IMAGING VARIATIONS IN THE CENTRAL ANDEAN MANTLE AND THE SUBDUCTING NAZCA SLAB WITH TELESEISMIC TOMOGRAPHY

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SIGNED: Alissa Scire
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ABSTRACT

The Nazca-South America convergent margin is marked by the presence of the Andean mountain belt, which stretches along the 8000-km long western margin of the South American plate. The subduction zone is characterized by significant along-strike changes in both upper plate structure and slab geometry that make it an ideal region to study the relationship between the subducting slab, the surrounding mantle, and the overriding plate. This dissertation summarizes the results of three finite frequency teleseismic tomography studies of the central Nazca-South America subduction zone which improve our understanding of how along-strike variations in the Andean mountain belt and the subducting Nazca plate interact with each other and with the surrounding mantle. This is accomplished by first focusing on two smaller adjacent regions of the central Andes to explore upper mantle variations and then by using a combined dataset, which covers a larger region, to image the deeply subducted Nazca slab to investigate the fate of the slab. The first study focuses on the central Andean upper mantle under the Altiplano-Puna Plateau where normally dipping subduction of the Nazca plate is occurring (18° to 28°S). The shallow mantle under the Eastern Cordillera is generally fast, consistent with either underthrusting of the Brazilian cratonic lithosphere from the east or a localized “curtain” of delaminating material. Additional evidence for delamination is seen in the form of high amplitude low velocities under the Puna Plateau, consistent with proposed asthenospheric influx following lithospheric removal. In the second study, we explore the transition between normal and flat subduction along the north edge of the Altiplano Plateau (8° to 21°S). We find that the Peruvian flat slab extends further inland along the projection of the Nazca Ridge than was previously
proposed and that when re-steepening of the slab occurs, the slab dips very steeply (~70°) down through the mantle transition zone (MTZ). We also tentatively propose a ridge parallel tear along the north edge of the Nazca Ridge. Both of these observations imply that the presence of the Nazca Ridge is at least locally influencing the geometry of the flat slab. The final study investigates along-strike variations in the deeply subducted Nazca slab along much of the central Nazca-South America subduction zone (6° to 32°S). Our results confirm that the Nazca slab continues subducting into the lower mantle rather than remaining stagnant in the MTZ. Thickening of the slab in the MTZ north of 16°S is interpreted as folding or buckling of the slab in response to the decreased slab sinking velocities in the lower mantle.
INTRODUCTION

The Nazca-South America subduction zone is characterized by significant along-strike variations in magmatism, upper crustal shortening, crustal thickness, lithospheric structure, and slab geometry that make it an ideal region to study the relationship between the subducting slab, the mantle wedge, and the overriding plate. Since subduction zones are one of Earth’s largest recycling systems, transferring oceanic lithosphere, sediment, and water from the Earth’s surface into the mantle (Stern, 2002), understanding the dynamics and evolution of the Nazca-South America subduction zone and the interaction between the subducting and overriding plate can help us increase our understanding of the global plate tectonic cycle.

The western margin of the South American plate is marked by the presence of the Andean mountain belt, which stretches along the entire length of the modern subduction zone. The central Andes, between ~15° and 30°S, is dominated by the Central Andean Plateau (CAP), which is defined as the region above 3 km elevation. The CAP is one of the largest high plateaus in the world, second only to the Tibetan Plateau. The central part of the CAP can be divided into two internally drained segments, the Altiplano in the north, a relatively flat plateau with an average elevation of ~3.5 km, and the Puna in the south, which is characterized by higher elevations, averaging ~4.5 km, and more topographic relief (Isacks, 1988; Whitman et al., 1996). The Altiplano-Puna Plateau is bounded to the west by the modern volcanic arc, the Western Cordillera and to the east by a fold and thrust belt, the Eastern Cordillera. Deformation in the Eastern Cordillera occurred between ~40 and 15 Ma (McQuarrie et al., 2005; Oncken et al., 2006) before shifting to the Subandean Zone, the modern, thin-skinned fold and thrust belt north of
24°S, where deformation began after 10 Ma (Allmendinger et al., 1997; Oncken et al., 2006). South of 24°S, deformation in the backarc shifts from the thin-skinned deformation of the Subandean Zone to thick-skinned formation in the Santa Barbara System and finally to the basement cored uplifts of the Sierras Pampeanas (Kley and Monaldi, 2002).

