Imaging a magma plumbing system from MASH Zone to magma reservoir

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1 Abstract

2 The Puna Plateau of the Central Andes is a well-suited location to investigate
3 the processes associated with the tectono-magmatic development of a Cordilleran
4 system. These processes include long-lived subduction (including shallow and steep
5 phases), substantial crustal thickening, the emplacement of large volumes of
6 igneous rocks, and probably delamination. To elucidate the processes associated
7 with the development of a Cordilleran system, we pair Common Conversion Point-
8 derived receiver functions with Rayleigh wave dispersion data from Ambient Noise

Tomography. The resulting high-resolution shear wave velocity model of the southern Puna Plateau reveals the details of a lithospheric-scale magma plumbing system. Slow velocities near the crust-mantle transition are interpreted as a MASH Zone (a partially molten zone where mantle-derived melts interact with the lithosphere and undergo density differentiation) with ~4-9% melt. After differentiation, less dense and presumably more felsic melts propagate to shallower depths within the crust (~20 km below surface) and comprise vertically (~10 km) and laterally (~75 km) extensive slow velocity bodies that span the frontal arc and plateau interior. These large slow velocity bodies represent a partially molten mid-crust (up to 22%) where magma can further evolve to higher silica concentrations. The periodic influx of melt from the underlying MASH zone into these mid-crustal bodies may serve as a trigger to the eruption of the voluminous ignimbrites observed in the southern Puna Plateau. Many of the active tectonic processes operating along the southern Puna Plateau are thought to be analogous to the processes that formed the North American Cordillera. Thus, these results could provide insight into some of the processes associated with the development of a Cordilleran margin.

Keywords: Cordillera; Andes; Magmatic Arcs; Seismology; Crustal Structure

1. Introduction

Spatially extensive subduction-related magmatic products form at Andean (Cordilleran) margins. The spatial footprint of magmatism is complicated by long-term subduction, frontal arc migration, and changes in upper plate deformation over

time, all of which lead to various types of magmatic additions to the crust (DeCelles et al., 2009). The intrusion of large igneous bodies are the aggregate result of long term subduction at an Andean margin and represent the roots of frontal arcs and interior volcanism generally characterized by the eruption of felsic-intermediate ignimbrites (Ducea et al., 2015c; Best et al., 2009; Freymuth et al., 2015) . Despite in-depth petrological studies of Cordilleran-type frontal arcs and neighboring interior magmatism (Saleeby et al., 2003; Lipman and Bachmann, 2015; Ducea et al., 2015c; Best et al., 2016), vertically extensive seismic images of magma plumbing systems have only recently been resolved thanks to improvements in seismic coverage and imaging techniques (Ward et al., 2014; Huang et al., 2015; Kiser et al., 2016). An active type-example of these melt-forming and storing systems can be found in the South American Cordillera, where large-scale intermediate/felsic volcanism and ignimbrite flare-ups are thought to be the surface expression of batholith formation at depth (de Silva and Gosnold, 2007). Evidence is mounting from both petrological and geophysical data that large-scale igneous bodies stall in the mid-crust and undergo further crystal fractionation and mixing with crustal material as they transition from their more mafic initial compositions to felsic compositions (Kay et al., 2011; Ward et al., 2014; Comeau et al., 2015; Ducea et al., 2015a; Huang et al., 2015; Best et al., 2016). If these ponded magma reservoirs are laterally extensive, seismic imaging of these melt bodies is possible (Zandt et al., 2003; Ward et al., 2014).

Regions with long-lived volcanic activity are thought to be underlain by a Melting, Assimilation, Storage, and Homogenization (MASH) zone at the base of the

crust (Hildreth and Moorbath, 1988). The MASH zone represents a level of neutral buoyancy for mantle-derived basaltic melt where fractionation, differentiation, and mixing with lower crustal material occurs. This lower-crustal mixing zone eventually differentiates based on density, with the intermediate and felsic compositions rising to shallower depths in the crust and the mafic-ultramafic residuals or “cumulates” remaining near the base of the crust or sinking back into the mantle (Hildreth and Moorbath, 1988; Saleeby et al., 2003). Despite the petrologic evidence that these MASH zones exist beneath volcanically active regions, high-resolution seismic images of these zones are difficult to obtain.

The southern Puna Plateau was the site of a passive seismic experiment deployed from 2007 to 2009 by US, German, Argentine and Chilean institutions (Fig. 1) (Bianchi et al., 2013). Numerous seismic studies using different techniques, including teleseismic and regional distance body wave tomography (Bianchi et al., 2013), P- and S-wave receiver functions (Heit et al., 2014), two plane-wave tomography using earthquake-generated surface waves (Calixto et al., 2013), and body wave attenuation (Liang et al., 2014), characterize the first-order lithospheric structure and identify some important magmatic features in the region. However, the techniques utilized in these studies have some inherent resolution limitations within the crust.

We utilize a recently developed technique for the joint inversion of receiver functions and surface-wave dispersion data (Julia et al., 2000; Delph et al., 2015) from this dense array of seismic stations to obtain a 3D shear wave velocity model of this magmatically-active subduction-related continental plateau. This resulting

model has better crustal resolution than previous models and illuminates a magmatic plumbing system from the base of the lithosphere to the upper levels of the crust. We interpret these images in the context of the detailed tectonic and petrologic history of the southern Puna Plateau to understand the evolution of the continental lithosphere through subduction-related processes (e.g. the compositional variation of igneous rocks, plutonic-volcanic (P:V) ratios, and lithospheric foundering). As the western margin of South America is often interpreted as a modern analogue to the North American Cordillera, these results have broad implications for long-lived magmatic arcs and the development of Cordilleran systems.

