THE STRUCTURE OF THE CRUST AND UPPERMOST MANTLE BENEATH THE CENTRAL ANDES FROM AMBEINT NOISE TOMOGRAPHY: IMAGING THE NEOGENE TO MODERN BATHOLITH

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As members of the Research Committee, we recommend that this prepublication manuscript be accepted as fulfilling the research requirement for the degree of Master of Science.

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THE STRUCTURE OF THE CRUST AND UPPERMOST
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THE NEOGENE TO MODERN BATHOLITH

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0. Summary

The Central Andes of southern Peru, Bolivia, and northern Chile (between 12°S and 26°S) are comprised of a large orogenic plateau, the Central Andean Plateau (CAP). Despite numerous geological and geophysical studies across the plateau, the uplift history of the plateau remains uncertain. Integrating data from 151 broadband seismic stations, we use Ambient Noise Tomography (ANT) to image the shear-wave velocity structure of the crust and uppermost mantle of the CAP. Our major results include: (1) The high elevations of the CAP are supported by a thick (~70 km) low-velocity (and presumably low-density) crust. (2) A pervasive mid-crustal low-velocity zone underneath the western portion of CAP near the west side of the active arc (related to a thermally weakened crust). (3) A strong mid-crustal low-velocity zone beneath the Altiplano-Puna volcanic complex and the Los Frailes ignimbrites. The presence of a large and laterality extensive low-velocity zone suggests either a Neogene batholith at depth, a thermally weakened crust capable of lateral flow, or both. Our interpretation of a large Neogene batholith associated with active volcanism revisits the idea of magmatic addition as a contributing mechanism to the growth of the western portion of the CAP.

1. Introduction

The South American Cordillera is considered the type example of a convergent margin where oceanic lithosphere is subducting underneath continental lithosphere inducing a compressional stress regime in the overriding plate. Commensurate with oceanic subduction, abundant arc magmatism is observed along the South American margin with large volcanic gaps associated with flat-slab subduction geometries in the
subducting Nazca Plate [Ramos and Folguera, 2009; Ramos, 1999]. In the Central Andes, the Central Volcanic Zone (CVZ) consists of Holocene volcanic activity (between 14°S and 26°S) truncated to the north by Peruvian flat-slab subduction and to the south by Pampean flat-slab subduction [Cahill and Isacks, 1992; Siebert and Simkin, 2002-]. Along this stretch of the South American Cordillera, the Andean mountain belt is widest and highest forming the largest high plateau in the world associated with arc magmatism. The CAP as defined by the 3 km elevation contour extends over 1,800 km along the active margin reaching a maxim width around 400 km (Figure 1) [Allmendinger et al., 1997]. Despite the Plateau’s impact on the dynamic evolution of the Central Andes, the structure and uplift history of the CAP remains a lingering question in the geosciences.

Three competing tectonic processes have emerged as possible mechanisms for the Plateau’s uplift since ~10 Ma: (1) slow and steady uplift driven by underthrusting of the Brazilian shield associated with shortening in the eastern Plateau [Barke and Lamb, 2006; Jordan et al., 1997], (2) rapid uplift associated with isostatic rebound following delamination of the lower Altiplano-Puna crust [Garzione et al., 2006; Schurr et al., 2006; Kay et al., 1994], and (3) thickening of the Altiplano crust through lower crustal flow possibility augmented by magmatic addition or underplating [Husson and Sempere, 2003]. Each model makes specific predictions about the evolution of the plateau’s climate, elevation, lithospheric thickness, composition, and magnitude of deformation. In part, the controversy between models persists because of a lack in fundamental data sets that overlap in time and space. As part of the Central Andes Uplift and the Geodynamics of High Topography (CAUGHT) project, the main objective of the seismological component is to map the crustal and upper mantle structure of the Central Andes using
multiple techniques including Ambient Noise Tomography. In this ANT study, we use
data from 151 broadband seismic stations from 12 different international seismic
networks, deployed incrementally in the Central Andes from May 1994 through March
2012, to image the vertically polarized shear-wave velocity (Vsv) structure of the Central
Andes (Figure 1). A complete shear-wave velocity model will provide constrains on the
formation of the CAP.

2. Tectonic Setting

The Central Andes of southern Peru, Bolivia, and northern Chile (between 12°S
and 26°S) form a typical Andean-type orogeny where tectonics is driven by oblique
subduction of the oceanic Nazca Plate under the continental South American Plate. The
geometry of the east-dipping subducting slab is normal (~30°) and bounded to the north
(14°S) and south (26°S) by flat-slab segments [Cahill and Isacks, 1992]. Characteristic
volcanic cessation associated with the onset of the Peruvian (11 Ma) and Pampean (12
Ma) flat-slab subduction defines the northern and southern termination of the CVZ of the
Central Andes [Ramos and Folguera, 2009]. Estimates of the current rate of convergence
between the Nazca and South American Plates are in the range of 58-80 mm yr\(^{-1}\)
[Kendrick et al., 2003] with up to 15 mm yr\(^{-1}\) distributed over a broad zone of
deformation that is the Central Andes [Norabuena et al., 1998]. Preferred shortening
estimates are largest (>300 km) where the Andean orogen is widest (~20°S) with active
deformation over 800 km inland from the trench [see summary in Gotberg et al., 2010].
With the exception of the Altiplano-Puna, the major morphotectonic structures can be
traced along the entire length of the CAP and from west to east include: the forearc, the
Western Cordillera, the Altiplano-Puna, the Eastern Cordillera, and the Subandes (Figure 2).

The width of the forearc remains relatively constant throughout the study area (~200 km) with a notable exception around 23°S where the active volcanic arc forms a westward concave bend that encompasses the Atacama Basin. Tectonic erosion since the late Neogene has led to the onset of subsidence and normal faulting in the offshore section of the forearc [Clift et al. 2003; von Huene and Ranero 2003]. Normal faulting in the forearc continues onshore where the tectonic regime transitions from subsidence to uplift forming the Costal Cordillera [Allmendinger and González, 2010]. The Chilean Precordillera, the eastern most extent of the onshore forearc has experienced near-uniform trenchward tilt during the Neogene and forms the western flank of the CAP [Isacks 1988; Victor et al. 2004; Jordan et al., 2010].

The Holocene aged volcanic arc that defines the Western Cordillera is a collection of andesitic to dacitic stratovolcanoes that have erupted through older ignimbrite sheets. Variations in the mode of subduction since the mid-Eocene is responsible for a wider volcanic arc during the early Miocene that thermally weakened the lithosphere between 14°S and 24°S [Ramos and Folguera, 2009; Mamani et al., 2010; Allmendinger et al., 1997]. Previous seismological studies have imaged a consistently thick (~70 km) central Andean crust below the active volcanic arc and compositional variations in subduction-related igneous lithologies suggest the crustal thickness has been increasing since the mid-Oligocene [Schmitz et al., 1999; Beck and Zandt, 2002; Yuan et al., 2002; Mamani et al., 2010].
The Altiplano and Puna comprise the core of the CAP and are characterized by areas of high elevation with low (Altiplano) to moderate (Puna) relief. During the Cenozoic, the Altiplano has been the locus of sedimentation, with thick sequences of red-beds, evaporites, and volcanics [Sobel et al., 2003; Strecker et al., 2007]. The Corque–Corocoro structure, a fold and thrust zone that deforms thick sections of Eocene or Oligocene to Pliocene red-bed sequences occupies most of the northern Altiplano [Lamb, 2011]. Lake Poopo and the Salar de Uyuni are the remnants of a large paleolake referred to as the Lake Tauca highstand that covered the entire southern Altiplano as recently as 14.1 ka [Placzek et al., 2006]. Located at the transition between the Altiplano (average elevation 3,700 m) and the Puna (average elevation 4,200 m) is the Altiplano-Puno Volcanic Complex (APVC), a large silicic volcanic field with multiple young (~12 Ma to 1 Ma) ignimbrite eruptions [de Silva, 1989; Salisbury et al., 2011]. Underlying the APVC at a depth of 15-20 km is a very-low-velocity zone (~1 km s⁻¹ shear-wave velocities) referred to as the Altiplano-Puno Magma Body (APMB) [Chmielowski et al., 1999; Zandt et al., 2003].

The high topography that defines the eastern flank of the CAP and the Eastern Cordillera is composed of folded and faulted Paleozoic rocks intruded by Triassic and Tertiary granitoid bodies [Gillis et al., 2006; Benjamin et al., 1987]. In the north where the plutons outcrop along the western section of the Eastern Cordillera, elevations can exceed 6,000 m and form the Cordillera Real. The eastern section of the Eastern Cordillera is predominantly the oldest stratigraphic section of the orogen with Ordovician aged and locally older rocks outcropping.
The Subandean Zone accommodates the transition from the high elevations of the CAP to the foreland plains through synclinal basins separated by thrust-faulted anticlines [McQuarrie et al., 2008]. The Alto Beni syncline, a piggyback basin that marks the western most extent of the Subandean Zone, has accommodated 6,500-7,000 m of Tertiary fill [Baby et al., 1995].