The tectonic development of the Altiplano-Puna Plateau is debated, with different studies arguing for different mechanisms and timing of plateau uplift. While some studies based on structural reconstructions (e.g., McQuarrie et al., 2005; Elger et al., 2005; Oncken et al., 2006; Lamb, 2011) argue for gradual uplift of the plateau due to shortening and thickening of the crust between 40-30 and 10 Ma, paleoelevation studies (e.g. Ghosh et al., 2006; Garzione et al., 2008) argue for a rapid period of uplift since 10 Ma resulting from large scale delamination of the lithosphere. Recent work by Saylor & Horton (2014) argues that uplift of the plateau may have been spatially and temporally non-uniform, and therefore could be result from locally varying crustal shortening or localized removal of mantle lithosphere rather than uniform shortening along the entire CAP or large scale delamination of the lithosphere. Evidence for local delamination of the lithosphere has been found in several regions of the Andes in both geochemical (e.g., Kay & Kay, 1993; Kay et al., 1994; Kay & Coira, 2009) and seismic studies (e.g., Myers et al., 1998; Schurr et al., 2006; Asch et al., 2006; Koulakov et al., 2006; Heit et al., 2008; Bianchi et al., 2013; Beck et al., 2015).

The Nazca plate is between 40 and 45 Ma old when it subducts beneath South America (Müller et al., 2008). Motion of the Nazca plate with respect to South America varies along the plate margin, averaging ~8.5 cm/yr relative to a stable South American
reference frame (Somoza & Ghidella, 2012). Subduction of the Nazca plate beneath South America occurs at an angle of ~30° down to depths of about 300 km along much of the margin, with two regions of flat or shallow subduction occurring under Peru between 5° and 14°S and under central Chile and Argentina between 30° and 33°S (Cahill and Isacks, 1992; Hayes et al., 2012). The flat slab region in Peru was first identified by Barazangi & Isacks (1976, 1979). It has been estimated that the slab flattens at about 100 km depth for several hundred kilometers inland (Cahill & Isacks, 1992; Hayes et al., 2012; Phillips & Clayton, 2014) although some uncertainty in the exact geometry of the Peruvian flat slab exists due to the limited slab seismicity. The Peruvian flat slab extends ~1500 km along the Nazca-South America subduction zone from ~14°S under Peru to ~5°S where the slab steepens near the Grijalva Fracture Zone. The southern edge of the flat slab segment near 14°S corresponds to the location of the Nazca Ridge, a bathymetric high on the Nazca plate thought to have been formed by volcanism from the Easter-Salas hotspot (Ray et al., 2012). The subduction of a large oceanic plateau, termed the Inca Plateau, is proposed as an explanation for the lateral extent of the Peruvian flat slab (Gutscher et al., 1999). Subduction of this hypothetical plateau supports flat subduction in the northern section of the Peruvian flat slab, although Skinner & Clayton (2013) argue that plate reconstructions place the Inca Plateau too far east to support the flat slab. The development of regions of flat slab subduction has been linked to the subduction of overthickened oceanic crust although debate exists over the importance of this relative to other factors (e.g., Gutscher et al., 2000; van Hunen et al., 2002, 2004; Espurt et al., 2008; Martinod et al., 2013; Gerya et al., 2009; Skinner & Clayton, 2013). Regions of flat slab subduction are also characterized by an absence of arc volcanism due to the
displacement of the asthenospheric mantle wedge by the slab. Arc volcanism appears to have shut off at ~4 Ma above the present day location of the Nazca slab (Rosenbaum et al., 2005), indicating that the subducting slab flattened in this region around the end of the Miocene.