Figure 1: The seismic station distribution (yellow squares: intermediate or broadband stations; inverted triangles: short-period stations) in the southern Puna Plateau and surrounding region. Elevations >3 km represent the southern Puna Plateau (unshaded). Only stations that had high-quality receiver functions are plotted. Irregularly shaped polygons represent ignimbrite deposits colored by eruptive age, and black semi-circular lines represent mapped ignimbrite calderas (from the Andes Ignimbrite Database, Freymuth et al., 2015 and refs. therein). Black units are basaltic deposits mapped at the surface, many of which show a geochemical signature for delamination (Kay et al., 1994; Ducea et al., 2013; Murray et al., 2015). Red star represents location of shear velocity profile in Fig. 4 within the Cerro Galan Caldera.

2. Methods

2.1 Receiver Function Computation

This dataset arises from the analysis of 92 short, intermediate, and broadband period range stations deployed from throughout the southern Puna Plateau from the REFUCA project (19 stations, 2002-2003) and the complimentary German – American – Chilean – Argentine projects of PUDEL (PUna DELamination) and SLIP (Seismic Lithospheric Imaging of the Puna Plateau; Sandvol & Brown 2007) that operated from 2007-2009. We computed 5919 receiver functions from earthquakes with magnitudes between 5.5 and 7.6 from distances between 30 and 90 degrees.

For short period instruments, the instrument response was deconvolved from the waveforms to enhance the sensitivity to lower frequency signals in the seismograms (Niu et al. 2005). The short period data was kept in units of velocity and incorporated into the larger dataset (Fig. 2A). All traces were then filtered to enhance the frequency range of interest and those that did not display clear first arrivals were discarded from further analysis. Radial receiver functions were computed using the iterative time-domain receiver function computation described

by Ligorria & Ammon (1999). We used a Gaussian pulse width of 2.5 due to its ability to resolve robust crustal structures (vertical resolution of ~ 1 km). The receiver functions were then inspected using the FuncLab package (Eagar and Fouch 2011) for further quality control.

2.2 Ambient Noise Tomography

The ambient noise tomography dataset of Ward et al. (2013) was augmented by recently available data from the PUDEL (Heit et al., 2007), SLIP (Sandvol and Brown, 2007), and PLUTONS (West and Christensen, 2010) deployments. These datasets were processed as described by Ward et al. (2013) to obtain phase velocities at 13 periods between 8 and 50 seconds (Supp. Fig. S1). To assure only high quality data was used in the calculation of phase velocities, all cross-correlations with a signal-to-noise ratio <10 , an interstation spacing of <3 times the wavelength at the period of interest, and unrealistic phase velocities (<1.5 km/s and >5 km/s) were discarded from the 2-D phase velocity inversion (Barmin et al., 2001). Also, any data that had residuals >3 seconds after the 2-D tomographic inversion were also discarded. The resulting dispersion model has a spatial resolution of $\sim 0.25^\circ$ based on the Rayleigh criterion (Barmin et al., 2001), and a 50 km Gaussian smoothing parameter was used to stabilize of the inversion. Thus, amplitudes <50 km may be underestimated in the phase velocity results. Because of the station geometry available previous to this study, the inclusion of the newly available data imposed few changes to the phase velocity results of Ward et al. (2013), and the period range used in this study has good sensitivities down to ~ 80

km (Fig. 3). PLUTONS data were not included in the receiver function analysis, as they were located outside our study area.

2.3 CCP-derived Receiver Function Computation

We use the approach described by Delph et al. (2015) to create receiver functions from Common Conversion Point (CCP) stacking analysis (Dueker & Sheehan 1997). This approach has the advantage that it maps receiver functions back to their theoretical locations in the Earth and linearly stacks the amplitudes of receiver functions that fall within a limited cell to create a 3D volume of impedance contrasts throughout the study area. This helps correct for backazimuthal structural variations beneath a station and decreases noise present in the receiver functions. In our CCP stacking analysis, the spatial dimensions of the cells over which to average the receiver function amplitudes have a minimum dimension of 0.2° laterally by 0.5 km vertically. We allow the lateral dimension of our bins to dilate in increments of 0.1° until 10 rays are incorporated into the bin or a width of 1° is reached. A bin-spacing of 0.1° is used to match the spacing of the ambient noise tomography volume. We use a simple 1D velocity model consisting of 65 km thick crust with a 6.1 km/s P-wave velocity and a 1.74 V_p/V_s ratio, and a mantle of 7.5 km/s P-wave velocity with a V_p/V_s of 1.78 to migrate receiver functions to depth. While assuming a 1D velocity model throughout the study area is an oversimplification, the velocity model used in the computation of the CCP stacks will have a very small effect on our resulting CCP-derived receiver functions, as we migrate our CCP stacks back to time using the same velocity model.

As the receiver functions are ray-traced to depth, they spread out laterally beneath a station. This leads a higher likelihood of data from different stations being incorporated into the same bin and increases the spatial resolution of the receiver functions with depth. However, the incorporation of data into a vertical profile that may not have been present in shallower depth bins can lead to unrealistic amplitude jumps in the resulting CCP-derived receiver functions that have a non-Gaussian shape (Fig. 2B, black line). To reconcile this, Delph et al. (2015) used wide bin widths to decrease the effect that these amplitude artifacts could have on the resulting shear wave velocity profiles when jointly inverted with dispersion data. In order to make this approach more robust and allow finer lateral resolution, we modify the approach presented by Delph et al. (2015) to filter out spurious amplitude changes and obtain a smooth waveform that can be accurately modeled by the forward problem of the receiver function portion of the joint inversion.