3. Data and Methods

The fundamental basis for ANT is that the cross-correlation of ambient seismic noise as recorded by two contemporaneously operating seismometers can be used to extract surface wave empirical Greens functions (EGFs) for the path between the two stations [Sabra et al., 2005; Shapiro et al., 2005]. While in principle both body waves [Roux et al., 2005] and surface waves [Shapiro et al., 2005; Moschetti et al., 2007; Yang et al., 2007] can be measured in this way, the main sources of coherent ambient noise are surface waves which dominates the amplitude of the Greens function between seismic stations making them easier to extract. The use of an automated frequency time analysis (FTAN) to measure the dispersion of group and phase velocities from inferred Rayleigh waves EGFs is well established in continental areas [Benson et al., 2007; Moschetti et al., 2007, Lin et al., 2008]. The spatial resolution of ANT is governed by the number, density, and azimuthal distribution of good quality interstation dispersion measurements, which to first order, are a function of the number of seismic stations in the array, the geometry of the array, and the length of the time window for which contemporaneous seismic data is available. For this reason, seismic arrays with close interstation spacing,
large footprints, and deployment durations of a year or longer are best suited for an ambient noise study [Benson et al., 2007].

3.1 Data and Initial Processing

We recently deployed 50 broadband seismic stations as part of the National Science Foundation (NSF) Continental Dynamics (CD) project: Central Andes Uplift and the Geodynamics of High Topography (CAUGHT) in the Central Andes of northern Bolivia and southern Peru. We augment the CAUGHT network with seismic data from other temporary deployments including 14 PULSE stations (PerU Lithosphere and Slab Experiment), 20 IPOC stations (Integrated Plate boundary Observatory Chile), 5 PLUTON stations (Magma intrusion beneath Andean supervolcanoes), and 5 PeruSE stations (Peru Subduction Experiment), to construct an array ideally suited for an ambient noise study of the Central Andes (Figure 1). Data from an additional 4 permanent stations including NNA, LCO, and LVC from the Global Seismic Network (GSN) and LPAZ from the Global Telemetered Seismic Network (GTSN) are also included in our study.

An intermediate step in the data processing is inverting for 2-D phase velocity maps from interstation dispersion curves. The horizontal extent and resolution of these 2-D phase velocity results are limited by the number and azimuthal distribution of crossing interstation ray paths. Although the ambient noise method requires the cross-correlation of contemporaneously operating seismometers, we exploit the methods ability to incorporate interstation dispersion curves measured from previous deployments by supplementing our main array with data from four additional temporary deployments spanning three additional time periods (Table 1). In total, we use data from 151
broadband seismic stations from 12 different international seismic networks, deployed incrementally in the Central Andes from May 1994 through March 2012, to image the crust of the Central Andes using ambient noise tomography.

We follow the method outlined by Benson et al. [2007] to obtain and quality control interstation surface wave dispersion measurements and briefly describe the data processing scheme used in this paper. Individual seismic station processing in the time domain includes cutting the waveform data into single-day segments, removing the mean, trend, and instrument response, followed by applying a 5 to 150 s bandpass filter. Temporal normalization is done using a running-absolute-mean normalization that acts to identify and remove the obscuring signals of earthquakes, instrument irregularities, and non-stationary noise sources. The normalized time series are transformed into the frequency domain using a Fast Fourier Transform (FFT) and spectrally whitened to produce a broader-band signal. Single-day segments from each possible station pair are cross-correlated, transformed back into the time domain, and stacked to increase the signal-to-noise ratio (SNR). The causal and acausal components of the cross-correlated time function represent surface waves traveling in opposite directions between stations and are added to each other to reduce the effects of inhomogeneous source distribution. The resulting symmetric waveform component is an EGF from which a frequency-time analysis is performed to measure the phase velocities at every integer period between 8 and 50 seconds. The automation of calculating large numbers of interstation dispersion measurements allows for the introduction of spurious measurements into the data set and quality control measures must be applied. An initial attempt to remove obvious outliers discards dispersion measurements with phase velocities below 1.5 km s\(^{-1}\) and above 5.0
km s$^{-1}$. Individual phase velocities at periods with less than twice the wavelength of the interstation distance or with SNRs of less than 10 are removed from the dispersion measurements.

3.2 Phase Velocity Inversion

We use the remaining dispersion measurements to invert for phase velocity maps at every integer period between 8 and 50 s using the method outlined by Barmin et al. [2001]. As a final measure of quality control, dispersion measurements with residuals greater than three seconds are removed reducing the total number of interstation paths used in this study to 32-63% of the maximum 4,591 possible paths (Table S1). The relativity dense station spacing used in this study allows us to partition our study area into small 0.1° by 0.1° grid points and use less restrictive regularization parameters in the inversion. The penalty function used in the inversion is dependent on three user defined regularization parameters: $\alpha$ is a measure of the damping, $\beta$ controls how data is smoothed in areas of varying path density, and $\sigma$ is the Gaussian smoothing width in km.

As suggested by Barmin et al. [2001], the selection of the best regularization parameters is somewhat arbitrary but can be improved upon by systematically varying the parameters and comparing the results with a priori information. We rigorously test the effects of varying the damping value, smoothing value, and smoothing length and conclude that the best regularization parameters for our study area are $\alpha = 200$, $\beta = 100$, and $\sigma = 100$ km. Inverting for phase velocity maps also yields a measure of spatial resolution at each grid point that is defined as the minimum separation distance at which two $\delta$-like anomalies can be resolved [Barmin et al., 2001].
3.2.1 Sensitivity Test for Phase Velocity Inversion

We test the sensitivity of our phase velocity results by investigating the effects of systematically varying the regularization parameters and grid spacing used in the inversion. A comparison of phase velocity maps using 0.25° by 0.25° with 0.1° by 0.1° grid spacing introduces minimal variation in the model results and we prefer the smooth appearance of the shear-wave inversion results by using 0.1° by 0.1° grid spacing. Varying the regularization parameters $\beta$ and $\sigma$ affects the amplitudes of anomalies on the edge of our array with lower $\beta$ and $\sigma$ values allowing for larger amplitude ranges. We address the arbitrary nature of selecting the regularization parameters $\beta$ and $\sigma$ by favoring conservative values ($\beta = 100$ and $\sigma = 100$ km) and focusing our interpretations on areas with good (“bright”) resolution. The selection of the regularization parameter $\alpha$ (damping value) has the largest effect on our phase velocity results and warrants careful consideration.

We introduce a new metric that we call the ‘normalized roughness’ in an attempt to quantify the best damping value to use for our study area as defined by equation 1.

$$\xi(\alpha, \lambda, \phi) = \frac{1}{T_{\text{max}} - T_{\text{min}}} \sum_{i=T_{\text{min}}}^{T_{\text{max}} - 1} |c_{i+1}(\alpha, T_{i+1}, \lambda, \phi) - c_i(\alpha, T_i, \lambda, \phi)|$$  

where $\xi$ is the normalized roughness, $\alpha$ is the damping value, $\lambda$ is the latitude, $\phi$ is the longitude, $T$ is the period, and $c$ is the phase velocity. The normalized roughness is a measure of the average rate of change between periods for a given 1-D phase velocity profile as a function of damping value and location. However, even a perfectly resolved model would exhibit varying average rates of change between periods as a function of location due to the seismic heterogeneity of the Earth. We remove the effect of seismic
heterogeneity by defining a reference normalized roughness calculated from the highest
damping value used and subtract it from the normalized roughness calculated at every
location for every damping value following equation 2.

\[ \xi'(\alpha, \lambda, \phi) = \xi(\alpha, \lambda, \phi) - \xi(\alpha_{\text{max}}, \lambda, \phi) \] (2)

Plotting the normalized roughness \((\xi')\) against the damping value \((\alpha)\) used at every point
in the array with at least fair resolution, we form a qualitative measure of selecting \(\alpha = 200\)
as the best damping value to use for our study area (Figure S1).

### 3.3 Shear Velocity Inversion

From the phase velocity maps at 8, 10, 12, 14, 16, 20, 25, 30, 35, and 40 seconds,
we construct phase velocity profiles at every grid point in our array. At each grid point
where all ten periods used to construct the phase velocity profiles have at least fair
resolution (<800 km), we iteratively invert for the 1-D shear-wave velocity structure
[Herrmann, 1987; Snoke and James, 1997; Larson et al., 2006]. We use a linearized least
square inversion method that requires a 1-D shear-wave velocity starting model and then
iteratively inverts for the best 1-D shear-wave velocity structure that minimizes the misfit
between the calculated and observed 1-D phase velocity profile [Herrmann, 1987]. After
three iterations, we observe no significant reduction in the misfit to the data and terminate
the iterative inversion.

