Central Andean Crustal Structure from Receiver Function Analysis

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Abstract

The Central Andean Plateau (15°-27°S) is a high plateau in excess of 3 km elevation, associated with thickened crust along the western edge of the South America plate, in the convergent margin between the subducting Nazca plate and the Brazilian craton. We have calculated receiver functions using seismic data from a recent portable deployment of broadband seismometers in the Bolivian orocline (12°-21°S) region and combined them with waveforms from 38 other stations in the region to investigate crustal thickness and crust and mantle structure. Results from the receiver functions provide a more detailed map of crustal thickness than previously existed, and highlight mid-crustal features that match well with prior studies. The active volcanic arc and Altiplano have thick crust with Moho depths increasing from the central Altiplano (65 km) to the northern Altiplano (75 km). The Eastern Cordillera shows large along strike variations in crustal thickness. Along a densely sampled SW-NE profile through the Bolivian orocline there is a small region of thin crust beneath the high peaks of the Cordillera Real where the average elevations are near 4 km, and the Moho depth varies from 55 to 60 km, implying the crust is undercompensated by ~5 km. In comparison, a broader region of high elevations in the Eastern Cordillera to the southeast near ~20°S has a deeper Moho at ~65-70 km and appears close to isostatic equilibrium at the Moho. Assuming the modern-day pattern of high precipitation on the flanks of the Andean plateau existed since the late Miocene, we suggest that climate induced exhumation can explain some of the variations in present day crustal structure across the Bolivian orocline. We also suggest that south of the orocline at ~20° S, the thicker and isostatically compensated crust is due to the absence of erosional exhumation and the occurrence of lithospheric delamination.

Keywords: crustal thickness, Central Andes, receiver functions, isostasy

1. Introduction

The Central Andean Plateau (CAP) is the second highest plateau on Earth and is associated with a convergent margin where the oceanic Nazca plate subducts beneath the continental South American plate. The central Andes are an ideal place to study convergent margin processes and the structure of overthickened continental crust. The CAP, defined as the region over 3 km in elevation, consists of three distinct provinces from west to east; the active volcanic arc of the Western Cordillera, the internally drained Altiplano and the inactive fold and thrust belt of the Eastern Cordillera (McQuarrie and DeCelles, 2001). Further east is the active fold and thrust belt of the Subandean zone, and the modern foreland of the Beni basin. Previous

studies have shown that the CAP is characterized by thick continental crust and greater than 275 km of crustal shortening in the back-arc region (McQuarrie et al., 2005; McQuarrie et al., 2008; Oncken et al., 2006; Eichelberger, et al., 2015 and other references therein).

In 2010 an array of 50 broadband seismometers was deployed for two years in northwest Bolivia and southern Peru as part of the Central Andean Uplift and Geodynamics of High Topography (CAUGHT) project funded by the NSF-Continental Dynamics program. In this study we use data recorded by the CAUGHT seismic array, combined with additional data from permanent seismic stations and previous portable broadband deployments, to determine the crustal thickness and structure of the crust in the Bolivian orocline section of the central Andes in northern Bolivia and southern Peru. Our goal in this paper is to provide more detailed images of the crustal structure of the northern and central Andean Plateau that can be used as constraints to gain insights into possible mechanisms of plateau formation and uplift. We also identify two sections of undercompensated crust in the Eastern Cordillera that may be a result of their proximity to regions of higher precipitation.

2. Tectonic Setting

The CAP region of northwestern Bolivia and Southern Peru extends roughly 400 km from the volcanic arc to the Subandean zone. The Altiplano is an internally drained basin, with average elevations of ~3.8-4 km, between the high peaks of the Western and Eastern Cordilleras that reach up to 6 km. Volcanoes define the present-day extent of the magmatic arc in the Western Cordillera (Fig. 1). The Eastern Cordillera along this section of the Andes is an inactive fold and thrust belt that contains Triassic and Miocene plutons (Isacks, 1988; Kley, 1999; McQuarrie and DeCelles, 2001; Gilles et al., 2006). The Interandean zone, located between the

Eastern Cordillera and Subandean zone, is a region that was uplifted by imbricated basement material between 10 and 5 Ma, as shortening propagated east into the active, thin-skinned fold and thrust belt in the Subandean zone (Kley, 1996; McQuarrie et al., 2005).

Of interest in this study is the northern portion of the CAP, from roughly 12°-20° S and 63°-72° W. This region lies in one of the widest sections of the CAP at the Arica Bend, and has undergone significant upper crustal shortening in the backarc. Numerous studies have found higher amounts of shortening, up to 350 km, in the orocline region than in either the Puna region to the south, where backarc shortening is closer to ~200 km, or in the Andean region farther north in Peru where shortening is ~175 km (Kay and Kay, 1993; Allmendinger and Gubbels, 1996; Kley and Monaldi, 1998; McQuarrie et al., 2008; DeCelles, et al., 2011; Gotberg et al., 2010). McQuarrie (2008) partitions the overall upper crustal shortening between 14° and 18°S into the tectonic provinces along the orocline axis in roughly the following amounts: 39 km in the Altiplano, 171 km in the Eastern Cordillera and Interandean zones, and 66 km in the Subandean zone. Upper crustal shortening has occurred in the region, at an essentially constant rate, from roughly 45-0 Ma (McQuarrie et al., 2008; Eichelberger and McQuarrie, 2015). While exhumation rates are fairly consistent with proposed shortening models from 45-0 Ma, there is proposed surface uplift of ~1.5-3 km since 10 Ma with little associated upper-crustal shortening in the Eastern Cordillera and continued exhumation of the Eastern Cordillera from 15-0 Ma (Gillis et al., 2006; Garzione et al., 2006; McQuarrie et al., 2008; Leier, et al., 2013). Recent climate studies have instead suggested that uplift of the region in the 15-0 Ma time frame was less than one kilometer (Poulsen et al., 2010; Insel et al., 2012).

These differences in the timing of upper-crustal shortening and exhumation, as well as ~12 Ma low relief surfaces identified throughout the Eastern Cordillera, have been used to argue

that two separate phases of uplift occurred, one associated with upper-crustal shortening that caused initial uplift of the Eastern Cordillera, and another that led to a rapid rise of the Eastern Cordillera and Altiplano in the last 10 Ma (Isacks, 1988; Gubbels et al., 1993; Kennan et al., 1997; Husson and Sempere, 2003; Gillis et al., 2006; Hoke and Garzione, 2008; Whipple and Gasparini, 2014). To account for the rise of the Eastern Cordillera without upper crustal shortening since ~10 Ma (e.g. Isacks, 1988), mechanisms have been proposed that include underplating of fore-arc material, under-thrusting of the Bolivian craton, removal of material from the lower crust and upper mantle and lower crustal flow (Baby et al., 1997; Giese, et al., 1999; Allmendinger, 1996, Husson and Sempere, 2003; Oncken, et al., 2006, Garzione, et al. 2006, Molnar and Garzione, 2007, Lamb, 2011). Foundering, or lithospheric removal, has also been postulated as a means of creating cycles of orogenic uplift in addition to regional high topography such as that seen in the Western U.S. or Andean Plateau. This process involves eclogitic roots that form under mountain systems, with their removal being a potential driving force in the creation of high mountain systems as an attempt to achieve isostatic equilibrium (Yuan et al., 2000; Yuan et al., 2002; Zandt et al., 2004; Shurr et al., 2006; Molnar and Garzione, 2007; Garzione et al., 2008; Kay and Kay, 1993; DeCelles et al., 2009). Other studies have argued against the need for foundering to cause large-scale regional uplift, and have instead supported crustal thickening models that create high topography with minor or no contributions from lithospheric removal (Lamb, 2011; Barnes et al., 2012; Hindel et al., 2005; Eichelberger et al., 2015 and other references therein).

2.1 Previous Geophysical Studies

Previous geophysical work in the central Andes has shown that there are significant changes in crustal character across the width of the Andes, with crustal thicknesses ranging up to

70 km or more underneath the Western and Eastern Cordilleras, up to ~60 km underneath the Altiplano and as little as 35 km further east into the Bolivian craton (Dorbath et al., 1991; Dorbath et al., 1993; Beck and Zandt, 2002; Yuan et al., 2002; McGlashan et al., 2008). Most of these studies overlap slightly with or are south of our study area. There are large magma bodies documented to the south in the Altiplano-Puna region (Chmielowski et al., 1999; Zandt et al., 2003, Ward et al., 2013), and evidence for low-angle thrust faults with little to no surface representation between ~21-24°S (Wigger, et al., 1994; Schmitz and Kley, 1997). Further, the thick crust underneath the central Andes near 19-20°S, including the 5-6 km elevations of the Western and Eastern Cordillera peaks, appear to be in isostatic equilibrium (Beck and Zandt, 2002; Yuan et al., 2002). High seismic velocities under the Western Cordillera match well with those expected of mafic material thought to be associated with active arcs, and are not present in either the Eastern Cordillera or underneath the Altiplano (Myers et al., 1998; Graeber and Asch, 1999).