2.3.1 Gaussian Filtering of CCP-derived Receiver Function

Receiver functions can be generalized as the difference in a recorded signal between a radial-component seismogram and a vertical-component seismogram convolved with a Gaussian pulse width. Thus, a receiver function can be characterized by the following equation in the time domain (t):

$$r(t) = f(t) * g(t) \quad (1)$$

where $r(t)$ is our resulting receiver function, $f(t)$ is the Green's function (in this case, a spike train of the differences between vertical and radial components) and $g(t)$ is a Gaussian filter whose width is defined by the alpha parameter (α_g), which has the general form:

$$g(t) = e^{-\alpha g^2 t^2} \quad (2)$$

This leads to very smooth waveforms that can be defined by the summation of Gaussian pulses centered at different times. In receiver functions derived from the CCP volume, the number of receiver functions incorporated into a bin may vary based on bin size, bin spacing, and the location of the back-projected receiver functions at depth. This has the potential to cause our CCP-derived receiver functions to have unrealistic, sharp changes in amplitude. These unrealistic changes in amplitude are generally small and have a negligible effect on the resulting shear-wave velocity profile when used in a joint inversion. However, we build on the approach of Delph et al. (2015) to smooth out these artifacts at the slight cost of vertical resolution.

In this application, we filter our CCP-derived receiver functions with a Gaussian filter $h(t)$ to dampen unrealistic structures. This makes our new receiver function have the following form:

$$r_{filt}(t) = f(t) * g(t) * h(t) \quad (3)$$

where $h(t)$ has the same form as $g(t)$, but a different Gaussian parameter. The convolution of a Gaussian filter with our CCP-derived receiver function changes the “effective” alpha parameter of our resulting receiver function by broadening the Gaussian pulse. Convolution has commutative and associative algebraic properties, thus:

$$r_{filt}(t) = [f(t) * g(t)] * h(t) = f(t) * [g(t) * h(t)] \quad (4)$$

meaning that this multi-step Gaussian filter technique is equivalent to producing a receiver function with a Gaussian pulse width equal to that of produced by the

convolution of two Gaussian pulses with different widths. This relationship allows us to choose a Gaussian filter that will produce a receiver function with a Gaussian pulse width of our choice (which must be lower than the initial Gaussian pulse width of the receiver functions). For example, if we want to obtain a 2.2 (α_{filt}) Gaussian equivalent receiver function from initial 2.5 (α_g) Gaussian receiver function data, we could solve for the necessary Gaussian filter width parameter with the following equation:

$$\alpha_h = \sqrt{\frac{-(\alpha_{filt}^2 \times \alpha_g^2)}{\alpha_{filt}^2 - \alpha_g^2}} \quad (5)$$

where the subscripts correspond to Gaussian functions in Eqn. (4). Using this equation, we find that a Gaussian filter with an alpha parameter of ~ 4.6 would produce a 2.2 Gaussian equivalent receiver function. These are the values we use in this study. This process, however, does not retain amplitude information, thus we must also perform an amplitude correction for the filtered receiver functions.

2.3.2 Amplitude Correction

The amplitude of a receiver function is linearly related to the Gaussian pulse width and is a function of ray parameter. Thus, we can define the relationship as follows:

$$A(p) = m(p) \times \alpha + b(p) \quad (6)$$

where $A(p)$, $m(p)$, and $b(p)$ is the receiver function amplitude, slope, and y-intercept of the linear trend as a function of ray parameter, and α is the Gaussian pulse width. To calculate the amplitude change between two receiver functions with

a different alpha parameter, we simply divide the amplitude of the receiver function of interest by the amplitude of the receiver function of our data:

$$\frac{A_1(p)}{A_2(p)} = \frac{m(p) \times \alpha_1 + b(p)}{m(p) \times \alpha_2 + b(p)} \quad (7)$$

In this case, α_1 is a 2.2 and α_2 is our input data (2.5). We find that, for any ray parameter, the change in the amplitude between two receiver functions of different Gaussian pulse widths is constant (i.e. the relationship between the y-intercept (b) to the slope (m) is constant). This property simplifies this relationship to the following form:

$$\frac{A_1}{A_2} = \left(\frac{1}{\alpha_2 + b/m} \right) \alpha_1 + \left(\frac{b/m}{\alpha_2 + b/m} \right) \quad \text{where } b/m = 0.226 \quad (8)$$

Thus, from the above equation with our Gaussian pulse widths of interest ($\alpha_1 = 2.2$, $\alpha_2 = 2.5$), we find that the amplitude of a 2.2 Gaussian receiver function will have 89% of the amplitude of a 2.5 Gaussian receiver function.

In summary, in this altered new approach, we effectively filter our CCP-derived receiver functions with a Gaussian pulse. Convolution of a “2.5” receiver function with a Gaussian filter that has an alpha parameter of 4.6 produces a smooth receiver function with an effective alpha parameter of 2.2 (Fig. 2B), thus smoothing across unrealistic artifacts that are the result of the incorporation of different data into a bin at the cost of vertical resolution. Changing the Gaussian pulse width of a receiver function also has an effect on the resulting amplitude of a receiver function, as lower alpha parameters yield much broader Gaussian pulses. The relationship between Gaussian pulse width and amplitude is linear with respect to ray parameter

and crustal structure, and changing from a 2.5 to a 2.2 Gaussian receiver function yields an amplitude that is 89% of the 2.5 Gaussian receiver function amplitude.

Figure 2: (A) Record of a M 6.5 earthquake that occurred on Sept. 24, 2009 recorded on the vertical component of two seismic stations. Time-series begins 10 seconds before the P-wave arrival. (1) Raw short period seismogram, (2) Short period velocity seismogram after deconvolution of instrument response and filtering, (3) Velocity seismogram from intermediate period sensor after filtering. The similarities of the deconvolved short-period recording and intermediate recording for a teleseismic event indicate that the short period data may be used to compute receiver functions. (B) End-member example of a CCP-derived receiver function profile for a single gridpoint extracted from the CCP volume (black) to the Gaussian-filtered CCP-derived receiver function with amplitude scaling (red). The Gaussian filter mitigates spurious artifacts that can be introduced into the vertical receiver function profile by the incorporation of new receiver function data into a depth bin.

