortening for the Altiplano (e.g. Kley & Monaldi 1998) allow for the presence of sufficient crustal material to compensate for large crustal thicknesses. However, although the Altiplano basin is relatively flat, elevations do range from ∼3.3 to ∼4.1 km. This difference, while significant, is not nearly enough to explain why large thickness deviations (82–57 km) appear to exist in the supporting crustal root between adjoining areas. One explanation of how the regions which appear to have a thick crust were formed requires that the thick mantle lithosphere under the Altiplano (Whitman *et al.* 1996) be tightly coupled to the overlying crust, resulting in a basal sinking force. This would result in the elevation of the Altiplano being artificially depressed despite the presence of a thick crust (e.g. Gerbault *et al.*

2005). Another effect could be that the thick lithospheric root of the Altiplano exerts a gravitational pull significant enough to open a planar channel within the lower crust. This would allow lower crustal material to flow through this channel thereby creating differences in the crustal thickness in adjacent regions and, given models like that of Beaumont *et al.* (2001), such flow would be unimpeded by topography on the Moho. Further, the fact that topography exists on the Moho precludes the possibility of lower crustal flow along the Moho discontinuity as ponding would occur under thinner sections of the crust, and thereby re-equilibrate the crustal thickness.

The area in the Puna centred on 25°S, 68°W is worth noting because here an unusually thin crust of ~45 km is observed. This region of thinner crust is peculiar because despite the apparent lack of a very thick crust, the region is at an elevation of ~4000 m, again violating the isostatic relation between elevation and thickness. Yuan *et al.* (2002) derived an average crustal thickness of 42 km in this region and also pointed out that this block may not be in isostatic equilibrium. They further observed that isostatic equilibrium could not be obtained even if the crust is purely felsic and the lithosphere is 50 km thick. Yuan *et al.* (2002) suggest a solution would be to mechanically couple this region with an adjacent region of thickened crust. The physical properties of a thin crust at high elevation combined with the presence of a thin mantle lid (Whitman *et al.* 1996) could be explained in another way. The dynamic nature of this area strengthens the argument for attributing these features to an episode of lower crustal delamination (e.g. Kay and Kay 1993; Kay *et al.* 1994; Schurr *et al.* 2006).

Southern Peru

Further north, the Moho results for Peru reveal a thick (60–70 km) crust at the southern margin of the Peruvian flat-slab indicating that a deep Moho is not limited to the highest elevations of the Altiplano plateau. There is also a divergence in this region between the crustal thicknesses expected from isostatic thickness based on elevation and that recorded by the precursor technique.

Chilean Flatslab

New crustal thicknesses for the area above the Chilean flatslab are in accord with thickened crust under the Andes extending eastward under the Pampean ranges of northern Argentina (Fig. 10). These data are in agreement with crustal thicknesses from the CHARGE experiment (Fromm *et al.* 2004; Calkins *et al.* 2006; Gilbert *et al.* 2006) as well as those of Regnier *et al.* (1994) based on earthquake data and Introcaso *et al.* (1992) from gravity data. Taken together, these data indicate a crustal thickness of ~60 km under the Main Cordillera, an eastward shallowing to between 60 and 50 km, and a rather abrupt thickness change to between 40 and 30 km to the east of 67°W.

Gilbert *et al.* (2006) and Calkins *et al.* (2006) used the results from the CHARGE experiment to argue for the presence of eclogite at the base of the crust under the Argentine Precordillera and Pampean ranges. The occurrence of eclogite and garnet granulite was inferred from weak *Ps*-wave conversions recorded at the Moho that were explained by the small impedance contrast that would result at a boundary between upper mantle peridotite and lower crustal eclogite. The presence of a mafic root in the deep crust of this region might be supported by the presence of garnet-granulite xenoliths in eastern Precordillera in Miocene volcanic rocks (Kay *et al.* 1996), although these xenoliths could alternatively be from the Precambrian crust beneath the Precordillera. General elevations in the Sierras Pampeanas near 1 km and locally over 2 km in a region of a deep crustal root can then be attributed to eclogite in the lower crust acting as a dense sinker mechanically depressing the region. A similar explanation has been used to explain the relatively low elevation (0.9–1.2 km) of the southern Ural mountains in central Russia. There, the explosive-source deep seismic reflection profile of the URSEIS experiment confirmed the presence of a thick (~55 km), mafic, seismically fast eclogitic root (Carbonell *et al.* 1996; Knapp *et al.* 1996; Steer *et al.* 1998).