### 3.3.1 Sensitivity Test for Shear Velocity Inversion

Inverting the phase velocity profiles for 1-D shear-wave velocity profiles adds an
additional complexity in quantifying the uncertainty of our results. We investigate the
effects of varying the velocity structure and layer thicknesses of the 1-D shear-wave
velocity starting profile as well as the effects of fitting phase velocity profiles generated
using different damping values at 10 representative locations in our study area (Figure 2).
Surface waves at the periods measured have broad sensitivity kernels (Figure S2) that
broaden with increasing wavelengths making them less sensitive to sharp velocity
contrasts such as the Moho. Previous work over limited regions of our study area
interpreted large variations in the Moho depth both perpendicular to and along strike of
the Andes [Beck and Zandt, 2002; Yuan et al., 2002; Sodoudi et al., 2011; Tassara and
Echaurren, 2012], which make extrapolating a detailed contour map of the Moho for our
entire study at the scale we are able to resolve problematic. Hence, we favor a constant
velocity starting model and allow the iterative inversion technique to model vertical
velocity variations.

We test four different constant velocity-starting models at 10 representative
locations and observe a negligible (<0.05 km s\(^{-1}\)) effect on the 1-D shear-wave results
above 70 km. We use the depth at which the constant velocity-starting model dependent
1-D shear-wave velocity results diverge as a proxy for the lower limit of our depth
resolution (~60-70 km across our study area). For each of the 10 representative locations,
we systematically invert for the 1-D shear-wave velocity profile using each possible
permutation of three differently damped input phase velocity models (150, 200, 250), two
different layer thicknesses in the top 75 km of the starting model (1 km and 5 km), three
Vp/Vs ratios that are held constant (1.70, 1.75, and 1.80), three different Q values (50,
200, 500), and four different constant velocity-starting models (3.1, 3.6, 4.1 and 4.6 km
s\(^{-1}\)). The minimum and maximum shear-wave results for any given depth at each location
define an envelope of uncertainty in our model results as shown in Figure 3. The relatively tight nesting of our shear-wave results across all end-member input values and different morphotectonic structures indicate our results are data driven and less dependent on a priori assumptions.

4. Results

We show the 2-D phase velocity results obtained using the regularization parameters of $\alpha = 200$, $\beta = 100$, and $\sigma = 100$ km.

4.1 Phase Velocity Results

Comparing the phase velocity profiles constructed at each of the ten representative points demonstrates the large range of modeled phase velocities in our study area (Figure 4). For periods that preferentially sample the upper crust, 8-20 s, the measured phase velocities of the Chilean forearc are more than 1.0 km s$^{-1}$ faster than phase velocity measurements made for the same periods at the APVC location. The lowest phase velocities we observe, $\sim$2.4 km s$^{-1}$ are nested in an area defined by the junction of the Bolivian, Argentinian, and Chilean political boundaries, reaching a minimum between 12 and 16 s. The general character of the phase velocity profiles between different locations in the same morphotectonic providences are consistent and vary by only $\sim$0.1 km s$^{-1}$ between locations.

The lowest period we can resolve across our array is 8 seconds (Figure 5), which reaches a peak sensitively around 7-9 km (Figure S2) depending on the 1-D shear-wave velocity structure. Comparing the phase velocity map at the 8 second period with known
geologic features and morphotectonic provinces is one way to appraise the robustness of our results both for 8 seconds and phase velocity measurements made for longer periods. In our 2-D phase velocity results at 8 seconds, we observe two distinct zones of high \((\geq 3.1 \text{ km s}^{-1})\) phase velocity and two distinct zones of low \((\leq 2.9 \text{ km s}^{-1})\) phase velocity where we have good resolution \(< 150 \text{ km}\). The northern high velocity \((3.0 \text{ km s}^{-1})\) zone is a narrow band with pockets of higher velocity \((3.2 \text{ km s}^{-1})\) and tracks the Eastern Cordillera well before merging into the Western Cordillera north of the Altiplano proper. The coastal high velocity is a broad high \((3.4 \text{ km s}^{-1})\) with slightly lower velocities \((3.2 \text{ km s}^{-1})\) in the coastal area of the Bolivian Orocline. The northern most zone of low velocity \((\leq 2.8 \text{ km s}^{-1})\) correlates well with the Subandes and has an area of slightly higher velocity \((2.9 \text{ km s}^{-1})\) along the border between Bolivia and Peru. The central low velocity represents about 50\% of our study area and the low velocity outlines an area defined by the Altiplano-Puna and has at least 3 pockets of lower velocity \((\leq 2.5 \text{ km s}^{-1})\).

Phase velocity measurements at 20 seconds are the best constrained (Table S1) and also show the largest range \((2.4 – 3.9 \text{ km s}^{-1})\) of velocities for any one period (Figure 6). The forearc is still the fastest \((\geq 3.4 \text{ km s}^{-1})\) feature observed at 20 seconds and forms the western boundary of a low velocity zone associated with the core of the Andean plateau with two distinct bodies of lower velocity \((\leq 2.5 \text{ km s}^{-1})\). The most prominent feature at this depth slice is the anonymously low velocity zone \((\approx 2.4 \text{ km s}^{-1})\) that corresponds to the location of the APMB imaged by previous seismic studies [Chmielowski et al., 1999; Zandt et al., 2003].
Our preferred shear-wave inversion starting model parameters include a constant shear-wave velocity of 4.6 km s\(^{-1}\), a Q value of 200, a Vp/Vs ratio of 1.75, and 1 km layer spacing in the top 75 km of the model. A more detailed explanation of the selection of our preferred starting model parameters is included in the supplementary section entitled “Selection of Parameters”.

4.2.1 Crustal Thickness

Although surface waves at the periods we measure are not sensitive to sharp velocities contrasts, such as the Moho, we use the absolute shear velocity variations across our study area to infer first-order variations in the thickness of the crust below the Central Andes. Figure 7 shows the results of five cross-sections though our study area where we have chosen the color palate to saturate to white when velocities exceed 4.5 km s\(^{-1}\), shear velocities associated with the upper mantle (equivalent to Vp = 7.8 km s\(^{-1}\) for a Poisson solid). Defining the color palette in this manner allows for a quick first-order evaluation of the thickness of the crust. A prominent feature in our results is the CAP, which possesses the slowest and thickest crust of the Andes (Moho depth ~60 to >70 km). The 4 km s\(^{-1}\) (Vp = 6.9 km s\(^{-1}\)) contour lies at a depth of ~55 km beneath the eastern CAP, shallowing to ~45 km beneath the Eastern Cordillera. Moving west into the high elevations of the Western Cordillera and dropping elevation into the Precordillera and forearc, we observe a rapid increase in crustal velocity reflected by the shallowing of the 4.0 km s\(^{-1}\) contour to depths of 20-30 km. The crustal thickness is quite variable, locally correlating with the location of the top of the subducting Nazca Plate (red line). On the
opposite side a gradual eastward thinning of the crust is observed below the eastern flank of the Plateau transitioning into the Subandes and the foreland.

4.2.2 Forearc

At the upper limit of our resolution, a striking correlation is observed in our 5 km depth slice with the morphotectonic provinces (Figure 8a). The onshore forearc is characterized by fast velocities (~3.5 km s\(^{-1}\)) that follow the Coastal Cordillera, the mapped location of a Paleozoic batholith [Mamani et al., 2010]. Near the coastal section of the axis of the Bolivian Orocline (~18° S), the onshore section of the forearc is furthest from the trench and the Coastal Cordillera narrows and is no longer mapped onshore. Our results appear to capture this transition from high (~3.5 km s\(^{-1}\)) velocities to low (<3.0 km s\(^{-1}\)) and correlate well with the location of the southern Moquegua Basin [Rousse et al., 2005] suggesting the southern Moquegua Basin is a deeper sub-basin. Where the forearc is widest (between 22°S and 24°S), we observe low velocities similar to the velocities of the southern Moquegua Basin associated with the Atacama and Calama Basins. At 15 km depth (Figure 8b), the forearc is as fast as any other area at this depth and at 35 km (Figure 8c) it is the fastest morphotectonic province with localized velocities of 4.5 km s\(^{-1}\). At 50 km depth (Figure 8d) the forearc is characterized by velocities that range from 3.75 km s\(^{-1}\) to >4.5 km s\(^{-1}\). Assuming that the 4.5 km s\(^{-1}\) value represents mantle material, this observation suggests a highly variable but generally large crustal thickness, given the relatively low elevations of the forearc. However, it is well known that forearc regions are often underlain by hydrated mantle that manifests itself with shear velocities that are low enough to appear crustal [Bostock et al., 2002]. Previous seismic studies in the
Chilean forearc have interpreted the apparent large crustal thickness as an indicator of hydrated mantle [Schmitz et al., 1999] and our results are generally supportive of this idea. This topic is discussed further later in this paper.