2.3.4 Joint Inversion

Surface wave dispersion data and receiver functions are highly complementary datasets, as the shortcoming of each individual dataset is the strength of the other (Julia et al. 2000; Delph et al. 2015). While surface wave dispersion data are good at recovering absolute shear wave velocities as a function of depth, the broad sensitivity kernels (e.g. Fig. 3A) lead to a final model that is highly dependent on an often poorly constrained *a priori* model, as discontinuities

275 will not be resolved by the inversion of surface wave dispersion data alone. In
276 contrast, receiver functions are very sensitive to impedance contrasts in the Earth,
277 but have virtually no sensitivity to the absolute velocities on either side of a
278 boundary. By combining these two datasets in a joint inversion approach, we reduce
279 the non-uniqueness of the inverse problem, mitigating the need to have an *a priori*
280 starting model that is close to the real Earth structure (Delph et al. 2015). In this
281 study, the resulting 2.2 effective Gaussian receiver functions derived from CCP-
282 stacking are paired with the augmented ambient noise-derived dispersion data from
283 Ward et al. (2013) following the procedure outline by Delph et al. (2015) (Fig. 3B).
284 The joint inversion algorithm requires a dataset weighting parameter that defines
285 the relative importance of each dataset to the inversion penalty function, with a
286 value of 0 corresponding to an inversion of only the receiver function data and a
287 value of 1 corresponding to an inversion of only dispersion data (Julia et al. 2000).
288 We choose a weighting parameter of 0.3, as it leads to a good balance between
289 receiver function fit and dispersion fit. The fit to the dispersion data generally does
290 not improve for weights higher than this value (Delph et al. 2015). The initial model
291 for the joint inversion used in this study consists of a 4.5 km/s half-space with 2 km
292 thick layers, as the vertical resolution of a 2.2 Gaussian receiver function is ~ 2 km.
293 Assuming a very simple velocity model with no *a priori* information about structure
294 assures that all information in the resulting shear wave velocity model is necessary
295 to satisfy the constraints imposed by the receiver functions and dispersion data.

Figure 3: (A) Sensitivity kernels of Rayleigh wave phase velocities to shear wave velocity structure for the periods for a simple half-space (constant velocity) model. **(B)** Dispersion curves (1) and CCP-derived receiver functions normalized to coherence peak (2) along Latitude -26. Each row of (1) is a phase velocity for a specified period (8, 10, 12, 14, 16, 18, 20, 25, 30, 35, 40, 45, 50 seconds) at a tomographic grid-point. These curves are paired with the CCP-derived receiver functions (2) for the joint inversion (resulting shear velocity cross-section shown in Fig. 5E). Receiver functions are normalized to ray parameter of 0.06 s/km. Purple circles: locations of calderas that sourced ignimbrite eruptions; Red triangles: Holocene volcanoes; Inverted triangles: seismic stations; Black diamonds: mafic volcanic samples from Ducea et al. (2013) and Murray et al. (2015).

3. Results

Bianchi et al. (2013) proposed the presence of an areally extensive magma body beneath the southern Puna Plateau based on a broad low-velocity anomaly extending from the crust into the upper mantle. Further evidence for the existence of the “Southern Puna Magmatic Body” (SPMB; Bianchi et al. 2013) was obtained via earthquake-generated dispersion data (Calixto et al. 2013) and attenuation (Liang et al. 2014), with the largest anomaly lying beneath the Cerro Galan Caldera (Heit et al. 2014). In this study, we image the finer-scale crustal structure of the SPMB and observe three separate, tabular slow velocity bodies that together comprise the

318 SPMB. These bodies are spatially coincident with regions of active arc volcanism in
319 the Lazufre Volcanic Region (Lazufre Anomaly, LA), the southernmost Central
320 Volcanic Zone (Incahuasi Anomaly, IA), and in the interior of the southern Puna
321 plateau beneath the largest ignimbrite center in the region, the Cerro Galan Caldera
322 (Cerro Galan Anomaly, CGA; also imaged by Heit et al. 2014). These bodies are
323 approximately ~10 km thick, with the top of the slow velocity zones ~16-21 km
324 below the surface and the slowest seismic velocities at ~20-25 km depth (Fig. 4, 5).
325 These bodies appear distinctly separate at mid-crustal depths, and have trench-
326 perpendicular widths of ~50-100 km based on the ~3.3 km/s average crustal shear
327 wave velocity contour (top 50 km of our model; Fig. 6A). We choose this velocity
328 contour because profiles with average shear wave velocities lower than ~3.3 km/s
329 most likely contain partial melt, as it is slower than expected for crystallized,
330 isotropic crustal compositions at mid-crustal depths (Christensen, 1996).

Figure 4: Jointly inverted shear wave velocity profile (A) from dispersion (B) and receiver function (C) data for a gridpoint in the Cerro Galan Caldera (Red Star – Fig. 1). Black points (B) and black line (C) are observed data, while the red lines are predicted by the red model shown in (A). Velocities <3.3 km/s are slow for crustal compositions at mid-crustal depths, and are inferred to contain some amount of partial melt. Shaded red regions in (B) and (C) are responsible for the slow shear wave velocities in (A). The seismic Moho (SM) is thought to be ~59 km (Heit et al., 2014). Note the slow shear velocities and a very low velocity gradient across the seismic Moho (SM). CGA, Cerro Galan Anomaly; NRMS, Normalized Root Mean Square.