A recent regional study further to the north of our study area, in southern Peru (~13-17°S), found crust up to 75 km thick, and a positive impedance contrast at 40 km depth that the authors interpreted as possible under-thrusting of the Brazilian Shield underneath the Eastern Altiplano (Phillips et al., 2012; Phillips et al., 2014). These studies cover an area between the flat-slab subduction segment in central Peru, and the normal dipping (30°) slab segment in southern Peru; ~13-16°S and ~75-70°W. In these studies, the Moho underneath the Altiplano has also been shown to have minimal topography, and maintains a thickness of 70 km from north to south.

In our study region, Dorbath et al. (1993) estimated lithospheric structure based on local tomography along a line of stations across the Andes in Bolivia that corresponds to our densest

line of stations (Figure 2a, B-B'). They estimated the crustal thickness to be 60 km under the Altiplano and 50 km under the Eastern Cordillera based on changes in P-wave seismic velocity. However, there are few other locally derived determinations of crustal thickness in the CAP in northern Bolivia or southern Peru where there are high elevations and large amounts of crustal shortening.

A number of recent studies have developed continental-scale crustal thickness models for South America using seismic and gravity techniques (Feng et al., 2007; Lloyd et al., 2010; van der Meijde et al., 2013; Assumpção et al., 2013; Chulick et al., 2013) but there is limited data on the depth to Moho in the central and northern Altiplano. These studies show a smooth Moho with thick crust (~60-70 km) beneath the high elevations of the Andes and lower elevations (30-40 km) beneath the low elevations east of the Subandean zone. The results from our study will improve continental scale models in the future.

3. Data

The Central Andean Uplift and Geodynamics of High Topography (CAUGHT) seismic deployment was part of a larger collaborative National Science Foundation Continental Dynamics funded project, and consisted of 50 STS-2 broadband seismometers deployed in a temporary network for two years (2010-2012) in northwest Bolivia and southern Peru (Fig. 2a). We have supplemented that data with additional records from the BANJO-SEDA deployment; the permanent GSN station near La Paz, Bolivia (LPAZ); an array installed by the Observatorio San Calixto in Bolivia; stations from the Integrated Plate Boundary Observatory Chile (IPOC); an international project by GFZ Potsdam; and a new permanent station installed in northern Chile by the Chile National Network (Fig. 2a; Table 1). We have used both teleseismic and local

earthquakes recorded by the CAUGHT temporary network and other nearby stations in this study.

We used teleseismic waveforms from events ($M_w > 5.5$) in the direct-P and PP distance ranges of $\sim 30\text{-}95^\circ$ and $\sim 100\text{-}150^\circ$, respectively. The distribution and location of the events used for calculating the receiver functions in this study are shown in Figure 2b. We identified 343 earthquakes with both a strong first arrival and high signal-to-noise ratios (greater than 10:1) for use in this study (Fig. 2b). For the permanent station LPAZ we used events over a 10-year period. Excess waveforms from the LPAZ station at dominant back-azimuths were removed to ensure even back-azimuthal distribution. Many of the stations used in this study had similar back-azimuthal coverage, despite their differing deployment times, and some stations had coverage on par with permanently deployed stations as shown in binned back-azimuth plots (Fig. 3). A figure showing examples of the raw data, before binning, is available in Supplemental Figure 1. While stations from the network deployed by the Observatorio San Calixto were used in this study, we had a limited number of receiver functions (10 or less for each station) because the stations had been recording for 6 months or less at the time of data collection.

In order to constrain the average V_p/V_s ratio for the crust we use Wadati diagrams calculated using 9 local earthquakes with depths < 50 km. For our Wadati diagram we have used 200 local waveforms that were chosen to be less than 300 km distance from the line of densest station spacing (B-B', Fig. 2a) in the orocline region of the CAUGHT array to ensure that the direct-P was the first arrival. Of the nine events used, about half were located off the west end of the B-B' CAUGHT line and the rest were distributed along its length. The local waveforms had distinct P and S-wave first arrivals so that differential travel times could be determined.

4. Methods

4.1 Receiver function and CCP stacking method

Teleseismic waveforms were used to calculate receiver functions, which isolate converted phases from velocity discontinuities beneath a seismic station (Langston, 1979). Radial and tangential receiver functions are calculated by deconvolving the vertical component from the radial and tangential components of our teleseismic waveforms. P-to-S conversions at velocity discontinuities result in arrivals following the direct P-wave arrivals, which are visible on the radial component of the receiver function (Ammon et al., 1990; Ammon, 1991; Cassidy, 1992). Tangential receiver functions can provide additional information because in the absence of anisotropy or dipping structures, they should show no energy.

In order to calculate our receiver functions, we first filtered the waveforms using a Butterworth bandpass filter with corners at 0.1 and 5 Hz to isolate the initial P-wave arrival and eliminate any long-period noise. After rotation of the horizontal components into the radial-transverse reference frame we applied the iterative time-domain deconvolution technique of Liggoria and Ammon (1999) using a Gaussian pulse of 2.5 (corner frequency ~ 1 Hz). We allowed up to 400 iterations in the deconvolution. Any receiver functions showing initial negative troughs, very high amplitudes after the first arrival, or significant resonance and ringing in their coda were discarded. Final quality control consisted of elimination of any receiver functions with more energy (higher amplitude) on the transverse receiver function than the radial receiver function at the time of the initial P-wave arrival. In our final data set we have ~ 2000 receiver functions from 343 teleseismic earthquakes, corresponding to ~ 20 receiver functions per station (Fig. 2b).

Teleseismic receiver functions used in this study were stacked using the common conversion point (CCP) method, which stacks receiver functions in subsurface bins based on the location of the P-to-S conversion to increase the signal to noise ratio (Dueker and Sheehan, 1997; Sheehan, et al., 2000). This stacking method allowed us to share information between adjacent bins; a portion of the signal amplitude of a source bin is shared with nearby bins in an effort to smooth the resulting model and close small gaps in receiver function coverage. A phase-weighting filter was applied to the stacks to reduce signal-generated noise (Frassetto et al., 2010), and any bins sampled by fewer than five ray paths were removed. An example of the effects of different phase-weighting strengths is shown in supplemental Fig. S3. All the receiver functions are corrected for elevation and the stacks are plotted as depth below sea level.

During the stacking process all of the receiver functions were migrated from the time domain into depth using a 3D S-wave velocity model determined from a combination of the ambient noise tomography study by Ward et al., (2013) above 60 km, and the IASP91 global model below 60 km. The ambient noise model used periods for 8-40 seconds (Ward et al., 2013). We used a constant V_p/V_s ratio of 1.75 in the upper 60 km, and a ratio of 1.80 below 60 km to calculate a P-wave velocity model from the 3D S-wave velocity model. We used local earthquakes and the Wadati diagram method, to determine the average V_p/V_s ratio of 1.75 in the mid to upper crust (Whitcomb et al., 1917; Fig. 4). This also matches well with previous estimates of V_p/V_s ratios calculated for the northern CAP (Cunningham and Roecker 1986; Dorbath et. al. 1996; Chulick et. al. 2012). While it is common the use the HK method (Zhu and Kanamori, 2000) to estimate the V_p/V_s for individual stations, we found the HK results for our stations to be generally unreliable (a sample HK plot is shown in supplemental Fig. S4). Receiver functions were back-projected through this 3D volume starting with a 1D average of

the entire volume to produce an initial ray path. Refinement of that ray path was then done using the velocities sampled along the initially calculated ray path through the 3D velocity volume (Frassetto and Thybo, 2013).

Receiver functions were stacked along dozens of different cross-section profiles, of which we present the five cross-sections, through the regions of the volume with the highest density of station coverage (Fig. 5, Fig. 6a-d). We used rectangular bins that can vary both along-profile and perpendicular to the profile line in order to maximize the coverage of each cross-section. Along the cross-section where we have the densest station spacing we defined the bins with a 15 km width along-profile and 30 km width perpendicular to the profile. Our dense profile has two trends; B-B' SW of the bend has a trend of 32° and NE of the bend has a trend of 58° (Fig. 5). For the other cross-section profiles we used 50 km bins to better match the station spacing: a trend of 60° was used for the northernmost profile (Fig. 6a), a trend of 90° was used to re-process the original BANJO-SEDA line (Fig. 6b), trends of 132° and 172° were used to fit the Western Cordillera (Fig. 6c), and trends of 130° and 170° were used along the high topography of the Eastern Cordillera profile (Fig. 6d). Our bins average about a dozen piercing points each at 50 km depth, and we eliminate bins with less than 5 rays contributing to the stack (Supplementary Figure S2).

We used the CCP stack cross-sections (~ 25 orthogonal and ~ 10 parallel sections) with 50 km bins through the entire volume to calculate the depth of the Moho, not just the five cross-sections shown in Figures 5 and 6, in conjunction with single station stacks to identify the conversion from the continental Moho, which is usually the strongest velocity contrast in the receiver functions. We identified the P-to-S conversion from the Moho (relative to sea level) and several other discontinuities in the crust and upper mantle. We made a map of the depth to the

Moho and used the same volume of stacked receiver functions and nearest neighbor interpolation between adjacent bins to contour our map (Fig. 7). Adjacent areas outside of the station coverage were constrained based on the regional scale Moho map by Tassara et al., (2006). The Ps piercing points and the bin sharing geometry at 50 km depth is shown in supplemental Figure S2 and provide a qualitative measure of the data coverage for the Moho map. While we note that some bins are based on interpolation between stations, the coverage is significantly better than previously available.