4.2.3 Western Cordillera

The Western Cordillera is of intermediate velocity (3.0-3.5 km s\(^{-1}\)) at the 5 km depth slice (Figure 8a). At 15 km depth (Figure 8b), an almost continuous along strike low velocity body (<3.25 km s\(^{-1}\)) underlies the Western Cordillera north of 18°S. South of this latitude the correlation between the Western Cordillera and the low-velocity body is more complicated. Between 18°S and 21°S the location of the low velocity body shifts eastward beneath the western edge of the Altiplano, and between 21°S and 24°S, the body widens and underlies Western Cordillera and the northern Puna. This low velocity body appears to terminate against the northern edge of the Salar de Atacama near 24°S. Despite this complex relationship between the Western Cordillera morphotectonic province and the low velocity body, the active arc as marked by Holocene volcanoes is consistently located along the western edge of the body, until it terminates near 24°S. By 35 km depth (Figure 8c) the Western Cordillera is no longer distinguishable by a distinctive velocity, but rather appears as a gradient zone between the higher velocities in the forearc and the lower velocities in the interior of the Altiplano-Puna.

4.2.4 Altiplano-Puna

Although most of the Andean Plateau velocities at 5 km depth (Figure 8a) are below 3.25 km s\(^{-1}\), the Altiplano-Puna is distinctly different from the rest of the Plateau
with large sections of the Altiplano-Puna below 3.0 km s\(^{-1}\) and locally below 2.75 km s\(^{-1}\).

In the Altiplano between 17\(^\circ\)S and 18\(^\circ\)S, a low velocity body (2.75 km s\(^{-1}\)) at 5 km depth corresponds with the location of the Corque–Corocoro structure, a synclinal fold-and-thrust zone that deforms and preserves a thick sections of Eocene or Oligocene to Pliocene red-bed sequences [Lamb, 2011]. By the 15 km depth slice (Figure 8b) the Altiplano-Puna is no longer a distinct feature. However, at the 35 km and 50 km depth slices (Figure 8c, 8d) the CAP, which continues north of the Altiplano, is a continuous zone of low-velocity flanked by higher velocities extending the length of the study area.

4.2.5 Eastern Cordillera

The Eastern Cordillera is perhaps one of the best-resolved regions in the 5 km depth slice (Figure 8a). Flanked to the northeast by the thin-skinned fold-and-thrust belt of the Subandes with thick piggyback basins and to the southwest by the Altiplano Basin, the basement-cored fold-and-thrust style of the Eastern Cordillera has formed a very different shallow subsurface. The low velocities (<3.0 km s\(^{-1}\)) of the Altiplano and Subandean basins stand in sharp contrast to the much higher velocities (>3.25 km s\(^{-1}\)) of the Eastern Cordillera where basement rocks outcrop or are located near the surface. A few additional pockets of even higher velocity (>3.5 km s\(^{-1}\)) are distributed along the Cordillera Real segment of the Eastern Cordillera and correlate in space with the location of Triassic and Tertiary granitoid plutons (Figure 2) [Gillis et al., 2006; Benjamin et al., 1987]. In the 15 km depth slice (Figure 8b), the distinction between the Subandes and the Eastern Cordillera is not as apparent as in the shallower results but the Eastern Cordillera maintains a generally higher velocity that defines the eastern limit of the Altiplano and
the western edge of the Subandes. At greater depths (Figure 8c, 8d) this region becomes part of the lateral gradient associated with the westward crustal thickening into the Andean Plateau.

4.2.6 Subandes

The Subdandes accommodate the transition from the high topography of the CAP to the lowland elevation of the foreland basin through a series of active thin-skinned fold-and-thrust type structures. We have less coverage over this section of the Andes but where we can resolve the velocity structure, we observe the lowest velocities (<3.0 km s\(^{-1}\)) at the 5 km depth slice in the Subandes (Figure 8a). The lowest velocities correlate with the Alto Beni Basin, a large piggyback basin [Baby et al., 1995]. Shear-wave velocities within the Subandes are mostly slower than in the Eastern Cordillera at 15 km (Figure 8b) but by 35 km (Figure 8c) it is faster than the Eastern Cordillera. At 50 km depth (Figure 8d), the Subandes are as fast as the fore-arc with pockets of mantle velocity (~4.5 km s\(^{-1}\)) likely reflecting the transition to a thinner crust in the foreland.

4.2.7 APVC

The Altiplano terminates near 21.5°S and is replaced by the high standing Puna province. The Altiplano-Puna Volcanic Complex is a late Cenozoic large-volume silicic volcanic center located on the northern end of the Puna [de Silva, 1989]. The APVC and the crustal body of partial melt that supplies it (APMB) provide an additional test on the ability of our results to resolve crustal structure. Previous work has characterized the location of a zone of partial melt or mush below the APVC that is a regionally extensive,
sill-like “magma body” centered at 15-20 km depth (Figure 8b) [Chmielowski et al., 1999; Zandt et al., 2003]. Our results at the 15 km depth slice show a regionally extensive low-velocity body (<2.75 km s\(^{-1}\)) in the area of the APMB and our results in the cross-section (Figure 7d) show a sill-like body centered at a depth of 15 km. Although smaller in volume, the Los Frailes Volcanic Complex (LFVC) at 19.5°S is a similar type ignimbrite flare-up where our results at the 15 km depth slice also show a comparable low-velocity zone below the LFVC. It is interesting to note that at greater depths (~50 km), the regions under the large silicic volcanic fields are characterized by higher velocities than surrounding areas of the Altiplano-Puna. These higher velocities may be the signature of a mafic residue zone and will be discussed in more detail later.

5. Discussion

Two first-order crustal-scale features observed in the ANT results are the thick high-velocity crust in the forearc, and the presence of an extensive mid-crustal low-velocity body under the western margin of the Andean Plateau. Both features have been recognized in earlier studies, but the larger-scale, uniform-resolution coverage of the ANT study provides a more regional context for their interpretation. In this section, the ANT results are compared to previous seismic studies where there is overlap, and compared to other geophysical and geological studies to better constrain the interpretations of these features.

5.1 Central Andes Forearc Crustal Structure
The Chilean forearc between 21°S and 25°S has been the focus of many seismic studies, notably as part of a series of German-Chilean collaborative projects [Oncken et al., 2006]. Both active source and passive source studies are summarized and a detailed geologic interpretation is presented in Schmitz et al. [1999]. More recent studies focused in the back-arc region are reviewed later.

Using P and S wave arrival times from local earthquakes, Graeber and Asch [1999] inverted for a three-dimensional Vp and Vp/Vs velocity model. Their Vp structure along an east-west cross-section profile of the forearc at 23.25°S is compared to the ANT Vs model at the same latitude in Figure 9. Although the ratio between compressional-waves and shear-waves can vary as a function of composition, presence of fluids, and anisotropic effects, Graeber and Asch’s Vp/Vs model show that west of the active arc, the Vp/Vs variation is principally a function of depth, <1.74 above 40 km and >1.74 below; therefore the lateral structures between the Vp and Vs models should be similar. Comparison of the two models shown in Figure 9 show the first order structures are consistent and there is agreement in even some smaller, subtler features. For example, centered around 67.9°W in both cross sections, there is a near-vertical transition in the mid-crust from slow to fast velocities passing into the forearc. A subtler feature is the eastward inclination of the isovelocity contours west of 69°W. A detail that helps establish the depth of the ANT resolution is the agreement in the 60 km depth of the continental Moho between 69°W and 69.5°W. Further west, both models also show the westward shallowing top of the subducting oceanic plate. This comparison supports our contention that our results are robust to a depth of ~70 km with exceptional detail resolvable in the middle to lower crust.
The geologic interpretation of the detailed seismic studies in the Chilean forearc is presented in Schmitz et al. [1999] and Giese et al. [1999], and only briefly summarized here. The prominent seismic discontinuity in the forearc at a depth of 35 to 45 km observed in both the active source and passive source models (where Vp reaches ~ 7 km s\(^{-1}\)) is interpreted as the base of the Mesozoic lower crust, and its base as a “blurred” continental paleo-Moho. The depth interval between approximately 45 and 65 km is characterized by Vp between approximately 7 and 7.5 km s\(^{-1}\) and is interpreted as serpentinized upper mantle and the geophysical Moho at ~65 km is suggested to represent the base of the stability field of serpentine and a transition to amphibole bearing peridotites. This interpretation explains the presence of unusually thick crust, at least in a geophysical sense, in the absence of significant tectonic shortening [Giese et al., 1999].