Crustal thickness beneath the southern Puna Plateau vary from ~50-65 km, being thicker in the north and slightly thinner in the plateau interior (Wölbern et al., 2009; Tassara and Echaurren, 2012; Heit et al., 2014). The crust-mantle velocity transition throughout much of the region appears to be rather gradational based on broad low-amplitude receiver function conversions and a lack of an observable inflection in the phase velocity dispersion curves associated with the transition to

higher shear wavespeeds (Fig 3B; Supp. Figs. S3-7) (e.g. Lebedev et al., 2013). After the joint inversion, we observe shear wave velocities that are much slower than expected for typical mantle lithosphere down to the bottom of our model (<4.1 km/s vs. 4.5 km/s, Kennett and Engdahl, 1991; Figs. 4, 5C-H). Further to the south along the transition from the Puna Plateau to the Sierras Pampeanas, this gradational velocity contrast gives way to a much sharper transition to faster wavespeeds associated with typical mantle lithosphere (Fig. 5H, 6B).

We note that our joint inversion results are generally consistent with the receiver function study of Heit et al. (2014) but there are some differences that warrant brief discussion. Heit et al. (2014) identified a strong Moho arrival for much of the region, and also identified a relatively prominent lithosphere-asthenosphere boundary (LAB) at 75 to 80 km depth in their migrated P and S-receiver function cross-sections. We believe some of the differences we observe are due to the different processing of the receiver functions, as Heit et al. (2014) processed the data in a manner to emphasize the Moho and LAB arrivals. This included using somewhat lower frequency content (especially for the S-receiver functions) and stacking data in ~ 100 km bins along specific profiles. These are reasonable choices to characterize the first-order lithospheric structure for this region. Our processing involved higher frequency data aimed at maximizing resolution within the crust, on which we will focus in the following sections.

Figure 5: Mapslices (A-C) and cross-sections (D-H) relative to sea level through our velocity model. Purple circles: ignimbrite centers; White/Red triangles: Holocene volcanoes; Inverted triangles: seismic stations; Black diamonds: mafic volcanic samples from Ducea et al. (2013) and Murray et al. (2015); white lines in cross-sections: Moho depth from Tassara & Echaurren (2012). Shear velocity contour intervals every 0.2 km/s. (A) Mid-crustal magma bodies show strong spatial relationship with ignimbrite centers and Holocene volcanoes. White dots/bars along cross-section lines demarcate 100 km intervals. Cross-section A-B shown in Fig. 8. (B) LA and CGA appear to be merging with depth. (C) MASH zones apparent as very slow seismic velocities (~ 4 km/s) at mantle depths (>60 km) and correlate well with locations of mafic volcanics at the surface (black diamonds). (D-H) Cross-sections at different latitudes through our study area. Locations of cross-sections shown as horizontal lines in (A). Arrows indicate possible magma pathways from

MASH zone to mid-crustal magma reservoirs. CGA, Cerro Galan Anomaly; IA, Incahuasi Anomaly; LA, Lazufre Anomaly

4. Discussion

4.1 Crustal Magma Reservoirs

The trench-perpendicular width of the crustal slow velocity bodies (~50-100 km, Fig. 6A) makes them the most dominant features of the crust in the southern Puna Plateau, but what do these spatially extensive slow velocity bodies represent? Recent high-resolution seismic images in the northern Puna Plateau (Altiplano-Puna Volcanic Complex, APVC) imaged a similar seismic anomaly, interpreted as a magma reservoir that was feeding the prolific volcanism of the APVC (Ward et al. 2014). Based on the spatial correlation of these slow velocity bodies with surficial volcanism both in the arc and backarc, we also interpret these bodies to represent extensive mid-crustal magma reservoirs (or more appropriately, mush bodies) that are feeding volcanism in the southern Puna Plateau.

Figure 6: (A) Average shear-wave velocity in the top 50 km of our velocity model. Slow seismic velocities correspond well with the active Central Volcanic Zone (white

triangle) and the Cerro Galan Caldera. CGA: Cerro Galan Anomaly; IA: Incahuasi Anomaly; LA: Lazufre Anomaly. **(B)** Depth to the 4.2 km/s shear-wave velocity in our model (relative to sea level). Purple region indicates where a velocity of 4.2 km/s was not reached by the bottom of our seismic velocity model, and is indicative of regions that have undergone delamination. Data is filtered using a 0.3x0.3 degree boxcar filter to show only long-wavelength features. Inverted triangle are seismic stations used in this study, white circles are ignimbrite centers, white upright triangles are Holocene volcanoes, and black diamonds are mafic volcanic samples from Murray et al. (2015) that approximately delineates recent (<8 Ma) mafic volcanism in the southern Puna.

It is unlikely that these anomalies represent a single intrusion at depth, as an active magmatic arc that operates at any one time is generally much smaller in lateral extent (~35 km; Ducea et al., 2015b) than our slow velocity bodies, which are comparable to the width of the entire Sierra Nevada composite batholith (~80 – 120 km). However, the thermal history and longevity of magmatic arcs is an ongoing research problem (Gelman et al., 2013), and some of the spatial extent of the slow velocity bodies could be attributed to thermal effects of recently intruded, hot material on seismic velocities. Also, plutons exposed in the North American Cordillera generally have crystallization depths that are shallower than the depths to the top of our slow velocity anomalies (Chapman et al., 2012), and other seismically imaged magma bodies are generally smaller and shallower as well (Seccia et al., 2011; Kiser et al., 2016). Hence, we suspect that the zones of slow velocity imaged in this study may be sourcing similar shallow magma bodies that may be too small to resolve using this approach, such as the one imaged beneath the Lazufre region by Spica et al. (2015). Deeper exposed sections in the Sierra Nevada batholith and Famatinian arc immediately south of the Puna Plateau indicate that the dominant fabrics of arc rocks undergo a transition from vertical igneous-

metamorphic fabrics (e.g. stocks) to planar fabrics at ~20 km depth, in general agreement with the top of the LVZs in the southern Puna Plateau (Saleeby et al., 2003; Ducea et al., 2015a; Ducea et al., 2015c). Thus, the slow velocity bodies we image may represent zones of partially molten material that have stalled at this transition.