The uncertainty in the depth of a P-to-S conversion is a result of the trade-off between depth to a conversion layer and the velocity structure along the ray. Small changes in the V_p/V_s ratio have been shown to create large differences in the calculated depth of the converted arrival (Zhu and Kanamori, 2000), so uncertainty is reduced significantly by using a well-constrained V_p/V_s ratio. Migrated receiver functions are less sensitive to changes in the absolute P-wave velocity, with changes of V_p model used, from a simple two-layer model to one constrained using ambient-noise tomography, influencing arrival depth by less than ± 0.5 km. Using a Wadati diagram, we have identified the best-fit average mid to upper crustal V_p/V_s ratio (1.75) in the region to within 0.01, which reflects an uncertainty of about ± 1 km at 60 km depth (Fig. 4). However, larger errors are possible as the V_p/V_s ratio could change locally, and we have little constraint on the V_p/V_s ratio below ~ 40 km. In addition, dipping structures, or the presence of regions of high anisotropy, could also contribute significantly to the uncertainty of our arrivals.

5. Results

5.1 Moho depths

Our results are shown in Figures 5 and 6 as CCP cross-sections and in Figure 7 as a Moho depth map relative to sea level. Along our densest instrumented line, CAUGHT B-B' (Fig. 2a, 5), our CCP cross-section shows a prominent continental Moho increasing in depth from ~40 km on the eastern edge of the active fold and thrust belt to ~60 km on the eastern edge of the high plateau. The Moho depth reaches nearly 70 km in the center of the Altiplano and appears to remain deep under the western Cordillera but its identification there is complicated by complex crustal structure in the transition zone between the arc and western edge of the plateau. Beneath this transition zone the receiver function stacks show a complex layered structure suggestive of multiple low velocity layers in the crust and likely includes multiples from the prominent low velocity zone at ~15 km depth (TLVZ). The continental Moho location under the forearc is not clear in the CCP stacks. There is a prominent positive arrival at ~70 km depth but it is too deep to be the continental Moho and too flat to be the oceanic Moho of the subducting slab. We suggest the continental Moho here is weak due to partial serpentinization of the mantle above the slab as observed in other subduction zones (Wada, et al., 2008 and references therein).

The cross-section A-A', across southern Peru just south of the Peruvian flat slab, has limited coverage in the forearc and the narrower fold and thrust belt at this latitude (Fig. 6a). The Moho depth is 60 – 65 km under the plateau, shallowing to ~50 km under the Subandes. There are complexities and low-velocity zones associated with the wide volcanic arc in this region. Farther south the C-C' line (Fig. 6b) follows the BANJO-SEDA line interpreted in Yuan et al. (2000,2002) and Beck and Zandt (2002). The CCP stack cross-section along this line utilizes a few additional stations and uses PP phases, in addition to the P phases that were used in Beck and Zandt, (2002). Our results are very similar to those found along similar cross-sections, previously. This cross-section shows similar crustal thicknesses to line B-B' in the Western

Cordillera and Altiplano, and a 40 km thick crust in the Subandean zone, but the Moho under the Eastern Cordillera is significantly deeper, ~65-70 km.

Along the strike of the Western Cordillera, line D-D' (Fig. 6c), the Moho arrival is indistinct along most of the profile. There is a high amplitude arrival at 70-75 km depth where the cross-section intersects line B-B', but as we previously stated it is difficult to unambiguously identify the Moho among the multiple layers. However, north of 18°S there is a faint continuous arrival at 70-75 km that may mark a weak or gradational Moho under the active arc. South of 18°S along the western Cordillera, the CCP stacks show layering that is unlikely to be due to multiples, suggesting complex Moho structure (Fig. 6d). While an accurate Moho depth is difficult to pinpoint, it is unlikely that either the arrival at 90 km or the shallowest positive conversion at 40 km represent Moho depths.

The CCP cross-section along the strike of the Eastern Cordillera (line E-E', Fig. 6d) shows a continuous high-amplitude Moho at depths of 60-70 km at the NW and SE ends and a broad shallowing to depths of 55-60 km in the center. At the southern end, in the location of the BANJO-SEDA network there is at least locally a deep Moho beneath the Eastern Cordillera (~65-70 km). A deeper Moho (~70 km) is also observed to the northwest in the Eastern Cordillera as the Altiplano pinches out in southern Peru. In between the Moho has a broad upwarp defining locally thinner crust with the Moho reaching a minimum depth of 55 km just north of La Paz (Fig. 6d).

Figure 7 shows a smoothed contour map of depth to the Moho beneath our study region using receiver function results under the seismic station array. On the edges, and outside the array, we constrained the depth of Moho from Tassara et al., (2006) who used primarily gravity data in these regions. The map shows Moho depths < 40 km in the forearc and foreland. The 60

km depth contour encompasses most of the high elevations of the plateau. The deepest Moho and thickest crust occurs in southern Peru southwest of Lake Titicaca (Figure 7, “D”). The 45 km, 50 km, and 55 km contours generally parallel the eastern side of the CAP but with two incursions into the area of highest elevations of the Eastern Cordillera. One is a long wavelength (~300 km) “bowing” of the contours centered at ~17°S (Figure 7, “E”) and the other is a shorter wavelength (~150 km) “bite” into the Altiplano just north of La Paz (Figure 7, “D”), outlining the locally thin crust we highlighted in our earlier description of cross section B-B’.

5.2 Intra-crustal and uppermost mantle discontinuities

Along the CAUGHT dense line there are two major upper to mid-crustal conversions that stand out on the CCP cross-sections: a negative polarity arrival underneath the Western Cordillera at a depth between ~15 (Fig. 5, 6, TLVZ= top of low velocity zone) and a positive polarity arrival underneath the Eastern Cordillera and Subandean zone (Fig. 8; “A” and “B”). The negative arrival underneath the Western Cordillera, at 15 km or less, agrees well with the top of a low velocity zone identified in the ambient noise results of Ward et al., (2013). This negative arrival is also present along the C-C’ line (Fig. 6b) and is present along most of the along-strike D-D’ section (Fig. 6c).

Underneath the Subandean zone along the B-B’ cross-section there is a high amplitude positive conversion at 15 km that shallows to the east (Figure 8, “A”). There is a similar arrival at shallower depths underlying the Eastern Cordillera (Figure 8, “B”). The same feature appears to be more continuous further south, along the C-C’ profile (Figure 6b, “Detachment”). Figure 8 shows an enlarged version of this part of the CCP cross-section with an overlay of the balanced structural cross-section of the fold and thrust belt from McQuarrie et al. (2008). We note the

correlation of the positive conversion with the decollement fault at base of the highly faulted and folded layers and suggest the arrival marks the top of the “basement” rocks below the basal fault. Further supporting this interpretation is that the locations of vertical offsets of the arrival closely coincide with the geologic offsets in the structural cross-section.

Locally, there are some interesting lower crustal and upper mantle features in our cross-sections. We previously mentioned the complex lower crustal layering in the transition between the Western Cordillera and Altiplano along cross section B-B’ (Fig. 5). Along strike farther south cross-section C-C’ (Fig. 6b) exhibits the upper crustal low-velocity zone arrival but not the lower crustal and upper mantle ones. We observe some positive amplitude arrivals on the CCP cross-section B-B’ at depths of 60 to 200 km in the vicinity of the slab location based on Slab1.0 contours of Hayes, et al., (2012) but we do not image a continuously dipping slab. Another positive conversion arrival is visible underneath the Interandean and Subandean zones and perhaps extending under the Eastern Cordillera at 80-90 km depth along lines B-B’ and C-C’ (Fig. 5, 6b). We are uncertain whether these seismic discontinuities are older inherited features or active structures related to the underthrusting lithosphere.

6. Discussion

We have developed a higher resolution image of the depth to Moho beneath the CAP region of South America than was previously available (Fig. 7) and calculated a crustal thickness residual map based on a simple Airy isostasy model (Fig. 9b and 9c). Overall, our results agree well with the continental-scale models that were mentioned previously, while improving the resolution in many areas (Assumpção et al., 2013; Chulick et al., 2013 and the references therein). We have identified several mid crustal discontinuities across the region that correlate

well with known geology. The significance of the crustal thickness residuals and crustal discontinuities are discussed below starting with the Western Cordillera.

6.1 Western Cordillera Crust

The orocline region of the Western Cordillera contains a large number of active stratovolcanoes in southern Peru and extending southward on the border between Bolivia and Chile. Our study cannot image the specific plumbing of any one volcanic center or the associated shallow magma bodies, but we can image the overall regional crustal structure beneath the Western Cordillera. Our receiver function stacks show numerous conversions in the crust that produce multiples, making the receiver functions more difficult to interpret. There is a strong negative arrival indicating the top of a low velocity zone at ~5-20 km under much of the Western Cordillera-Altiplano boundary region (Fig. 5, 6c). Along cross-section A-A' the top of the low-velocity layer dips from ~5 km to 15-20 km toward the east (Fig. 6a). The ambient noise tomography shows S-wave velocities of ~3.0-3.25 km/s at 30 km depth under the active arc (Ward et al., 2013). We observe this decrease in velocity labeled TLVZ, top of low-velocity zone, on cross-section D-D' along the strike of the volcanic arc (Fig. 6c).