The ANT results are consistent with this interpretation. They also show that this thick geophysical crust extends the length of the forearc in the study area, from 15°S to 24°S. The ANT results further show that the uniformity of this “thick forearc crust” changes at the Arica bend at 18°S, where to the north the forearc crust is relatively uniformly thick, and to the south the forearc crust shows considerable variation in thickness between as little as 35 km and >70 km (Figures 8c and 8d). The possible interpretations of this observation are beyond the scope of this paper and will be addressed in future publications.

5.2 Central Andes Arc and Backarc

The other major result from the ANT study is the observation of a major mid-crustal low-velocity body along the western margin of the Andean Plateau. It has been
known for some time that the CAP is supported by a thick, low average velocity crust [e.g., Beck and Zandt, 2002] with a pervasive mid-crustal low-velocity layer, often called the Altiplano Low-Velocity Zone (ALVZ) [Yuan et al., 2002]. The low average velocity reflects a predominantly felsic composition [Beck and Zandt, 2002; Lucassen et al., 2001] and the low-velocity layer has been proposed to be a zone of crustal flow [Husson and Sempere, 2003]. We first review the existing geophysical studies from the arc and backarc regions of the Central Andes then we present a new interpretation of the mid-crustal low-velocity zone.

Numerous geophysical studies have imaged the subsurface structure of the arc and backarc regions of the Central Andes. These studies utilize a number of different techniques that include: local earthquake tomography [Schurr et al., 2003, Koulakov et al., 2006; Graeber and Asch, 1999; Myers et al., 1998; Dorbath and Granet, 1996], teleseismic earthquake tomography [Heit et al., 2008; Dorbath et al., 1993], surface wave studies [Baumont et al., 2002; Zandt et al., 2003; Vdovin et al., 1999], attenuation tomography [Haberland and Rietbrock, 2001; Schurr et al., 2003], seismic refraction and reflection surveys [Wigger et al., 1994; Schmitz et al., 1999; Oncken et al., 2003], converted wave studies [Sodoudi et al., 2011; Beck and Zandt, 2002; Yuan et al., 2002; Chmielewski et al., 1999; McGlashan et al, 2008], gravity studies [Kono et al., 1989; Whittman, 1999; Gotze and Krause, 2002; Tassara et al., 2007; Chase et al., 2009], and magnetotelluric studies [Brasse and Eydam, 2008; Brasse et al., 2002]. Comprehensive reviews have been published over the years by Allmendinger, Asch, Kay, Oncken, and Tassara [Allmendinger et al., 1997; Asch et al., 2006; Kay and Coira, 2009; Oncken et
al., 2003; Tassara and Echaurren, 2012]. A review of all these studies is beyond the scope of this paper but we review the most relevant results.

Using earthquake generated surface waves, Baumont et al. [2002] estimated the shear-wave velocity structure in the Central Andes. With large station spacing and a limited number of stations, the resolution available was about a degree or larger. A 1-D comparison of our ANT results with the surface wave tomography results is shown in Figure 10. As a result of the limited station coverage and large averaging distance of the earthquake study, a precise comparison is not possible, however, good agreement is observed between larger first-order scale structures. For example, at the LFVC location, a zone of mid-crustal (~20 km) low-velocity (~3.0 km s\(^{-1}\)) is well resolved by both studies. In the same area below the low velocity feature we observe a rapid increase in velocity with depth to 4.0 km s\(^{-1}\) at 50 km depth. This low over high velocity anomaly is significant because it is located below the LFVC, an area of 0-10 Ma silicic volcanic activity [de Silva and Francis, 1991].

The Altiplano-Puna Volcanic Complex of the Central Andes is the best-studied large silicic volcanic field associated with an active subduction margin. As a result, multiple geophysical techniques have imaged a low velocity, high conductivity crustal “magma body” centered on a depth of 15 km [Chmielowski et al., 1999; Zandt et al., 2003; Pritchard and Simons, 2004; Brasse et al., 2002], which corresponds to our low-velocity body (LVB) shown in Figure 7d and 8b. Thus, our results at 15 km depth with V\(_{sv}\) <3.0 km s\(^{-1}\) most likely represent a magma “mush” zone with 15-20% partial melt [Schilling and Partzsch, 2001; Schilling et al., 2006]. Magnetotelluric results from further north (~18°S) do not resolve a mid-crustal high conductivity zone [Brasse and Eydam,
2008; Schilling et al., 2006], which is consistent with our results at this location. Near 18°S, the 3.25 km s⁻¹ LVB narrows and pinches out, dividing the LVB into a northern and southern section suggesting the LVB does not contain interconnected zones of partial melt. However, velocities observed at this pressure and temperature regime are too low for an isotropic (melt free) granitic composition. We address this discrepancy by discussing the possible effects or radial anisotropy on our results in the following section.

5.3 The Central Andes Neogene Batholith

It has been proposed that large silicic volcanic activity (ignimbrite eruptions) is the surface expression of batholith formation at depth [de Silva and Gosnold, 2007; Lipman, 1984, 2000; Bachmann et al., 2007]. Although a complete description of the complex evolution of volcanic systems is beyond the scope of this paper, the formation of a batholith of intermediate to felsic composition implies the existence of a dense mafic root (Ducea, 2001; Saleeby et al., 2003; Lee et al., 2006; Tassara, 2006). Through magmatic fractionation and segregation processes, denser residuum preferentially reside at depth while lighter felsic melts rise producing a seismic velocity structure of low over high. We explore the relationship between batholith formation and the shear-wave velocity structure using the APVC as a guide because more geologic studies have been done in this area, and we have good resolution in the area of the APVC.

Working under the hypothesis that voluminous ignimbrites are the surface expression of batholith formation at depth, de Silva and Gosnold [2007] incorporate geochemical, ignimbrite eruptive histories, gravity data, and seismic data to elucidate the linkage between the spatiotemporal development of ignimbrite flare-ups and episodic
batholith formation at depth. Their analysis suggests a thermally induced positive feedback magmatic system that promotes the accumulation of dacite magma bodies that evolves with time suggesting the presence of a large batholith of intermediate to felsic composition beneath the APVC. We adopt a similar multi-disciplinary approach as de Silva and Gosnold [2007] by combining out ANT results with isostatic residual gravity data and mapped Neogene ignimbrite centers to search for evidence of a large batholith as predicted by de Silva and Gosnold [2007].

Starting with the well-defined location of the APMB at approximately 15 km depth [Zandt et al., 2003; Leidig and Zandt, 2003], we overlay the horizontal extent of the APMB onto our 15 km depth slice results and observe a good correlation between the outline of the APMB and the 3.0-3.25 km s\(^{-1}\) velocity contour (Figure 8b). Although not as well imaged as the APMB, the LFVC to the northeast has a low velocity body at centered around 15 km depth which if defined by the 3.0-3.25 km s\(^{-1}\) velocity contour outlines the approximate surface expression of the LFVC. In addition to both the APVC and LFVC being associated with young (<12 Ma) ignimbrite eruptive centers, negative isostatic residual gravity anomalies [Whitman, 1999] coincide with the locations of both volcanic complexes providing further evidence for a low density body in the crust. Figure 11 shows the correlation between negative isostatic residual gravity anomalies, the location of Neogene ignimbrite eruption centers, and the 3.25 km s\(^{-1}\) velocity contour from our 15 km depth slice as a proxy for the approximate location of a Neogene batholith that extends along the length of the CAP. This interpretation is further supported by the distribution of active arc Holocene volcanoes along the western edge of
the proposed batholith, and the correlation with the locations of back arc volcanic activity.

One problematic issue is the fact that the Vsv <3.25 km s\(^{-1}\) is too low for an isotropic granitic composition even at high temperature [Christensen 1996; Sobolev and Babeyko, 1994]. A simple explanation for this could be the presence of small amounts of partial melt, but the absence of a high conductivity anomaly as discussed earlier precludes this as a possible explanation along the entire length of the low-velocity body. Another possibility is the presence of radial anisotropy where vertically polarized shear-wave velocities (Vsv) are slower than horizontally polarized shear-wave velocities (Vsh). This possibility is suggested by previous work in Tibet where earthquake generated Rayleigh and Love surface wave dispersion curves require a mid-crustal zone of radial anisotropy with Vsh amplitudes between 10% and 20% greater than Vsv to match observed surface wave dispersion curves [Shapiro et al., 2004; Duret et al., 2010; Yang et al., 2012]. A similar mid-crustal zone of radial anisotropy in the Central Andes would increase our effective isotropic velocities by 5% to 10%. Uniformly increasing the results in 15 km depth slice by 10% will not change the shape of the proposed batholith, but the resulting higher velocity contour of ~3.5 km s\(^{-1}\) is a better fit for a granitoid composition. Velocities below 3.0 km s\(^{-1}\) such as that observed beneath the APVC and LFVC, likely represents the presence of partial melt [Schilling et al., 2006].