4.2 The Cerro Galan Magma Body

The Cerro Galan Anomaly (CGA) is the largest slow velocity anomaly imaged in this study and ranges in depth from 12 – 24 km below sea level (BSL) or 16 – 28 km below surface, with an estimated volume of ~22,000 km³ based on the 2.9 km/s velocity contour (a rather conservative estimate for regions containing partial melt, see Ward et al., 2014). This anomaly appears to be a robust feature of the crust, as it has been imaged by multiple other techniques at varying resolutions (Bianchi et al., 2013; Calixto et al., 2013; Heit et al., 2014; Liang et al., 2014). The depth of the CGA in this study shows remarkable consistency with geobarometric measurements of Cerro Galan-sourced ignimbrites that indicate magmatic hybridization takes place ~16 km BSL (20 km below surface; Kay et al., 2011) (Fig. 4). Approximating this slow velocity body as an oblate spheroid, it has a horizontal to vertical aspect ratio of ~4.5:1, slightly less than the ~10:1 ratio of the slow velocity body associated with the Altiplano-Puna Volcanic Complex (Ward et al., 2014) and on the smaller end of other imaged magma reservoirs which tend to have a range between 5-10:1 (Bachmann and Bergantz, 2008).

In plateaus, the underthrusting and burial of upper crustal igneous rocks and sediments due to foreland thrusting creates an environment that is fertile for

anatexis (Decelles et al., 2009) (i.e. the formation of migmatites), as residence time in the mid-crust is extended due to low erosion rates in plateau interiors. A possible exposed analog for the CGA is found in the Angel Lake metamorphic suite located in the Ruby Mountains (Nevada, USA). This metamorphic suite is a 20-30 km exposure of the Cretaceous-Eocene Nevadaplano crust (Howard et al., 2011). It contains a large proportion (>50%) of in-situ or short-traveled silicic melts derived from the partial melting of local Precambrian and Paleozoic rocks (migmatites) intermixed with sills of Eocene mafic to intermediate intrusives that were generated near the base of the crust. In the context of the southern Puna Plateau, the periodic intrusion of mafic-intermediate melts into a mid-crustal migmatized zone as a result of the steepening of the subduction angle (and possibly delamination) could act as rejuvenation events to the magma reservoir, prolonging its state as a partially crystallized mush zone and increasing its lateral extent (e.g. Cao et al. 2016). The influx of this new, hot material from below could have resulted in the eruption of the ignimbrites associated with the Cerro Galan Caldera ranging in age from ~6-2 Ma.

Following the approach of Ward et al. (2014), estimates of plutonic to volcanic (P:V) ratio can be calculated based on the cumulative erupted volume of the Cerro Galan Caldera (~1300 km³; Kay et al., 2011), and the percentage of material encompassed by the 2.9 km/s contour that is considered to be plutonic (Fig. 7A). Considering an estimate that 50% of the total volume of the slow velocity body is plutonic (i.e. in-situ melt and igneous rocks that are coeval with volcanism) based on observations from the Ruby Mountains leads to a P:V ratio of ~8:1. This is a very conservative estimate for the P:V ratio since we are neglecting melt that was most

likely introduced via subduction processes. Also, if a higher seismic velocity is used to estimate the volume of plutonic material (possibly up to ~ 3.3 km/s), this ratio increases to values $>10:1$, which is larger than often cited ranges for P:V ratios ($\sim 5:1$; White et al., 2006). At velocities >3 km/s, the volume of the CGA no longer closes in our model, which hinders the meaningful estimation of P:V ratios if a higher velocity is assumed to contain partial melt. However, the range of P:V ratios that we obtain in this study accompanied by recent estimates of $\sim 25\text{-}35:1$ in the APVC (Ward et al., 2014) and the $>10:1$ in the San Juan Batholith (Lipman and Bachmann, 2015) indicate that much higher P:V ratios exist than previously expected. The interactions between the host rock and these large volumes of intruded material from the mantle may be a significant contributor to the formation of more felsic continental compositions, especially in long-lived systems like the North and South American Cordillera. Estimating P:V ratios for the frontal arc (IA and LA) is not possible at the present time because: 1) accurate estimates of the volume of erupted volcanic material in the frontal arc do not exist at this time (it is hard to measure the depth extent of volcanic material below the surface in long-lived arcs), 2) the shear wave velocity anomalies are not fully enclosed by our current results (e.g. Fig. 5A).

Using the relationship between shear velocity and melt fraction of Takei (2002) and the contiguity-porosity relationships from Yoshino et al. (2005) for a felsic system, we can estimate the percentage of partial melt in the CGA. The relationship between melt fraction and shear wave velocity is dependent on the contiguity of the melt in a crystal matrix (a function of dihedral angle), the melt fraction (or porosity) of the partially molten region, and the densities of both the

crystalline rock and its compositional equivalent for melt (Takei, 2002). The dihedral angle of a crystal mush (crystals + melt) is largely dependent on the composition of the melting material, and was experimentally derived for various compositions by Yoshino et al. (2005). The lowest shear wave velocity we obtain in the CGA is ~ 2.7 km/s: a 25% decrease in shear velocity assuming that a compositionally similar, crystalline rock would have a shear velocity of a granite/granodiorite (~ 3.6 km/s) (Fig. 7B). This decrease in seismic velocity indicates a maximum melt percentage of $\sim 22\%$ in the CGA. The rheology of a silicate melt is thought to increase rapidly after the melt percentage decreases below $\sim 60\%$, inhibiting the ability of the partially molten body to internally convect (Marsh, 1981; Lejeune and Richet, 1995). Thus, barring a rejuvenation event by melt from below, the CGA is unlikely to represent an eruptible body.