We observe a sharp increase in velocity in the receiver function images at ~40-45 km depth along the dense line B-B' (Fig. 5) perpendicular to the strike of the volcanic arc (where we have the smallest bin size) and on C-C', but we do not see this as a continuous discontinuity (limited extent near 71°W) on the along strike cross-section (D-D') with the larger bin spacing (Fig. 6c). Where we see this increase in velocity it may be the bottom of a mid-crustal low velocity zone. This suggests that at least in this region there is a thick region of the crust (~15-40 km) with very low S-wave seismic velocities. This low S-wave velocity region in the mid-crust

is consistent with high temperatures and some small amount of partial melt assuming an intermediate to felsic composition (Ward et al., 2013).

We also observe a strong increase in velocity at ~40-45 km depth (Fig. 5) with a decrease in velocity at ~60 km right above the Moho along B-B' suggesting a low velocity region in the lower crust. This might be related to the melting, assimilation, storage and homogenization (MASH) zone proposed by Hildreth and Moorbath (1988) in central Chile. We do not see this lower crust structure everywhere along the arc but it is observed near the intersection of the Peru, Chile, and Bolivia border, beneath the active arc (Fig. 5, 6a).

6.2 Detachment structure across the retroarc fold and thrust belt

Comparing the midcrustal discontinuities from our receiver functions with a balanced structural cross-section across the same region (McQuarrie, et al., 2008) highlights important similarities between the interpreted subsurface geology and our geophysical results along line B-B' (Fig. 8). In the Subandean zone there is very good agreement between the basal detachment in the balanced cross-section of McQuarrie et al. (2008) and a discontinuity (increase in velocity with depth) in our CCP stacks that dips from a depth of ~5 km at the eastern edge of the Subandean zone to ~15 km at the boundary with the Interandean zone (Fig. 8, "A"). The discontinuity disappears beneath the Interandean zone presumably because as the detachment goes into the basement there is not an impedance contrast across it. We see a shallow, near horizontal, discontinuity indicating an increase in velocity beneath the Eastern Cordillera at ~5 km that corresponds to the basal detachment for the westward verging backthrust zone (Fig. 8, "B"). Beneath the Altiplano we observe a weak amplitude discontinuity at ~20 km that is consistent with a step down in the detachment on a basement ramp in the cross-section of McQuarrie et al. (2008), where the detachment is at a depth of ~20 km beneath the Corque

Syncline (Fig. 8, “C”). The Tertiary basin fill is ~12 km thick on the Altiplano with ~8 km of Paleozoic section, indicating that basement is at ~20 km beneath this portion of the Altiplano (Horton et al., 2002; Lamb and Hoke, 1997), consistent with a positive arrival observed on our cross-section (Fig. 8, “C”). The detachment at the basement interface would produce a discontinuity with an increase in velocity with depth, consistent with our observations and those of McQuarrie, et al. (2008).

6.3 Variations in crustal thickness

To better investigate whether the crust is isostatically compensated at the Moho for the region as a whole, we have used a topographic model to derive the predicted Airy isostatic crustal thickness to compare to our observed crustal thickness from receiver functions. Figure 9b uses topography (GTOPO, 30 arc seconds, Lindquist, et al., 2004) low-pass filtered at 10 km x 10 km (100 km²) to better match the expected spatial wavelength of isostatic compensation (Molnar and England, 1990). We then use the following form of the isostasy equation to calculate the Airy isostatic compensation thickness for each crustal column (Molnar and England, 1990):

$$T = C + H + H * \left[\frac{\rho_c}{\rho_m - \rho_c} \right]$$

The above equation defines the crustal thickness predicted by Airy compensation (T) to be the sum of the reference crustal thickness of crust (C) plus the elevation above sea level of the column (H) and the thickness of root material needed to reach compensation based on the assumed densities of the crust and mantle, ρ_c and ρ_m . Performing this calculation for all the columns in the topographic model yields a map of predicted crustal thickness based on an initial reference crustal thickness of 38 km, $\rho_c = 2.8 \text{ g/cm}^3$ and $\rho_m = 3.3 \text{ g/cm}^3$ (Fig. 9b). We tested reference crustal thickness values ranging from 35 to 40 km that all showed similar features with

differences mainly in their amplitudes. We compared these predictions to the observed crustal thickness we obtained by correcting our depth to Moho results (Fig. 7) to be crustal thickness including topography (Fig. 9a). We subtracted the calculated crustal thickness model from our observed crustal thickness to create a crustal thickness residual map (Fig. 9c). Our image shows areas of undercompensated crust, where there is less crust than that needed to support elevations, in blue, and areas of overcompensated crust in red, based on the simple assumptions mentioned above. We do not include results in the forearc, which were not properly represented by the constant density model ($\rho_c = 2.8 \text{ g/cm}^3$ and $\rho_m = 3.3 \text{ g/cm}^3$) that we applied throughout the region. The three largest anomalies in the residual map include an excess of $\sim 15 \text{ km}$ of crust in the Western Cordillera and northern Altiplano of southern Peru (anomaly F in Fig. 9c) and two zones of deficit of $\sim 10 \text{ km}$ of crust in parts of the Eastern Cordillera (anomalies D1, D2, and E). We note that the smaller anomalies D1 and D2 are better sampled by our seismic stations than anomaly E. For purposes of this paper, we will also refer to the spatially related D1 and D2 anomalies as a single anomaly, D.

Our map of depth to Moho, Figure 7, indicates that much of the Western Cordillera and Altiplano regions of the CAP have Moho depths of 60-70 km (relative to sea level) similar in depth to what previous authors have found farther south (Wigger et al., 1994; Swenson et al., 2000; Yuan et al., 2000; Yuan et al., 2002; Beck et al., 1996; Zandt et al., 1996; Beck and Zandt, 2002; Wölbern et al., 2009). The crustal thickness residual map (Fig. 9c) shows the Altiplano crust in southern Peru is thicker by $\sim 5\text{-}10 \text{ km}$ (anomaly F) than the central Altiplano crust in northwest Bolivia. The thicker crust is consistent with the results of Phillips et al., (2012) and Phillips and Clayton (2014) just north of our study area. The $\sim 5\text{-}10 \text{ km}$ difference would suggest at least a 1 km change in elevation between the central and northern Altiplano, if the crust was

being entirely compensated at the Moho. The surface elevation increases from ~3.8 km in the central Altiplano to ~4.3 km in the northern Altiplano but that is accounted for in our crustal thickness residual calculation. There is no indication of a larger amount of crustal shortening in southern Peru. However, we cannot rule out differences in density between the crust and uppermost mantle that might contribute to the difference. Preliminary results from a surface wave tomography study of the same region (Ward et al., 2015) show an upper mantle high velocity anomaly beneath the area of overcompensated crust suggesting an upper mantle density anomaly may be partly responsible for holding down the elevations in this area.

Locally, parts of the Eastern Cordillera crust (east of La Paz) are thinner than predicted by Airy isostasy based on smoothed topography. There are two regions of relatively high smoothed elevations with crust that are ~10 km thinner than what is predicted from our simple isostatic calculation (Fig. 9c). The crust in these areas is also ~10-15 km thinner than what we observe along-strike in the Eastern Cordillera in central Bolivia, near ~19-20°S. Figure 10 shows a comparison of the depth to Moho and topographic profile along two transects across the Andes, one north of the Bolivian orocline (Fig. 5 B-B') and the other south of the orocline (Fig. 6 C-C'). Although the Eastern Cordillera is very different in width in the two profiles the average maximum elevation is similar yet the crustal thickness differs by 10-15 km. Also, the topographic profiles are very different in the Interandean and Subandean zones but there is no evidence for different amounts of shortening in these two regions (McQuarrie et al., 2005; McQuarrie et al., 2008). The topographic difference was recognized by Masek et al. (1994) who attributed the difference to differences in climate-controlled erosion and by Horton (1999) who suggested that increased erosion rates in the north influenced the kinematics of thrust belt

development. An additional difference is that the southern profile passes through the Los Frailes Volcanic Field, a Late Cenozoic ignimbrite center (Baker, 1981).

The two regions of thinned crust in Fig. 9c have been designated anomalies D (1 and 2) and E. Both regions of thinner crust are spatially small and isolated; anomaly D, beneath the northern Eastern Cordillera is ~150 km by 50 km, while the thin crust in anomaly E extends ~300 km along-the strike of the Eastern Cordillera and is roughly ~100 km wide. The western edges of the thin crust in the Eastern Cordillera correlate well with Triassic and Miocene plutonic bodies, but the mapped surface area of those plutons is an order of magnitude less than the area over which thinner crust is observed.

6.4 Tectonic implications

There are a number of possible mechanisms for modifying the crustal thickness, including climate controlled erosion, lower crustal flow, and lithospheric delamination. In the following paragraphs we briefly discuss how each of these mechanisms may fit with our observations of isostatic compensation.