Tomographic images of deeply subducted slabs indicate that the interaction between subducting slabs and the mantle transition zone (MTZ) varies globally, with some slabs penetrating through into the lower mantle with only minor deformation in the MTZ while in other regions, slabs flatten and stagnate in the MTZ for long periods of time (e.g., Bijwaard et al., 1998; Li et al., 2008; Fukao et al., 2009). Constraints on the geometry of the deeply subducted Nazca slab in the MTZ and lower mantle have previously come from deep slab seismicity combined with teleseismic and global tomography studies. Both Engdahl et al. (1995), who used teleseismic and regional P phases to image the South American subduction zone between 5° and 25°S down to ~1400 km depth, and global tomography studies (e.g., Bijwaard et al., 1998; Zhao 2004; Li et al., 2008; Fukao et al., 2001, 2009; Zhao et al., 2013) observe that the Nazca slab continues subduction into the lower mantle although the high velocity anomaly associated with the slab becomes more amorphous below the MTZ. However, both Engdahl et al. (1995) and global tomography studies (Bijwaard et al., 1998; Zhao 2004; Li et al., 2008; Fukao et al., 2001, 2009; Zhao et al., 2013) note thickening of the Nazca slab in the MTZ, consistent with some deformation and temporary stagnation of the slab in the transition zone before it resumes subduction into the lower mantle.

In this dissertation I present the results of three finite frequency teleseismic tomography studies which image the mantle under the central South American Andes.
The first two studies use different datasets to image the upper mantle between 18° and 28°S where normally dipping subduction of the Nazca plate is occurring (Appendix A) and the region under northwestern Bolivia and southern Peru (8° and 21°S) where the angle of subduction changes from normally dipping subduction in the south to flat subduction in the north (Appendix B). The third study (Appendix C) combines the datasets used in Appendices A and B in order to increase the aperture of the array allowing us to resolve anomalies at 895 km depth so that we can image the Nazca slab in the lower mantle to offer new constraints on the fate of the Nazca slab.

The mantle beneath the central Andes of South America has been imaged using local (e.g., Myers et al., 1998; Graeber and Asch, 1999; Schurr et al., 2006; Koulakov et al., 2006) and teleseismic (e.g. Dorbath et al., 1993; Engdahl et. al, 1995; Heit et al., 2008; Bianchi et al., 2013) tomography. While each of these studies has provided new insights into the dynamics of the subduction system, they have focused on the region defined by the individual footprint of the seismic array used. In Appendix A, I present results from a teleseismic P-wave tomography study which combined data from eleven temporally and spatially distinct networks deployed in the central Andes between 1994 and 2009. By using data from these networks we were able to obtain a more detailed regional scale model of the central Andean mantle between 18° and 28°S. The increased array aperture resulting from the combined data set allowed us to image the mantle to the base of the mantle transition zone at 660 km depth.

The north central Andes between 8° and 21°S is characterized by a change in the geometry of the subducting slab, shifting from normal (~30° dip) subduction in the south to near horizontal subduction in the north. We deployed two arrays of seismic stations,
the CAUGHT (Central Andean Uplift and Generation of High Topography) array in northwestern Bolivia and southern Peru, and the PULSE (PerU Lithosphere and Slab Experiment) array which was deployed over the southern part of the Peruvian flat slab, to investigate the structure of the mantle and the subducting Nazca plate in this region. In Appendix B, we use finite frequency teleseismic P- and S-wave tomography to investigate the transition from flat to normal subduction and its effect on the Nazca plate and the surrounding mantle. Since previous seismological studies in this region have been limited by the number of stations, our work provides new constraints on the geometry of the subducting Nazca slab.

The interaction between subducting slabs and the transition between the upper and lower mantle varies globally, with global tomography studies imaging some slabs penetrating into the lower mantle with only minor broadening in the MTZ while other subducting slabs flatten out and stagnate in the MTZ over long distances (e.g., Bijwaard et al., 1998; Li et al., 2008; Fukao et al., 2009). By combining the datasets from the fourteen separate arrays used in the studies discussed in Appendices A and B, we are able to increase the maximum depth of resolution and image the Nazca slab in the lower mantle. In Appendix C, we explore the morphology of the subducting Nazca slab in the MTZ and the lower mantle between 6° and 32°S down to depths of ~900 km using teleseismic P-wave tomography. This allows us to offer new constraints on the interaction between the subducting Nazca slab and the boundary between the upper and lower mantle.
SUMMARY OF WORK

The methods, results, and conclusions of my work are presented in the three papers that are appended to this dissertation. I present a summary of the most important findings of each study below.