Figure 7: (A) P:V ratios of the CGA depending on shear velocity used to constrain the volume of partially molten material and the percentage of that material which is considered plutonic. In order to make a conservative estimate of the P:V ratio for the CGA, we use the 2.9 km/s shear velocity contour and 50% plutonic material as observed in the migmatites in the Ruby Mountains (Howard et al., 2011) to obtain a P:V ratio of 8:1. This estimate does not include plutonic material introduced via subduction processes. **(B)** Normalized shear wave velocity plotted against melt

percentage from the relationships of Yoshino et al. (2005) and Takei (2002). For the CGA, the maximum melt percentage is ~22% (assuming a Vs of 3.6 km/s should the LVZ crystallize completely). In the MASH zone, melt percentage ranges between ~4 – 9% depending on bulk composition and a Vs of 4.5 km/s should the slow velocity material crystallize completely.

4.3 Melt at the Crust-Mantle Transition: Seismic Imaging of a MASH Zone

Beneath volcanics arcs lies the melting, assimilation, storage, and homogenization (MASH) zone (Hildreth and Moorbath, 1988) located at the crust-mantle transition. In this zone, the lower densities associated with crustal material causes slightly denser mantle-derived melts to stall, mix with lower crustal material, and undergo density differentiation. From this level, less dense (and more felsic) melts can rise to shallower depths in the crust, leaving behind dense “cumulates” of mafic-ultramafic material that either remain near the crust-mantle transition or gravitationally sink back into the mantle (Hildreth and Moorbath, 1988; Saleeby et al., 2003). The MASH zone is particularly difficult to image due to its dominantly mafic-ultramafic composition with small amounts of partial melt (Hildreth and Moorbath, 1988), as partially molten mafic material will have seismic velocities similar to that of both a crystallized felsic-intermediate lower crust as well as warm upper mantle (asthenospheric) material.

The presence of a felsic lower crust is unlikely based on geochemical constraints given by basaltic samples at the surface, which indicate a lower crust of dominantly pyroxenitic compositions (Ducea et al., 2013; Murray et al., 2015). Also, a gradational crust-mantle transition at ~50-65 km in the southern Puna Plateau is consistent with more mafic lower crustal compositions (Tassara and Echaurren, 2012; Heit et al., 2014). Based on the very slow seismic velocities imaged beneath

the central Puna Plateau at mantle depths (<4.1 km/s vs. 4.5 km/s for typical mantle; Kennett and Engdahl, 1991), we infer that these slow seismic velocities represent the presence of partially molten mafic-ultramafic compositions at and near the crust-mantle transition (i.e. MASH zone) (Fig. 4, Fig. 5D-H). Assuming an overall ultramafic bulk composition in the lower crust (Kay et al., 1994; Ducea et al., 2013), this partially molten zone is estimated to contain ~ 4 -9% partial melt (Takei, 2002; Yoshino et al., 2005) (Fig. 7B).

We use the shallowest depth where our velocity model reaches 4.2 km/s to indicate the transition across the seismic Moho, as generally only crystallized ultramafic rocks will have wavespeeds greater or equal to this value (e.g. hornblendite, pyroxenite, peridotite; Christensen, 1996). Similar to the western United States, significant removal of dense lower crust and mantle lithosphere is thought to have occurred through the process of lithospheric foundering (Ducea and Saleeby, 1998; Zandt et al., 2004; Erdman et al., 2016) based on teleseismic tomography (Scire et al., 2014; Beck et al., 2015), structural data (Schoenbohm and Carrapa, 2015), and the isotopic signatures of mafic volcanism in the region (Kay et al., 1994; Ducea et al., 2013; Murray et al., 2015) (Fig 6B). Regions where velocities of 4.2 km/s are not reached by 80 km may represent areas that have undergone delamination, which has allowed for partial melt to accumulate due to the influx of asthenosphere and associated decompression melting (purple region, Fig. 6B). Interestingly, the few mafic volcanic deposits in the Puna Plateau spatially correlate with the slowest seismic velocities in the uppermost mantle and anti-correlate with inferred mid-crustal magmatic reservoirs (black diamonds, Fig. 6A). Our preferred

explanation is that the seismically faster material that separates the magma reservoirs is crystalline crust that does not impede the low viscosity mafic melts from propagating to the surface. However, these melts can only make it to the surface where they are not captured by magma reservoirs in the mid-crust (Figs. 5A, 6A). If these rising basaltic melts intersect a magma reservoir, they are captured and incorporated into the magma reservoir, extending its lifetime. This could also lead to a rejuvenation and resulting eruption if the influx is high enough.

Should delamination have occurred in this region, it is unlikely that significant mantle lithosphere underlies the southern Puna Plateau. However, a negative velocity contrast is observed in S-receiver functions at ~80 km that has been interpreted as the lithosphere-asthenosphere boundary (Heit et al. 2014). We cannot verify the existence of a decrease in velocity at this depth with our results (maximum depth = 80 km), but this interpreted LAB should be located near the base of the MASH zone.