Enhanced erosion could explain crustal thinning if there is no shortening in the same region. This would lead to a thinning of the crust and slightly lower average elevations (Molnar and England, 1990). Furthermore, if the erosion occurs non-uniformly (i.e. due to the incision of deep valleys and the exhumation of steep peaks), the resulting high relief landscape, after isostatic adjustment, will have a slightly lower average elevation but the Moho will move up as much as the tallest peaks are exhumed. This process could explain the existence of the high relief Cordillera Real with peaks as high as 6 km, in an area of relatively thin crust (Fig. 5).

Although our crustal thickness residual map suggests that areas with thin crust are undercompensated, this result is based on a low-pass filtering of the topography of the entire

study region. To get a more accurate sense of the isostasy relationships in the Eastern Cordillera, we selected an area bounded by a rectangle that encloses the Eastern Cordillera between $\sim 15^{\circ}\text{S}$ and 18°S (yellow box in Fig. 9c). The mean elevation in this area is 3.1 kilometers, which would indicate a predicted isostatic Moho depth of 55 km. From Figure 1 it is clear that there is a steep topographic gradient in this area, so we divided the rectangle lengthwise into a southwestern half and a northeastern half and repeated the isostasy calculation for each. The western half has a mean elevation of 3.9 km with a predicted isostatic Moho depth of 60 km and the eastern half has a mean elevation of 2.2 km and a predicted Moho depth of 50 km. Comparing these predicted depths with the Moho depth profile along our dense line in Fig. 10 we observe the EC appears to be, on average, near isostatic equilibrium with an average Moho depth of 55 km. But in detail, the western half of the EC appears slightly undercompensated and the eastern half is slightly overcompensated. This more detailed result is generally consistent with our previous larger scale result in Fig. 9c, but the amplitudes of under- and overcompensation, ~ 5 km, are better constrained in Fig. 10. In Fig. 9c, we overlay contours of average annual precipitation from the Tropical Rainfall Measuring Mission (TRMM), using data from the years 1998 through the present (Huffman and Bolvin, 2015). The anomalously thin crust beneath parts of the Eastern Cordillera (anomalies D and E) is located slightly southwest (or upstream) of the locations of maximum annual precipitation, as expected if orographic precipitation is driving rock uplift (Norton and Schlunegger, 2011). Of course, this relationship would only hold for as long as the current pattern of orographic precipitation has existed. Various studies of this question generally conclude the current pattern was established in the Late Miocene, 10-15 Ma (Gillis et al., 2006; Mosolf et al., 2010; Barnes et al., 2012). The northern region of anomalous thin crust (Anomalies D1 and D2) has peaks up to 6 km that correspond to Triassic and Miocene plutons exposed at the

surface (Gillis et al., 2006). Gillis et al. (2006) used $^{40}\text{Ar}/^{39}\text{Ar}$ fission track thermochronology and U-Pb geochronology in the Eastern Cordillera to estimate that ~7.5 km of exhumation occurred during the Eocene-Oligocene with concomitant crustal shortening and ~3-5 km of rapid exhumation occurred during the Miocene. At face value, this suggests the ~ 5 km thinning in the western EC may be attributable to climate driven erosion.

Crustal flow may also be an important process in explaining our isostatic compensation model of the central Andean crust (Husson and Sempere, 2003; Kay and Coira, 2009). Eichelberger, et al. (2015), present evidence that crustal flow has had a significant influence on the evolution of crustal thickness in this region. Eichelberger, et al. (2015) used estimates of the crustal shortening rates to predict the crustal thickness and compared their results with our present-day observations of Moho depth. Their 3D volumetric crustal models are sensitive to the shortening rate, constrained by structural reconstructions done across the CAP and the initial cratonic thickness, which is poorly constrained. They suggest that with either initial crustal thicknesses of 35 km or 40 km some amount of lithospheric removal as well as some lower crustal flow likely occurred over the last 25 Ma to account for the modern distribution of crustal thickness that we observe (Eichelberger et al., 2015). Lower crustal flow could provide a mechanism to redistribute crust from the Eastern Cordillera near La Paz (Fig. 9 anomalies D1 and D2 with thinner than expected crust) to the northern Altiplano (Figure 9 anomaly F) where there is thicker than expected crust.

Lithospheric removal is an alternative mechanism that could lead to areas that are not in isostatic equilibrium. Numerous studies have concluded that lithospheric removal has occurred throughout much of the Andes, and may have been an important factor in the uplift of the CAP, although studies disagree as to the size and timing of the material removed (Kay and Kay, 1993;

Yuan et al., 2000; Yuan et al., 2002; Beck and Zandt, 2002; Molnar and Garzzone, 2007; Beck et al., 2015). In contrast to the region near La Paz, the Eastern Cordillera near 19-20°S has thick crust (~65-70 km) that is slightly thicker than the Altiplano and appears to be near isostatic equilibrium at the Moho (Fig. 6b). We note that there is also less erosion or exhumation in this region as evidenced by the San Juan Del Oro surface, a regionally extensive undeformed geomorphic surface that is dated as ~10 Ma (Gubbels et al., 1993). The Eastern Cordillera landforms farther south are only slightly modified and the topographic relief of the eastern CAP margin is more gradual and consistent with less erosion (Masek et al., 1994). Still, the difference in Moho depths beneath the eastern EC and adjacent Interandean Zone across the two transects is dramatic (Fig. 10). There is strong geophysical and geochemical evidence for late Cenozoic lithospheric delamination beneath the Los Frailes Volcanic Field (Myers et al., 1998; Beck et al., 2002; Kato et al., 2014). Both numerical modeling of delamination of lithosphere with weak lower crust (Krystopowicz and Currie, 2013) and seismological imaging of young delamination beneath the Puna plateau (Beck et al., 2015) show the post-delamination lithospheric characteristics include high elevations, thick felsic crust, and thin lithosphere. Thus, south of the orocline at ~20° S, the much thicker and isostatically compensated crust can be attributed to a combination of the absence of erosional exhumation and the occurrence of lithospheric delamination. The lack of a crustal root and abundant backarc volcanism in the EC between ~18° to 14° S suggests that if late Miocene delamination has occurred in the Altiplano as suggested by Garzzone et al. (2006) it has not progressed into the Eastern Cordillera at these latitudes. Further validation of this preliminary conclusion regarding delamination awaits ongoing studies of the upper mantle structure in this region.

7. Conclusions

We have used our new receiver function results combined with previous work to investigate the crustal structure in the Bolivian orocline region of the Central Andes, in southern Peru and northern Bolivia. Our principal results are listed below:

1. We determined a well-resolved depth to Moho map (relative to sea level) that shows values that range from 40 km to 75 km. In general, we observe thick, 65-75 km crust underneath the Western Cordillera and Altiplano, and the thinnest crust, 40 km, beneath the Subandean zone. In the Eastern Cordillera, our results show variable thickness crust ranging from 50-70 km depending on the location.
2. We identified the top of a low velocity zone at ~15 km beneath the Western Cordillera, corresponding to a low-velocity zone previously identified through an ambient noise study (Ward, et al., 2013).
3. Our results show an increase in velocity under the Eastern Cordillera and Subandean zone along B-B' that corresponds with the basal décollement in the balanced cross-section from McQuarrie, et al., (2008). We interpret an increase in velocity beneath the Altiplano at 20 km as the top of the basement.
4. We found two areas in the Eastern Cordillera with thinner than expected crust based on local Airy isostasy. These areas are adjacent to locations of high precipitation and erosion suggesting that climate induced exhumation may have contributed to thinning the crust, in addition to the possibility that lower crust was redistributed through the localized effects of crustal flow.
5. Our results confirm earlier results of thick crust near isostatic equilibrium in the eastern Cordillera near latitude 20° S where the Los Frailes volcanic field is located. We

conclude the thicker crust here is due to a combination of much lower rates of erosion and the occurrence of late Cenozoic delamination at this latitude.