Other lithospheric discontinuities

The general agreement between depth estimates for the Moho from the pmP technique and those from independent studies give us confidence that the precursors that have been identified generally correspond to reflections from the Moho. However, some precursory arrivals may correspond to discontinuities other than the Moho. For example, in the vicinity of Tarija in the Sub-Andean ranges, Yuan *et al.* (2002) report the Moho to be at ~45 km. Yet the most prominent depth-phase precursor we identify suggests a discontinuity at a depth of around 80 km (Fig. 4C). While such a large value might be expected beneath the Altiplano, such a thick crust seems inconsistent with both the geology and the relatively low average elevation of the Sub-Andean region. These arrivals could be reflections from the Hales discontinuity, an intermittently observed feature of the continental lithospheric mantle initially reported from a *P*-wave control-source study (Hales 1969). Subsequent studies have argued that the Hales discontinuity may be a zone of enhanced seismic anisotropy (Park & Levin 2001) or a relict mantle fabric that indicates the minimum thickness obtained by the stiff mantle lid during orogenic events (Levin & Park 2000). Although it is premature to generalize based on a single new observation, this underside reflection would suggest that the Hales discontinuity, if that is what is being detected, is a rather abrupt transition in bulk modulus or density and not merely a contrast in anisotropy.

In addition, other traces from the data set appear to depict reflections from intracrustal discontinuities at a depth of ~20 km [Fig. 4F(i)]. Such arrivals are analogous to those defining the 'bright spots' observed in Socorro, New Mexico (Brocher 1981) and Tibet (Brown *et al.* 1996) that have been interpreted to be magma bodies. The presence of intracrustal magma bodies in the Andes has previously been inferred from both seismic and petrologic data. In particular, seismic data were used in defining the Altiplano Puna Magma Body (Chmielowski *et al.* 1999), the existence of crustal melts on the ANCORP seismic line (ANCORP 2003) and beneath Tuzgle Volcano (Schurr *et al.* 2003, 2006). Coira & Kay (1993) used petrologic observations combined with geophysical results from the local seismic network of Cahill *et al.* (1992) to infer magma accumulation at ~20 km at an inferred crustal decollement beneath Tuzgle Volcano. The approximate 20 km depth observed by all of these techniques indicates that this is a preferred depth for upwardly ascending magmas to pool and initiate fractionation. The poor lateral resolution of the pmP method and the discrete nature of the crustal melt horizons place limits on the pmP technique as a tool in spatial mapping of these crustal discontinuities.

Fig. 4(F) also displays other sub-Moho reflections. While ambiguous, these reflections occur at depths comparable to those attributed to the Lithosphere Asthenosphere Boundary (LAB) (Heit *et al.* 2007)

CONCLUSIONS

Identification and mapping of teleseismic precursors to pP for intermediate to deep earthquakes beneath South America presents a rich new source of information on lithospheric structure beneath the Andes. Such precursors have allowed us to map crustal thickness variations over a large area without having a time-consuming field campaign. This technique is eminently suited to areas where field conditions are prohibitive and deep earthquakes plentiful. Where our results overlap previous findings using other techniques, they compare favourably. Most importantly, our results extend coverage over a large region where previous estimates are lacking. The success of the technique is due to the availability of significant numbers of seismic records from experiments conducted in the western US that are at the appropriate epicentral distance from South America.

The results obtained using the pmP technique combined with those from other studies show the crustal thicknesses under the Central Andes are more variable than expected from purely isostatic considerations. We infer that one or a combination of geodynamic processes such as lower crustal flow, lower crustal delamination as well as possibly mantle–crustal lithospheric coupling can explain the highly variable topography on the Moho. With this technique we have also found evidence of intracrustal (i.e. magma) and sub-Moho heterogeneity, indicating that the technique has uses well beyond mapping Moho topography.

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