An important question is what is the petrophysical explanation for this hypothesized anisotropy? In the case of Tibet, the crustal anisotropy was attributed to the preferred orientation of mica resulting from crustal thinning [Shapiro, et al., 2004]. In our case, there is no indication of crustal thinning, rather the western flank of the CAP
appears to be undergoing Late Cenozoic surface uplift as a crustal-scale monoclonal fold
of a horizontal foliation related to the crystallization of the batholith. Sawyer [1994]
emphasized that crustal melting, segregation, and crystallization takes place in a dynamic
environment in which shearing, melting, melt migration, and chemical processes interact
over different time scales. Outcrop studies of melt formation and segregation in
migmatitic terranes reveals a close correlation between melting locations and foliation
[Sawyer, 2001], suggesting that even after crystallization, a remnant anisotropy may
persist within the former melt zone. Observations of crustal seismic anisotropy in the
APVC crust [Leidig and Zandt, 2003] supports this idea, although the receiver function
method used in that study is sensitive to only azimuthal anisotropy. The presence of
azimuthal anisotropy does not preclude the presence of radial anisotropy. Future work is
planned that will directly investigate the effects of radial anisotropy.

5.4 Implications for Plateau Uplift
If our suggestion of a major Neogene to modern batholith associated with the
ignimbrites and arc volcanism is correct, then it implies a significant Neogene component
of magmatic addition along the western portion of the CAP [James, 1971; Kono et al.,
1989; Schmitz, 1994]. Jordan et al. [2010] recently reviewed the evidence for the Late
Cenozoic uplift of the western margin of CAP and summarized three competing tectonic
causes: (1) underthrusting of the Brazilian foreland crust, (2) delamination of an eclogitic
lower crust, and (3) lower crustal flow. They conclude, “Work remains before the
magmatic controls, lithospheric thinning controls (delamination), and crustal shortening
controls are fully disentangled and documented.” The new images of the Neogene batholith presented here, in conjunction with petrological and geochemical constraints, provide an opportunity to better quantify the role of magmatic addition in this debate.

6. Conclusions

Our first order results confirm that the high elevations of the CAP are supported by a thick (~70 km) low-velocity (and presumably low-density) crust with localized but regionally extensive mid-crustal low-velocity zones. Working under the hypothesis that voluminous ignimbrites are the surface expression of batholith formation at depth as exemplified by the APVC [de Silva and Gosnold, 2007], we combine our results with the locations of known Neogene ignimbrite eruptive centers and isostatic residual gravity and conclude the 3.25 km s$^{-1}$ contour at 15 km depth generally outlines the extent of a Neogene batholith with isolated pockets of partial melt (Figure 11). This newly outlined batholith generally correlates with the Western Cordillera, but locally extends into the western Altiplano, Eastern Cordillera, and Puna, and more surprisingly, locally extends into the forearc. The modern volcanic arc is located predominantly on the western margin of the batholith, suggesting some density control on the localization of melt migration paths. The batholith is relatively narrow (~100 km) between 16°S and 19°S, but splits into two branches farther north in Peru and in southern Bolivia it occupies a significantly greater width of the orogen. The young batholith mostly occupies regions of high elevation, suggesting the generation of low-density granitoids and the prevailing high temperatures must counteract the densifying effects of the formation of mafic residues at depth.
Although the prevailing consensus is that the Central Andes are built predominantly by crustal thickening due to compressional shortening, the identification of a large volume Neogene batholith recalls the possible important role of magmatic addition, especially in Peru and the Puna where shortening estimates are relatively smaller than in the central Bolivian segment. Additionally, recent multi-disciplinary studies documenting late Cenozoic uplift on the western margin of the central Andes, with relatively little documented shortening, also suggest magmatic addition as a possible important causative uplift mechanism. Future interdisciplinary studies should reexamine the potential role of magmatic addition and its interactions with crustal scale structures in the uplift history of the Andes.

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8. References


Ducea, M., 2001. The California Arc: thick granitic batholiths, eclogitic residues, lithospheric-scale thrusting, and magmatic flareups. GSA Today 11 (11), 4-10


Herrmann, R. B., 1987. in Computer Programs in Seismology, edited, St. Louis University, St. Louis, Mo.


9. Supplementary Material
In the supplementary material we present some additional analysis, figures, and sensitivity tests associated with our ANT study of the Central Andes.

**Selection Criteria for Frequency Content**

The selection criterion for rejecting interstation paths is described in the main text and ultimately discards 37-68% of the possible interstation paths. Table S1 shows the number of interstation paths that remain after quality control measures have been applied as a scalar value and as a percentage of the total possible interstation paths for each period between 8 and 50 seconds. The number of good quality interstation path measurements peaks around 20 seconds and quickly tapers above and below that period defining the upper and lower limits of the periods we can use in the shear-wave inversion. Below 8 seconds, very few (<100) interstation paths remain and above 40 seconds good resolution is confined to the northern section of our study area. We define areas of good resolution (“bright”) in our shear-wave inversion results by requiring every period used to construct the phase velocity profile at any given grid point to possess good resolution (<150 km), whereas areas of fair resolution (“dull”) in our shear-wave inversion results require every period used to construct the phase velocity profile at any given grid point to possess fair resolution (150-800 km). In choosing 40 seconds as the longest period to use in our shear-wave inversion as opposed to 50 seconds, we are trading depth resolution for horizontal resolution. Future work will include using ballistic surface waves that can provide good resolution below 100 seconds (~200 km depth) but for the purpose of this paper we focus on the upper lithosphere using 40 seconds as our lower limit.
Selection of Parameters

We present a more detailed consideration of how we selected our preferred regularization parameters for our inversions. Figure 3 in the main text shows our shear-wave results have some dependence on the regularization parameters used in the 2-D phase velocity inversion and shear-wave inversion-starting model. As mentioned in the main text, we use conservative values for the regularization parameters $\beta = 100$ and $\sigma = 100$ km. $\beta$ is a weighting factor associated with path density affecting the peripheral area of our shear-wave results where path density and our resolution quickly fade from good (“bright”) to fair (“fair”). We address the ambiguity of selecting $\beta = 100$ by limiting our interpretation to areas with good (“bright”) resolution. $\sigma$ defines the length over which 2-D phase velocity results are smoothed. Defining $\sigma = 100$ km reduces the appearance of speckling and helps to stabilize amplitude oscillations across our entire study area acting as a spatial smoothing parameter. We observe the largest effects in our 2-D phase velocity results and our shear-wave velocity results as a function of the regularization parameter $\alpha$ and as such the selection of this value merits more attention.

In the main text we rigorously define the normalized roughness ($\xi'$) and reserve a detailed description of how we use it to arrive at $\alpha = 200$ for this section. Figure S1a shows the normalized roughness plotted against the damping value used at every point in the array with at least fair (150-800 km) resolution. In choosing the best damping value for our study area, it should be noted that we are attempting to balance limiting the introduction of artificial velocity artifacts with aggressive smoothing whereby real velocity contrasts in the Earth are averaged (under-damped vs. over-damped). Figure S1b shows a “well behaved” dispersion curve (Peruvian Eastern Cordillera referenced in the
where the selection of the damping value does not change the velocity as a function of period as much as the “poorly behaved” dispersion curve (Chilean Forearc referenced in the text) shown in Figure S1c. In the poorly behaved dispersion curve, the lowest damping value of 50 is clearly too erratic to be representative of the velocity structure of the Earth and can be rejected as under-damped. However, there is a negative velocity swing around 30 seconds that every damping value below 350 produces. This suggests that damping values above 300 are over-damped for this specific location. Preforming this type of point-by-point location analysis is impractical as there are over 5,000 grid locations in our study area with good resolution alone, even more with fair resolution. Here is where the value of our normalized roughness metric can be utilized.

Highlighted in Figure S1a are the normalized roughness curves for both the well-behaved Peruvian Eastern Cordillera location (yellow) and the poorly behaved Chilean Forearc location (red). We can determine everything discussed above by looking at the two curves. The yellow curve associated with the well-behaved location has low normalized roughness values and is flat across all damping values indicating that increasing the damping value will have less of an effect on the roughness of the dispersion curve. By contrast, the red curve associated with the poorly-behaved location has high normalized roughness values at lower damping values and low normalized roughness values at higher damping values that flatten around the damping value of 200. This suggests that increasing the damping value will have a greater effect on the dispersion curve up to the value of 200 where increasing the damping value above 300 will not effect the roughness of the dispersion curve. For most locations, not all, the
damping value of 200 is where the gradient of the normalized roughness curves begins to flatten out and forms the basis of selecting 200 as our preferred damping value.