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9. References

- Allmendinger, R.W., & Gubbels T., 1996. Pure and simple shear plateau uplift, Altiplano-Puna, Argentina and Bolivia. *Tectonophysics*, **259**, 1-13, doi:10.1016/0040-1951(96)00024-8.
- Allmendinger, R. W., Jordan, T. E., Kay, S. M. & Isacks, B. L., 1997. The evolution of the Altiplano-Puna Plateau of the Central Andes. *Annu. Rev. Earth Planet. Sci.*, **25**, 139-174, doi:10.1146/annurev.earth.25.1.139.
- Ammon, C. J., Randall G. E., & Zandt, G, 1990. On the Nonuniqueness of Receiver Function Inversions. *J. Geophys. Res.*, **95**, 15,303-15,318, doi:10.1029/JB095iB10p15303.
- Ammon, C. J., 1991. The isolation of receiver effects from teleseismic P waveforms. *Bull. Seismol. Soc. Am.*, **81**, 2504-2510.
- ANCORP Working Group, 2003. Seismic imaging of a convergent continental margin and plateau in the central Andes (Andean Continental Research Project 1996 (ANCORP'96)), *J. Geophys. Res.*, **108**, 2328, doi:10.1029/2002JB001771.
- Assumpção, M., Feng, M., Tassara, A., Julià J., 2013. Models of crustal thickness for South America from seismic refraction, receiver functions and surface wave tomography, *Tectonophysics*, **609**, 82-96, doi: 0.1016/j.tecto.2012.11.014.
- Baby, P., Rochat, P., Mascle, G., & Herail, G., 1997. Neogene shortening contribution to crustal thickening in the back arc of the Central Andes, *Geology*, **25**, 883-886, doi:10.1130/0091-7613(1997)025<0883:NSCTCT>2.3.CO;2.
- Baker, M.C.W., 1981. The nature and distribution of upper Cenozoic ignimbrite centres in the Central Andes, *J. Volcanol. Geotherm. Res.*, **11**, 293-315, doi:10.1016/0377-0273(81)90028-7.
- Barnes, J.B., Ehlers, T.A., Insel, N., McQuarrie, N., Poulsen, C.J., 2012. Linking orography, climate, and exhumation across the central Andes, *Geology*, **40**, 1135-1138, doi: 10.1130/G33229.1,
- Beck, S.L., Zandt, G., Myers, S.C., Wallace, T.C., Silver, P.G. & Drake, L., 1996. Crustal-thickness variations in the central Andes, *Geology*, **24**, 407-410, doi:10.1130/0091-7613(1996)024<0407:CTVITC>2.3.CO;2.
- Beck, S.L. & Zandt, G., 2002. The nature of orogenic crust in the central Andes, *J. Geophys. Res.-Solid Earth*, **107**, 2230-2247, doi:10.1029/2000JB000124.
- Beck, S.L., Zandt, G., Ward, K.M., Scire, A., 2015. Multiple styles and scales of lithospheric foundering beneath the Puna Plateau, central Andes, *Geo. Soc. Mem.*, **212**, 43-60, doi:10.1130/2015.1212(03).
- Cassidy, J. F., 1992. Numerical experiments in broadband receiver function analysis, *Bull. Seismol. Soc. Am.*, **82**, 1453-1474.
- Chmielowski, J., Zandt, G. & Haberland, C., 1999. The Central Andean Altiplano-Puna Magma body, *Geophys. Res. Lett.*, **26**, 783-786, doi:10.1029/1999GL900078.
- Chulick, G.S., Detweiler, S. & Mooney W.D., 2013. Seismic structure of the crust and uppermost mantle of South America and surrounding oceanic basins, *J. South Am. Earth. Sci.*, **42**, 260-276, doi:10.1016/j.jsames.2012.06.002.

- DeCelles, P. G. and Horton, B. K., 2003. Early to middle Tertiary foreland basin development and the history of Andean crustal shortening in Bolivia, *Geol. Soc. Am. Bull.*, **115**, 58-77, doi:10.1130/0016-7606.
- DeCelles, P. G., Carrapa, B., Horton, B. K. & Gehrels, G. E., 2011. Cenozoic foreland basin system in the central Andes of northwestern Argentina: Implications for Andean geodynamics and modes of deformation, *Tectonics*, **30**, 1-30, doi:10.1029/2011TC002948.
- DeCelles, P. G., Ducea, M. N., Kapp, P. & Zandt, G., 2009. Cyclicity in Cordilleran orogenic systems, *Nat. Geosci.*, **2**, 251-257, doi:10.1038/ngeo469.
- Dorbath, C., Granet, M., Poupinet, G. & Martinez, C., 1993. A teleseismic Study of the Altiplano and the Eastern Cordillera in Northern Bolivia: New Constraints on a Lithospheric Model, *J. Geophys. Res.*, **98**, 9825-9844, doi:10.1029/92JB02406.
- Dorbath, L., Dorbath, C., Jimenez, E., & Rivera, L., 1991. Seismicity and tectonic deformation in the Eastern Cordillera and the sub-Andean zone of central Peru, *J. South Am. Earth Sci.*, **4**, 13-24, doi:10.1016/0895-9811(91)90015-D.
- Dorbath, C., & Granet, M., 1996. Local earthquake tomography of the Altiplano and the Eastern Cordillera of northern Bolivia, *Tectonophysics*, **259**, 117-136, doi:10.1016/0040-1951(95)00052-6.
- Dueker, K. & Sheehan, A., 1997. Mantle discontinuity structure from midpoint stacks of converted P to S waves across the Yellowstone hotspot track. *J. Geophys. Res.*, **102**, 8313-8327, doi:10.1029/96JB03857.
- Eichelberger, N., & McQuarrie, N., 2015. Kinematic reconstruction of the Bolivian orocline, *Geosphere*, **11(2)**, 445-462, doi:10.1130/GES01064.1.
- Eichelberger, N., McQuarrie, N., Ryan, J., Karimi, B., Beck, S., & Zandt, G., 2015. Evolution of crustal thickening in the central Andes, Bolivia, *Earth Planet. Sci. Lett.*, **426**, 191-203, doi:10.1016/j.epsl.2015.06.035
- Elger, K., Oncken, O. & Glodny, J., 2005. Plateau-style accumulation of deformation: Southern Altiplano, *Tectonics*, **24**, 1-19, doi:10.1029/2004TC001675.
- Feng, M., van der Lee, S., Assumpção, M., 2007. Upper mantle structure of South America from joint inversion of waveforms and fundamental mode group velocities of Rayleigh waves. *J. Geophys. Res.*, **112**, 1-16, doi:10.1029/2006JB004449.
- Frassetto, A., Zandt, G., Gilbert, H., Owens, T. J., Jones, C. J., 2010. Improved imaging with phase-weighted common conversion point stacks of receiver functions. *Geophys. J. Int.*, **182**, 368-374, doi: 10.1111/j.1365-246X.2010.04617.x.
- Frassetto, A., & Thybo, H., 2013. Receiver function analysis of the crust and upper mantle in Fennoscandia – isostatic implications, *Earth Planet. Sci. Lett.*, **381**, 234-246, doi: 10.1016/j.epsl.2013.07.001.b
- Garzzone, C. N., Molnar, P., Libarkin, J. C., & MacFadden, B. J., 2006. Rapid late Miocene rise of the Bolivian Altiplano: Evidence for removal of mantle lithosphere, *Earth Planet. Sci. Lett.*, **241**, 543-556, doi:10.1016/j.epsl.2005.11.026.
- Garzzone, C. N., Hoke, G. D., Libarkin, J. C., Withers, S., MacFadden, B., Eiler, J., Ghosh, P., Mulch, A., 2008. Rise of the Andes, *Science*, **320**, 1304-1307, doi: 10.1126/science.1148615.

- Giese, P., Scheuber, E., Schilling, F., Schmitz, M. & Wigger, P., 1999. Crustal thickening processes in the central Andes and the different natures of the Moho-discontinuity, *J. South Am. Earth Sci.*, **12**, 201-220, doi: 10.1016/S0895-9811(99)00014-0.
- Gillis, R. J., Horton, B. K. & Grove, M., 2006. Thermochronology, geochronology, and upper crustal structure of the Cordillera Real: Implications for Cenozoic exhumation of the central Andean plateau, *Tectonics*, **25**, 1-22, doi:10.1029/2005TC001887.
- Gotberg, N., McQuarrie, N., & Caillaux, V. C., 2010. Comparison of crustal thickening budget and shortening estimates in southern Peru (12-14°S): Implication for mass balance and rotations in the “Bolivian orocline,” *Geol. Soc. Am. Bull.*, **122**, 727-742, doi:10.1130/B26477.1
- Graeber, F. M. & Asch, G., 1999. Three-dimensional models of P wave velocity and P-to-S velocity ratio in the southern central Andes by simultaneous inversion of local earthquake data, *J. Geophys. Res.*, **104**, 20,237-20,256, doi:10.1029/1999JB900037.
- Hayes, G. P., Wald, D. J. & Johnson, R. L., 2012. Slab1.0: a three-dimensional model of global subduction zone geometries, *J. Geophys. Res.-Sol Ea*, **117**, doi:10.1029/2011JB008524.
- Hoke, G.D., & Garziona, C. N., 2008. Paleosurfaces, paleoelevation, and the mechanisms for the late Miocene topographic development of the Altiplano plateau, *Earth Planet. Sci. Lett.*, **271**, 192-201, doi:10.1016/j.epsl.2008.04.008.
- Horton, B.K. 1999. Erosional control on the geometry and kinematics of thrust belt development in the central Andes, *Tectonics*, **18**, 1292-1304, doi:10.1029/1999TC900051.
- Huffman, G.J. & Bolvin, D.T. 2012, last updated 2015: 3B42.V7.TMPA. NASA/GSFC, Greenbelt, MD, USA, NASA Goddard Earth Sciences Data and Information Services Center (GES DISC). Accessed 05/16/14 at <http://pmm.nasa.gov/data-access/downloads/trmm>.
- Husson, L. & Sempere, T., 2003. Thickening the Altiplano crust by gravity-driven crustal channel flow, *Geophys. Res. Lett.*, **30**, 1243, doi:10.1029/2002GL016877.
- Insel, N., Poulsen, C.J., Ehlers, T., Sturm, C., 2012. Response of meteoric $\delta^{18}\text{O}$ to surface uplift-Implications for Cenozoic Andean Plateau growth, *Earth Planet. Sci. Lett.*, **317-318**, 262-272, doi:10.1016/j.epsl.2011.11.039.
- Isacks, B. L., 1988. Uplift of the Central Andean Plateau and bending of the Bolivian Orocline, *J. Geophys. Res.-Solid Earth*, **93**, 3211–3231, doi:10.1029/JB093iB04p03211.
- Kato, J.J., Kay, S.M., Coira, B.L., Jicha, B.R., Harris, C., Caffè, P.J., Jimenez, N., 2014. Evolution and geochemistry of the Neogene Los Frailes ignimbrite complex on the Bolivian Altiplano Plateau, XIX Congreso Geológico Argentino, Junio 2014, Córdoba, in Volcanismo De Los Andes - Volcanismo Silíceo.
- Kay, S.M., and Coira, B.L., 2009, Shallowing and steepening subduction zones, continental lithospheric loss, magmatism, and crustal flow under the central Andean Altiplano-Puna Plateau, in Kay, S.M., Ramos, V.A., and Dickinson, W.R., eds., Backbone of the Americas: Shallow Subduction, Plateau Uplift, and Ridge and Terrane Collision: *Geo. Soc. Mem.* **204**, p. 229-260, doi: 10.1130/2009.1204(11).
- Kay, R. W. & Kay, S. M., 1993. Delamination and delamination magmatism, *Tectonophysics*, **219**, 177-189, doi:10.1016/0040-1951(93)90295-U.