4.4 The Evolution of the Southern Puna Magmatic System

The evolution of the southern Puna Plateau based on Kay and Coira (2009) and these new seismic images is shown in Fig. 8. Initially, arc magmatism is generated by “typical” subduction zone processes (e.g. wet melting above the mantle wedge), leading to the creation of a hot and gradational crust-mantle transition zone (MASH zone) beneath the rheologically weak crust of the active frontal arc. The subducting slab then shallowed, possibly related to the southward migration of the Juan Fernandez ridge (Yáñez et al., 2001) leading to an eastward migration of the frontal arc (Fig. 8A). At the same time, underthrusting of the foreland beneath the

591 eastern margin of the Puna Plateau increased (Babeyko and Sobolev, 2005), which
592 introduced sedimentary material into the mid-crust leading to the formation of a
593 migmatized (partially molten) zone. The shallowing of the subducting slab may have
594 also led to hydration of the mantle wedge, similar to what is seen below the Sierras
595 Pampeanas today (Porter et al., 2012). During this time, crustal thickening, tectonic
596 shortening, the build-up of cumulates in the lower crust due to mantle addition, and
597 underthrusting of foreland sediments results in pyroxenitic lowermost crustal
598 compositions, which has densities similar to or greater than that of the underlying
599 asthenosphere (DeCelles et al., 2009; Ducea et al., 2013). As the slab rolls back and
600 returns to a more steep subduction angle (~ 30 degrees), room is created for the
601 dense and gravitationally unstable lower crust to founder as Rayleigh-Taylor
602 instabilities (Elkins-Tanton, 2005; Schoenbohm and Carrapa, 2015) (Fig. 8B).
603 Foundering then drives upward asthenospheric flow into the hydrated mantle
604 beneath the plateau, leading to an increased flux of mantle-derived melting on either
605 side of the foundering body (Elkins-Tanton, 2005; Elkins-Tanton, 2007). This results
606 in the formation of a backarc MASH zone that sourced the recent eruptions of
607 ignimbrites related to the Cerro Galan Caldera. As the foundering body is removed,
608 these separate MASH zones assimilate into one large MASH zone that we image as a
609 central zone of slow velocities beneath the southern Puna Plateau. This
610 amalgamated MASH zone now feeds both the arc volcanism as well as the magma
611 reservoir associated with the CGA (Fig. 6B, 7C). Between the frontal arc and the CGA,
612 relatively fast (~ 3.6 km/s) average shear wave velocities are observed (Fig. 8D) that
613 appear to separate the magma reservoirs of the frontal arc and backarc. This zone

614 may act as a rheological barrier to rising silicic melts, possibly diverting them along
615 the thermally weakened pathways of long-lived magma ascent, as expected from
616 modeling of magmatic systems (Cao et al., 2016). This zone is relatively strong
617 because it has not been exposed to the fluids and high temperatures associated with
618 magma ascent nearly as long as the arc and backarc regions, as the delaminating
619 body may have insulated it from rising asthenosphere and associated melt (Elkins-
620 Tanton, 2007) (Fig. 8B). After the delaminating body is removed, lower crustal
621 pyroxenites may undergo small-scale delamination, melting as they sink into the
622 mantle. This melt is then re-incorporated into the MASH zone (Elkins-Tanton,
623 2007), leading to the isotopic characteristics of the mafic deposits seen at the
624 surface today (Kay et al., 1994; Ducea et al., 2013; Murray et al., 2015).

Figure 8: Conceptual evolutionary diagram of the magmatic system in the southern Puna Plateau. **(A)** A shallowing of the angle of subduction leads to an eastward migration of the frontal arc, and underthrusting of foreland material beneath the eastern Puna Plateau to mid-crustal depths leads to anatexis and the formation of a migmatite zone. **(B)** Initial stages of large-scale delamination of the uppermost mantle lithosphere and lower crust. Adiabatic melting of the inflowing asthenosphere leads to the creation of MASH zone in backarc distinct from the

frontal arc MASH zone. Local melting of the pyroxenitic lower crust is incorporated in the MASH zone, which is reflected in the geochemical signature of the mafic deposits at the surface. **(C)** After the delaminating body is removed, arc and backarc MASH zones merge into one centralized MASH zone, while small-scale pyroxenitic drips are partially melted as they fall into the mantle. Influx of magmatic material into the mid-crust follows rheologically weak, long-lived zones of magma ascent from Stage A and B. Blue square shows approximate cross-sectional area shown in D. **(D)**. Cross-section through our velocity model along A-B shown in Fig. 5A. Depths relative to sea level. White line, Moho depth from Tassara & Echaurren (2012); CGA, Cerro Galan Anomaly; IA, Incahuasi Anomaly.

5. Conclusions

In this paper, we jointly invert CCP-derived receiver functions and dispersion data from ambient noise tomography to image a lithospheric-scale magma plumbing system beneath the volcanically active Southern Puna Plateau. Three separate slow (<3.0 km/s) velocity bodies, which we interpret to represent magma reservoirs, are imaged in the mid-crust beneath the active frontal arc of the Lazufre volcanic field and the southernmost Central Volcanic Zone, as well as beneath the Cerro Galan Caldera. These bodies have vertical and lateral extents of ~10 km and up to 75 km, while the top of these zones occurs at ~16-21 km below the surface. The Cerro Galan Anomaly is the slowest anomaly in this study and represents a magma reservoir beneath the Cerro Galan Caldera, which erupted ~1300 km³ of material from ~6 – 2 Ma. When compared to the erupted volume, this 22,000 km³ slow velocity body represents a plutonic-volcanic ratio of at least 8:1 and contains a maximum of 22% melt. Eruptions may have been driven by the influx of new mantle-derived melt from an underlying MASH zone into this partially molten mid-crustal zone. This MASH zone is imaged as a broad, N-S trending slow velocity zone (<4.1 km/s) at mantle depths with ~4-9% melt, and we interpret the spatial extent

of this MASH zone to approximately delimit the areal extent of lithospheric delamination beneath the southern Puna Plateau. Felsic melts derived from this MASH zone rise into the crust and are diverted into the frontal arc and backarc magma reservoirs. This lithospheric-scale magma system has significant implications for the creation and evolution of continental lithosphere in Cordilleran systems.

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