**Surface Wave Sensitivity to Depth**

As shown in Figure S2, surface waves are dispersive meaning different periods are sensitive to different depths. In general, with surface waves the peak sensitivity depth and the thickness of sensitivity increases as the length of the period of the surface wave increases. It is this property of surface waves that makes them insensitive to sharp velocity contrasts.

**Resolution Definition**

In Figures S3-S12 we show phase velocity maps at 8, 10, 12, 14, 16, 20, 25, 30, 35, and 40 seconds with the associated resolution, ray paths, and path density plots. In addition to the phase velocity results, the resolution plots are significant as we use them to define the horizontal extent of our shear-wave velocity results. As discussed in the above supplementary material, the number of quality interstation paths peaks around 20 seconds and diminishes above and below that period which affects the resolution at each period differently. It should be noted that the resolution quickly transitions from good (>150 km) to fair (150-800 km) near the peripheral of our array coverage for any given period used and that the horizontal extent of the resolution is different for each period. We adopt a simple systematic approach to grading the robustness of our shear-wave results primarily defined by the extent of the 2-D phase velocity results. We require the same 10 periods be used for every grid point in the shear-wave inversion. Areas of shear-
wave results presented as “bright” are the result of inverting for 10 phase velocity periods that all have good (>150 km) resolution. Shear-wave results presented as “dull” are the result of inverting for 10 phase velocity periods that have between 1 and 10 periods of fair resolution (150-800 km). Implicit in this definition of fair resolution is that some percentage of the periods used will have good resolution and some percentage will have fair resolution resulting in a zone (“dull”) of shear-wave results with varying measures of robustness (less than good but better than bad, or fair). This zone can be thought of as a transition zone that incorporates all of the available data gradually transitioning from areas of good resolution to areas with no resolution.

Sensitivity of Input Starting Parameters

In Figures S13-S22 we show the individual effects of systematically varying each input starting model parameter and damping value input used to construct the uncertainty envelopes for each of the ten representative points shown in Figure 3 of the main text. We cannot present the individual effects all 216 possible permutations but rather show the effects of varying each individual parameter against our preferred damping and starting model values. We have already addressed why we prefer a damping value of 200 in the above text and here we establish the merits of selecting our preferred starting model values. We do not imply that we have 1 km horizontal resolution by parameterizing the top 75 km of our starting model with 1 km think layers but rather it results in a smooth model and we observe little difference in the final model results obtained from 5 km thick layers (see (d) of Figures S13-S22). Our shear-wave inversion results solve for Vs with a fixed Vp/Vs ratio that is held constant and we use a 1.75 ratio,
close to the Poisson value. We invert for a Vp/Vs ratio of 1.70 and 1.80 and observe very little difference in our results (see (c) of Figures S13-S22). Previous work in this part of the Central Andes suggests that an average Q value for the Altiplano is around 200 with other surrounding areas having a higher Q value [Baumont et al., 1999]. We invert for Q values of 50, 200, and 500 observing little change between 200 and 500 (see (b) of Figures S13-S22). We prefer a Q value of 200 in part because previous work has established it for the Altiplano and also higher Q values do not appreciably affect our shear-wave inversion results. The final value of our stating model that we systematically test is the starting shear-wave velocity. Our primary reason in choosing a constant velocity starting model is the variable and sometimes uncertain Moho depths across our study area. In future work we will incorporate a detailed Moho in such a study. It should be noted that introducing a Moho in the starting model affects only the results within ~10 km of the Moho depth which previous work has shown is between 70 and 40 km across our entire study area. Our choice to not include an approximate Moho depth in our starting model should not influence our mid-to-upper-crustal results and therefore we conclude our interpretations focused on upper crustal structures are robust. We prefer a starting Vs value of 4.6 km s\(^{-1}\) and test the effects of more realistic crustal values (3.1, 3.6, and 4.1 km s\(^{-1}\)) on our inversion results (see (a) of Figures S13-S22) and observe little (>0.05 1 km s\(^{-1}\)) variation in the upper 50 km of the lithosphere. We prefer a starting value of 4.6 km s\(^{-1}\) because the only value in the inversion that we can control is the half-space value, meaning whatever velocity we prescribe in the starting model for the half-space will be the final velocity in our results. In our starting model, the half-space is at depth of 315 km which corresponds to a shear-wave velocity of ~4.6 km s\(^{-1}\) from
global models [Kennett and Engdahl, 1991]. Again, since surface waves are not sensitive
to sharp velocity contrasts and at depths where our data do not drive the inversion (>70
km), the results gradually trend toward the half space value. Thus at depths where our
data stops driving the shear-wave inversion results, our starting model produces a gradual
velocity increase in our results approximating the global velocity model for that depth
range (~70-315 km).

10. Figure Captions

Table 1. Table summarizing seismic networks used in this study.

Figure 1. Map of study area showing the tectonic setting of the Central Andes, locations
of seismic stations used in this study, and Holocene volcanic activity from Siebert and
Simkin [2002-]. Colored diamonds are broadband seismic deployments and the black
diamonds are permanent stations from global networks. The black line follows the three-
kilometer topography contour and defines the horizontal extent of Central Andean
Plateau (CAP).

Figure 2 Map showing simplified geology [modified from Barnes and Ehlers, 2009],
Holocene volcanic activity (black triangles) [Siebert and Simkin, 2002-], and
morphotectonic provinces: Forearc, Western Cordillera, Altiplano, Puna, Eastern
Cordillera, and Subandes [white dashed lines modified from Tassara, 2005]. Also shown
are the locations of the ten representative locations a through j referred to in the text
(white/black stars).

Figure 3. Shear-wave velocity results for each of the ten representative locations referred
to in the text and shown in Figure 2 (a through j) where 0 depth refers to sea level. Gray
shaded areas represent the uncertainty envelope as defined by the maximum and
minimum shear-wave velocity results as a function of varying layer thickness, quality
factor, Vp/Vs, starting model Vs from the shear-wave inversion, and damping value from
the 2-D phase velocity inversion. The individual effects of varying each parameter are
presented in the supplementary material for each representative point. The dashed black
line is the results from our preferred model parameters. Note the nesting of the
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driven.

Figure 4. 1-D phase velocity results for periods 8 to 50 seconds for our 10 representative
locations (a through j) on Figure 2 shown in color and grouped by morphotectonic
province. Note the variability in phase velocities at all periods. Black lines represent the shear-wave inversions best fit to each 1-D phase velocity curve.

**Figure 5.** 2-D phase velocity results for the 8-second period. (a) Smoothed phase velocity results shown with the morphotectonic provinces from Figure 2 (Forearc: FA, Western Cordillera: WC, Altiplano: AP, Puna: PU, Eastern Cordillera: EC, and Subandes: SA). Phase velocity results are masked where resolution is less than fair (>800 km) and contoured (thin black line) where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.

**Figure 6.** 2-D phase velocity results for the 20-second period. (a) Smoothed phase velocity results shown with the morphotectonic provinces from Figure 2 (Forearc: FA, Western Cordillera: WC, Altiplano: AP, Puna: PU, Eastern Cordillera: EC, and Subandes: SA). Phase velocity results are masked where resolution is less than fair (>800 km) and contoured (thin black line) where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.

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**Figure 8.** Results for the 3-D shear velocity inversion at 5, 15, 35, and 50 km depths with the morphotectonic provinces from Figure 2 shown as solid black lines (Forearc: FA, Western Cordillera: WC, Altiplano: AP, Puna: PU, Eastern Cordillera: EC, and Subandes: SA). The lateral extent of the Altiplano-Puna Magma Body is outlined (dashed yellow line) in the 15 km depth slice [Zandt et al., 2003]. Velocity results are “dull” where resolution is fair (150-800 km) and “bright” where resolution is good (<150 km). Depths are referenced to sea level and Holocene volcanic activity is shown as black triangles [Siebert and Simkin, 2002-].

**Figure 9.** Comparison of a cross-section of our shear-wave velocity results with compression-wave velocity results along 23.25°S from Graeber and Asch [1999]. Note the active volcanic arc between 67.5°W and 68°W.

**Figure 10.** Comparison of our 1-D shear-wave velocity results (yellow dashed line) at our APVC location with earthquake surface-wave 1-D shear wave velocity results (red dashed line) from Zandt et al. [2003] (a), at our Central Altiplano location with earthquake surface-wave 1-D shear wave velocity results (red dashed line) from Baumont et al. [2002] (b), and at a Los Frailes location (19.5°S, 66.6°W) with earthquake surface-wave 1-D shear wave velocity results (red dashed line) also from Baumont et al. [2002]
Note the location we refer to as the Central Altiplano corresponds better with Baumont’s Northern Altiplano location. The earthquake surface wave results average the shear-wave velocity over a much larger area than our results.