- Kley, J., 1996. Transition from basement-involved to thin-skinned thrusting in the Cordillera Oriental of southern Bolivia, *Tectonics*, **15**, 763-775, doi: 10.1029/95TC03868.
- Kley, J., & Monaldi, C., 1998. Tectonic shortening and crustal thickness in the Central Andes: How good is the correlation? *Geology*, **26**, 723-726, doi:10.1130/0091-7613(1998)026<0723:TSACTI>2.3.CO;2.
- Kley, J., 1999. Geologic and geometric constraints on a kinematic model of the Bolivian orocline. *J. South Am. Earth Sci.*, **12**, 221-235, doi: 10.1016/S0895-9811(99)00015-2.
- Krystopowicz, N. J. & Currie, C. A., 2013. Crustal eclogitization and lithosphere delamination, *Earth Planet. Sci. Lett.*, **361**, 195-207, doi:10.1016/j.epsl.2012.09.056.
- Lamb, S., 2011. Did shortening in thick crust cause rapid Late Cenozoic uplift in the northern Bolivian Andes? *J. Geol. Soc. Lond.*, **168**, 1079–1092, doi:10.1144/0016-76492011-008.
- Leier, A., McQuarrie, N., Garzzone, C., Eiler, J., 2013. Stable isotope evidence for multiple pulses of rapid surface uplift in the Central Andes, Bolivia, *Earth Planet. Sci. Lett.*, **371-372**, 49-58, doi:10.1016/j.epsl.2013.04.025.
- Ligorria, J. P., & Ammon, C. J., 1999. Iterative Deconvolution and Receiver-Function Estimation. *Bull. Seismol. Soc. Am.*, **89**, 1395-1400.
- Lindquist, K. G., Engle, K., Stahlke, D. & Price, E., 2004. Global Topography and Bathymetry Grid Improves Research Efforts, *Eos Trans. AGU*, **85(19)**, 186.
- Lloyd, S., van der Lee, S., Sand França, G., Assumpção, M., Feng, M., 2010. Moho map of South America from receiver functions and surface waves, *J. Geophys. Res.*, **115**, 1-12, doi:10.1029/2009JB006829.
- Masek, J.G., Isacks, B.L., Gubbels, T.L. & Fielding, E.J., 1994. Erosion and tectonics at the margins of continental plateaus, *J. Geophys. Res.*, **99**, 13,941-13,956, doi:10.1029/94JB00461.
- McGlashan, N., Brown, L., & Kay, S., 2008. Crustal thickness in the central Andes from teleseismically recorded depth phase precursors, *Geophys. J. Int.*, **175**, 1013-1022, doi:10.1111/j.1365-246X.2008.03897.x
- McQuarrie, N., DeCelles, P., 2001. Geometry and structural evolution of the central Andean backthrust belt, Bolivia, *Tectonics*, **20**, doi: 10.1029/2000TC001232.
- McQuarrie, N., Horton, B. K., Zandt, G., Beck, S. & DeCelles, P. G., 2005. Lithospheric evolution of the Andean fold-thrust belt, Bolivia, and the origin of the central Andean plateau, *Tectonophysics*, **399**, 15-37, doi:10.1016/j.tecto.2004.12.013.
- McQuarrie, N., Barnes, J. B. & Ehlers, T. A., 2008. Geometric, kinematic, and erosional history of the Central Andean Plateau, Bolivia (15–17 degrees S), *Tectonics*, **27**, doi:10.1029/2006TC002054.
- Molnar, P. & England, P., 1990. Late Cenozoic uplift of mountain ranges and global climate change: chicken or egg? *Nature*, **346**, 29-34, doi: 10.1038/346029a0.
- Molnar, P. & Garzzone, C. N., 2007. Bounds on the viscosity coefficient of continental lithosphere from removal of mantle lithosphere beneath the Altiplano and Eastern Cordillera, *Tectonics*, **26**, 1-9, doi:10.1029/2006TC001964.
- Mosolf, J. G., Horton, B. K., Heizler, M. T., Matos, R., 2010. Unroofing the core of the central Andean fold-thrust belt during focused late Miocene exhumation: evidence from the

- Tipuani-Mapiri wedge-top basin, Bolivia, *Basin Res.*, **23**, 346-360, doi:10.1111/j.1365-2117.2010.00491.x.
- Myers, S., Beck, S., Zandt, G. & Wallace, T., 1998. Lithospheric-scale structure across the Bolivian Andes from tomographic images of velocity and attenuation for *P* and *S* waves, *J. Geophys. Res.*, **103**, 21,233-21,252, doi:10.1029/98JB00956.
- Norton, K. & Schlunegger, F., 2011. Migrating deformation in the Central Andes from enhanced orographic rainfall, *Nat. Commun.*, **2**:584, 1-7, doi: 10.1038/ncomms1590.
- Phillips, K., Clayton, R. W., Davis, P., Tavera, H., Guy, R., Skinner, S., Stubailo, I., Audin, L., Aguilar, V., 2012. Structure of the subduction system in southern Peru from seismic array data, *J. Geophys. Res.-Solid Earth*, **117**, B11306, doi:10.1029/2012JB009540.
- Phillips, K., Clayton, R. W., 2014. Structure of the subduction transition region from seismic array data in southern Peru, *Geophys. J. Int.*, **196**, 1889-1905, doi: 10.1093/gji/ggt504
- Poulsen, C.J., Ehlers, T., Insel, N., 2010. Onset of convective rainfall during gradual late Miocene rise of the central Andes, *Science*, **328**, 490–493.,doi: 10.1126/science.1185078.
- Schmitz, M. & Kley, J., 1997. The Geometry of the Central Andean Backarc Crust: Joint Interpretation of Cross-section Balancing and Seismic Refraction Data, *J. South Am. Earth Sci.*, **10**, 99-100, doi:10.1016/S0895-9811(97)00009-6.
- Sheehan, A. F., Shearer, P. M., Gilbert, H. J., & Dueker, K. G., 2000. Seismic migration processing of P-SV converted phases for mantle discontinuity structure beneath the Snake River Plain, western United States, *J. Geophys. Res.*, **105**, 19055–19065, doi:10.1029/2000JB900112.
- Storchak, D. A., Giacomo, D. D., Bondár, I., Engdahl, E. R., Harris, J., Lee, W. H. K., Villaseñor, A., Bormann, P., 2013. Public Release of the ISC-GEM Global Instrumental Earthquake Catalogue (1900-2009), *Seism. Res. Lett.*, **84**, 5, 810-815, doi: 10.1785/0220130034.
- Swenson, J., Beck, S. & Zandt, G., 2000. Crustal structure of the Altiplano from broadband regional waveform modeling: Implications for the composition of thick continental crust, *J. Geophys. Res.*, **105**, 607 – 621, doi:10.1029/1999JB900327.
- Tassara, A., Götze, H.-J., Schmidt, S. & Hackney, R., 2006. Three-dimensional density model of the Nazca plate and the Andean continental margin, *J. Geophys. Res.*, **111**, B09404, doi:10.1029/2005JB003976.
- van der Meijde, M., Julià, J. and M. Assumpção, M., 2013. Gravity derived Moho for South America, *Tectonophysics*, **609**, 456-467, doi: 10.1016/j.tecto.2013.03.023.
- Wada, I., K. Wang, J. He, and R. D. Hyndman (2008), Weakening of the subduction interface and its effects on surface heat flow, slab dehydration, and mantle wedge serpentinization, *J. Geophys. Res.*, **113**, B04402, doi:10.1029/2007JB005190.
- Ward, K. M., Porter, R. C., Zandt, G., Beck, S., Wagner, L., Minaya, E. & Tavera, H., 2013. Ambient noise tomography across the Central Andes, *Geophys. J. Int.*, **194**, 1559-1573, doi:10.1093/gji/ggt166.
- Whipple, K. X., & Gasparini, N.M., 2014. Tectonic control of topography, rainfall patterns, and erosion during rapid post–12 Ma uplift of the Bolivian Andes, *Lithosphere*, **6**, 251-268, doi: 10.1130/L325.1.