Figure 11. Summary figure combining the results of our 15 km depth slice of shear wave velocity with the locations of known Neogene ignimbrite eruptive clusters, shown as black areas [Salisbury et al., 2011], Neogene ignimbrite eruptive centers in southern Peru [Echavarria et al., 2006] and the APVC [Salisbury et al., 2011] shown as white areas with isostatic residual gravity from Whitman [1999]. The 3.25 km s$^{-1}$ velocity contour (thick black line from the 15 km depth slice) correlates well with areas of low isostatic residual gravity (thick white line) and the locations of known Neogene ignimbrite eruptive centers. We interpret this correlation as compelling evidence for the 3.25 km s$^{-1}$ contour outlining the general extent of a Neogene batholith.

Table S1. Table summarizing the number of high quality interstation paths used and the percentage of the maximum interstation paths for each period between 8 and 50 seconds.

Figure S1. The normalized roughness plotted against the damping value used at every point in the array with at least fair (150-800 km) resolution. (a) The normalized roughness for the Peruvian Eastern Cordillera (yellow line) and Chilean Forearc (red line) shown for reference. (b) Shows a “well behaved” dispersion curve (Peruvian Eastern Cordillera referenced in the text) where the selection of the damping value does not change the velocity as a function of period as much as the “poorly behaved” dispersion curve (Chilean Forearc referenced in the text) shown in (c).

Figure S2. Sensitivity kernels for each of the 10 periods used in the shear-wave inversion calculated from our final shear-wave velocity model at our Central Altiplano location shown in Figure 2 of the main text.

Figure S3. 2-D phase velocity results for the 8-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.

Figure S4. 2-D phase velocity results for the 10-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.

Figure S5. 2-D phase velocity results for the 12-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is
defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.

Figure S6. 2-D phase velocity results for the 14-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.

Figure S7. 2-D phase velocity results for the 16-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.

Figure S8. 2-D phase velocity results for the 20-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.

Figure S9. 2-D phase velocity results for the 25-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.

Figure S10. 2-D phase velocity results for the 30-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.

Figure S11. 2-D phase velocity results for the 35-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.
**Figure S12.** 2-D phase velocity results for the 40-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.

**Figure S13.** Analysis of systematically varying each parameter at the Subandean location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s⁻¹), (b) Q (50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.

**Figure S14.** Analysis of systematically varying each parameter at the Bolivian Eastern Cordillera location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s⁻¹), (b) Q (50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.

**Figure S15.** Analysis of systematically varying each parameter at the Peruvian Eastern Cordillera location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s⁻¹), (b) Q (50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.

**Figure S16.** Analysis of systematically varying each parameter at the Altiplano-Puna Volcanic Complex location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s⁻¹), (b) Q (50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.

**Figure S17.** Analysis of systematically varying each parameter at the Central Antiplano location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s⁻¹), (b) Q (50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.

**Figure S18.** Analysis of systematically varying each parameter at the Northern Arc location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s⁻¹), (b) Q (50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.

**Figure S19.** Analysis of systematically varying each parameter at the Peruvian Arc location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s⁻¹), (b) Q...
(50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.

**Figure S20.** Analysis of systematically varying each parameter at the Bolivian Arc location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s-1), (b) Q (50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.

**Figure S21.** Analysis of systematically varying each parameter at the Peruvian Forearc location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s-1), (b) Q (50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.

**Figure S22.** Analysis of systematically varying each parameter at the Chilean Forearc location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s-1), (b) Q (50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.
Table 1. Table summarizing seismic networks used in this study.

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| Number of Months of Data per Time Period: |
| Time Period I | 18 | 13 | 8 | 13 | 27 | * | * | * | * | 24 | * | 23 |
| Time Period II | * | * | * | * | * | 21 | * | * | * | 10 | * | 21 |
| Time Period III | * | * | * | * | * | 15 | 12 | * | 13 | * | 5 |
| Time Period IV | * | * | * | * | * | * | * | 16 | 16 | 16 | * |
Figure 1. Map of study area showing the tectonic setting of the Central Andes, locations of seismic stations used in this study, and Holocene volcanic activity from Siebert and Simkin [2002-]. Colored diamonds are broadband seismic deployments and the black diamonds are permanent stations from global networks. The black line follows the three-kilometer topography contour and defines the horizontal extent of Central Andean Plateau (CAP).
Figure 2 Map showing simplified geology [modified from Barnes and Ehlers, 2009], Holocene volcanic activity (black triangles) [Siebert and Simkin, 2002-], and morphotectonic provinces: Forearc, Western Cordillera, Altiplano, Puna, Eastern Cordillera, and Subandes [white dashed lines modified from Tassara, 2005]. Also shown are the locations of the ten representative locations a through j referred to in the text (white/black stars).
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</table>

Table S1. Table summarizing the number of high quality interstation paths used and the percentage of the maximum interstation paths for each period between 8 and 50 seconds.
Figure S1. The normalized roughness plotted against the damping value used at every point in the array with at least fair (150-800 km) resolution. (a) The normalized roughness for the Peruvian Eastern Cordillera (yellow line) and Chilean Forearc (red line) shown for reference. (b) Shows a “well behaved” dispersion curve (Peruvian Eastern Cordillera referenced in the text) where the selection of the damping value does not change the velocity as a function of period as much as the “poorly behaved” dispersion curve (Chilean Forearc referenced in the text) shown in (c).
Figure S2. Sensitivity kernels for each of the 10 periods used in the shear-wave inversion calculated from our final shear-wave velocity model at our Central Altiplano location shown in Figure 2 of the main text.
Figure S3. 2-D phase velocity results for the 8-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.
Figure S4. 2-D phase velocity results for the 10-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.
Figure S5. 2-D phase velocity results for the 12-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.
Figure S6. 2-D phase velocity results for the 14-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.
Figure S7. 2-D phase velocity results for the 16-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.
Figure S8. 2-D phase velocity results for the 20-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.
Figure S9. 2-D phase velocity results for the 25-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.
Figure S10. 2-D phase velocity results for the 30-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.
Figure S11. 2-D phase velocity results for the 35-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.
Figure S12. 2-D phase velocity results for the 40-second period. (a) Smoothed phase velocity results. Phase velocity results are contoured where resolution is less than fair (>800 km) and where resolution is good (>150 km). (b) Contoured resolution that is defined as the minimum separation distance at which two δ-like anomalies can be resolved [Barmin et al., 2001]. (c) High quality interstation paths used to invert for the 2-D phase velocity results (see Table S1). (d) Path density for selected period.
Figure S13. Analysis of systematically varying each parameter at the Subandean location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s$^{-1}$), (b) Q (50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.
Figure S14. Analysis of systematically varying each parameter at the Bolivian Eastern Cordillera location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s⁻¹), (b) Q (50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.
Figure S15. Analysis of systematically varying each parameter at the Peruvian Eastern Cordillera location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s$^{-1}$), (b) $Q$ (50, 200, 500), (c) $V_p/V_s$ (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.
**Figure S16.** Analysis of systematically varying each parameter at the Altiplano-Puna Volcanic Complex location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s\(^{-1}\)), (b) \(Q\) (50, 200, 500), (c) \(V_p/V_s\) (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.
Figure S17. Analysis of systematically varying each parameter at the Central Antiplano location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s\(^{-1}\)), (b) Q (50, 200, 500), (c) V\(_p\)/V\(_s\) (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.
Figure S18. Analysis of systematically varying each parameter at the Northern Arc location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s⁻¹), (b) $Q$ (50, 200, 500), (c) $V_p/V_s$ (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.
**Figure S19.** Analysis of systematically varying each parameter at the Peruvian Arc location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s\(^{-1}\)), (b) Q (50, 200, 500), (c) V\(_p\)/V\(_s\) (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.
Figure S20. Analysis of systematically varying each parameter at the Bolivian Arc location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s\(^{-1}\)), (b) \(Q\) (50, 200, 500), (c) \(V_p/V_s\) (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.
Figure S21. Analysis of systematically varying each parameter at the Peruvian Forearc location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s⁻¹), (b) Q (50, 200, 500), (c) Vp/Vs (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.
Figure S22. Analysis of systematically varying each parameter at the Chilean Forearc location shown in Figure 2. (a) shear-wave velocity (3.1, 3.6, 4.1, and 4.6 km s⁻¹), (b) $Q$ (50, 200, 500), (c) $V_p/V_s$ (1.70, 1.75, 1.80), (d) layer thickness in the top 75 kilometers (1 km, 5 km), and (e) damping value (150, 200, 250). Blue lines represent our preferred values with the black lines showing the effects of varying each parameter.