- Whitcomb, J. H., Garmany, J. D., Anderson, D. L., 1972. Earthquake Prediction: Variation of Seismic Velocities before the San Francisco Earthquake, *Science*, **180**, 632-635, doi: 10.1126/science.180.4086.632.
- Wigger, P. J., *et al.*, 1994. Variation in the Crustal Structure of the Southern Central Andes Deduced from Seismic Refraction Investigations, in *Tectonics of the Southern Central Andes*, pp.23-48, eds Reutter, K.-J., Scheuber, E. & Wigger, P.J., Springer-Verlag, Berlin, doi: 10.1007/978-3-642-77353-2_2.
- Wölbern, I., Heit, B., Yuan, X., Asch, G., Kind, R., Viramonte, J., Tawackoli, S. & Wilke, H., 2009. Receiver function images from the Moho and the slab beneath the Altiplano and Puna plateaus in the Central Andes, *Geophys. J. Int.*, **177**, 296-308, doi: 10.1111/j.1365-246X.2008.04075.x.
- Yuan, X. *et al.*, 2000. Subduction and collision processes in the Central Andes constrained by converted seismic phases, *Nature*, **408**, 958–961, doi: 10.1038/35050073.
- Yuan, X., Sobolev, S. V. & Kind, R., 2002. Moho topography in the central Andes and its geodynamic implications, *Earth planet. Sci. Lett.*, **199**, 389–402, doi: 10.1016/S0012-821X(02)00589-7.
- Zandt, G., Beck, S., Ruppert, S., Ammon, C., Rock, D., Minaya, R., Wallace, T. & Silver, P., 1996. Anomalous crust of the Bolivian Altiplano, central Andes: Constraints from broadband regional seismic waveforms, *Geophys. Res. Lett.*, **23**, 1159– 1162, doi:10.1029/96GL00967.
- Zandt, G., Leidig, M., Chmielowski, J., Baumont, D. & Yuan, X. H., 2003. Seismic detection and characterization of the Altiplano-Puna Magma Body, central Andes, *Pure Appl. Geophys.*, **160**, 789–807, doi: 10.1007/PL00012557.
- Zandt, G., Gilbert, H., Owens, T. J., Ducea, M., Saleeby, J. & Jones., C. H., 2004. Active foundering of a continental arc root beneath the southern Sierra Nevada in California, *Nature*, **431**, 41-46, doi: 10.1038/nature02847.
- Zhu, L., & Kanamori, H., 2000. Moho depth variation in southern California from teleseismic receiver functions, *J. Geophys. Res.*, **105**, 2969-2980, doi:10.1029/1999JB900322.

Array Name	Number of Stations Used	Time Period	Reference
CAUGHT (doi:10.7914/SN/ZG_2010)	51	10/2010 – 08/2012	1
BANJO-SEDA	23	05/1994 - 08/1995	1
Global Telemetered Seismograph Network (LPAZ, doi:10.7914/SN/IU)	1	08/1993 - present	2
Observatorio San Calixto	6	02/2012-present	3
Chilean National Seismic Network (GO01)	1	08/2011 - present	4
Plate Boundary Project; GFZ portion of IPOC	6	12/2006 - present	4

Table 1: Arrays, number of stations, and time period of the deployment for the stations used to create and migrate the receiver functions presented in this study. References are as follows:

1) IRIS DMS

IRIS Data Management System, Seattle, Washington

The facilities of the IRIS Data Management System were used for access to waveform, metadata or products required in this study. The IRIS DMS is funded through the National Science Foundation Cooperative Agreement EAR-1063471. Some IRIS activities are supported by the National Science Foundation EarthScope Program under Cooperative Agreements EAR-0733069, EAR-1261681.

2) GSN / IRIS Global Seismographic Network

(including H2, IC, II, IU, MS)

Global Seismographic Network is a cooperative scientific facility operated jointly by the Incorporated Research Institutions for Seismology (IRIS), the United States Geological Survey (USGS), and the National Science Foundation (NSF). <http://www.iris.edu/hq/programs/gsn>

3) Observatorio of San Calixto

The San Calixto Observatory is a private non-profit seismic observatory.

4) GE / (GEOFON)

GFZ Potsdam, Germany

The GEOFON network is funded and operated by GFZ Potsdam, Germany, in co-operation with almost 50 institutions worldwide.

<http://geofon.gfz-potsdam.de/geofon/>

Figure 1: Topographic map of the central Andes and geomorphic province boundaries for the Western Cordillera (WC), Altiplano (AP), Eastern Cordillera (EC), Interandean (IA), and Subandean Zone (SA) modified from McQuarrie et al., (2008) and Mosolf et al., (2010); Holocene volcanoes shown as red triangles. The Los Frailes Volcanic Field (LFVF) is outlined in dashed red. Yellow stars show the location of major population centers, and purple circles show the location of local events used in the Wadati diagram (Fig. 4).

Figure 2: a) Map of the seismic arrays from Table 1, and the cross-section locations of receiver function CCP stacks described in the text. Slab contours are plotted based on the slab1.0 model of Hayes et al., (2012). b) Global projection showing the location of the 343 earthquakes recorded by the stations shown in Figure 1 and used for calculation of the teleseismic receiver functions used in this study (list available in Supplemental Table T1).

Figure 3: Back-azimuth receiver function plots for four sample stations along the dense portions of the CAUGHT array. The three CAUGHT stations (CP11, CB25, CB2D) show similar back-azimuth coverage to the permanent GTSN station LPAZ; Moho arrival picks are outlined on the radial waveforms. An unstacked version can be found in Supplemental Figure S1.

Figure 4: Wadati diagram using 9 local events recorded on the CAUGHT network to constrain V_p/V_s . The solid line is a linear regression through the individual event-station points, giving a slope of $(V_p/V_s - 1)$ (Whitcomb et al. 1972). Arrival times were removed prior to plotting to obtain an average V_p/V_s based on the nine events.

Figure 5: CAP cross-section receiver function stack, at 1-to-1 scale, using a Gaussian pulse-width of $a=2.5$ and a phase-weighting filter strength of .25. Location of the cross-sections is shown on Figure 1b. The receiver function stacks are depicted as black traces drawn from the surface. International Seismological Centre earthquake locations (Storchak et al., 2013) are plotted as small black dots. The slab contour (Hayes et al., 2012) is shown as a heavy dashed black line. Positive Ps conversions are shown in red and negative conversions in blue, with the colors saturating at $\pm 15\%$ of the initial P arrival amplitude. Dark grey arrows show the locations where cross-section lines change direction, while the D and E markers on the section indicate the cross-over locations for the D-D' and E-E' line, respectively. Abbreviations for topographic provinces are same as in Fig. 1. TLVZ = top of low velocity zone. Bin spacing for the line is 15 km along the section, and 30 km perpendicular to the section. A phase-weighting comparison image is available (Supplemental Figure S3).

Figure 6a-d: Cross-section CCP receiver function stack, at 1-to-1 scale, using a Gaussian pulse-width of $a=2.5$ and a phase-weighting filter strength of .25. Labels are the same as in Figure 5, with locations of the lines shown in Figure 1b. Cross-section crossover locations are marked on lines B-B' and C-C'. THVZ = top of high velocity zone, other abbreviations same as figure 5. Bin spacing for these sections is 50 km both along and perpendicular to the section.

Figure 7: Depth to Moho map, relative to sea level, derived from our receiver functions at the individual stations and contoured using a nearest neighbor interpolation algorithm. Contours are drawn every 5 km until reaching the 40 km contour. The forearc and foreland are constrained using gravity measurements from Tassara et al., (2006). Anomalies D and E correspond to anomalies D and E marked on Figure 9. The bins used to derive this map are depicted in Supplemental Figure S2.

Figure 8: CCP stacked receiver function cross-section overlain with the balanced cross-section of McQuarrie, et al., (2008) to a depth of 75 km. Images are at a 1-to-1 scale and aligned based on tectonic province information from McQuarrie, et al., (2008). Anomalies A, B and C are all positive arrivals corresponding to transitions from low to high velocity that we interpret as the location of basement rock.

Figure 9: Crustal thickness maps calculated from a) receiver functions and b) Airy isostatic compensation depth for 100 km² filtered topography. c) Crustal thickness residual map created by subtracting the predicted Airy isostatic thickness using the filtered topographic model (100 km²) from the observed thickness based on receiver functions. Also plotted are contoured precipitation data taken from the TRMM 3B42 database (1998-2013) at contour intervals of 0.2 meter per year. Crustal residual information in the

forearc has been omitted due to lack of constraints from the observed receiver functions. Anomaly F, in the Altiplano, is an area with thicker than expected crust, while anomalies D1, D2, and E are thinner than our Airy isostasy calculations would predict.

Figure 10: A comparison of depth to Moho and topography profiles along two cross-sections of the Central Andes. The cross-section in brown is at 19.5°S, the BANJO-SEDA line, and the cross-section in blue is the dense CAUGHT line near La Paz, Bolivia. The cross-sections are aligned along the Altiplano-Eastern Cordillera boundary. Note the difference in the depth to Moho beneath the Eastern Cordillera and the difference in topographic profile along the two cross-sections. The vertical exaggeration below sea level in this image is 2:1.