The three studies use finite frequency teleseismic tomography to image the mantle under the South American Andes to varying depths. The original tomography technique of Aki et al. (1977) assumes that sampling occurs only along the infinitesimally thin theoretical ray path which represents arrivals of infinite frequency. This high frequency approximation was appropriate for first-arrival picks from band-limited data available in those early years and convenient for computational reasons when formulating the tomographic problem. The broader band data we use today and the arrivals we calculate by cross-correlation represents the traveltime of a finite-frequency seismic wave leading to the need for a more realistic approximation. The finite-frequency approximation defines the sampled area around the geometrical ray path by the first Fresnel zone, the width of which is dependent on both frequency and the distance along the ray path (Hung et al., 2000; Dahlen et al., 2000). The sampling in each model layer is determined by the width of the Fresnel zone, which increases with distance from the source or receiver, and the differential sensitivity within the Fresnel zone, resulting in theoretical “banana-doughnut” sensitivity kernels (Dahlen et al., 2000). Since the zone of sampling for each ray path is determined by the sensitivity kernels, multiple nodes in each model layer are sampled by a single ray rather than just the single node that lies along the geometrical ray path, increasing sampling of the model space over tomography algorithms that assume that sampling only occurs along the ray path. A detailed discussion of the specific
algorithm used to calculate the sensitivity kernels can be found in Schmandt & Humphreys (2010).

In Appendix A, my coauthors and I combined data from eleven temporally and spatially distinct seismic networks that were deployed in the central Andes between 1994 and 2009. We used data from 284 short period and broadband seismic stations which were deployed in the central Andes at various times. The dataset includes stations from ten temporary deployments, the permanent Plate Boundary Observatory network in Chile, and two permanent stations from the Global Seismograph Network and the Global Telemetered Seismograph Network. This dataset allowed us to use finite frequency teleseismic P-wave tomography to image velocity anomalies in the mantle from 95-660 km depth between 18° and 28°S. Since many studies of the central Andes have either looked at a small region under a single seismic network, which limits their maximum depth of resolution, or imaged the region as part of a global study, which limits their ability to resolve smaller scale features, the work presented in Appendix A offers a new regional scale view of the upper mantle in this part of the central Andes.

Significant along-strike variations in upper mantle structure are observed under the central Andes in our tomographic images. Fast anomalies under the Eastern Cordillera between 16° and 26°S is consistent with underthrusting of the Brazilian cratonic lithosphere from the east, which has previously been proposed to explain similar observations of anomalously high velocities under the Subandean zone (Myers et al., 1998; Beck and Zandt, 2002; Phillips et al., 2012). Alternately, a second possible cause of the fast velocity anomalies is that they represent a “curtain” of foundering lithospheric material. Low velocities anomalies at 95 and 130 km depth are observed beneath the
southern Puna Plateau (Anomaly D, Fig. 2), consistent with proposed asthenospheric influx following lithospheric delamination. Several other discontinuous low velocity anomalies are observed beneath the Altiplano to the north, one of which, Anomaly C, corresponds to the surface location of the Los Frailes Volcanic Field (Fig. 2). The southeast edge of the Puna Plateau is marked by fast velocity anomalies at 95 km depth that we argue could be the edge of the thicker mantle lithosphere of the Precambrian Pampean Terrane proposed by Ramos et al. (1986), which is thought to comprise the crystalline basement of the Sierras Pampeanas. The edge of this fast anomaly corresponds to the surface boundary of the Sierras Pampeanas.

Variations in the width of the slab anomaly in the deeper parts of the tomographic model show evidence for deformation of the slab between 300 and 660 km (Fig. 2). Thickening of the slab in the mantle transition zone (MTZ) is observed in several places, in agreement with the idea that the Nazca slab stagnates at least temporarily in the MTZ before resuming subduction into the lower mantle. Our study also images a sub-slab low velocity anomaly in the MTZ between 22° and 28°S (Anomaly G, Fig. 2), which is similar to those seen in other subduction zones, and is interpreted as either a local thermal anomaly or a region of hydrated material in the MTZ.

In Appendix B, we investigate the subducting Nazca plate and surrounding mantle in the transition from normally dipping subduction under northwestern Bolivia to flat subduction under southern Peru. We deployed two arrays of broadband seismic stations between 2010 and 2013, the CAUGHT (Central Andean Uplift and the Geodynamics of High Topography) array, which consisted of 48 stations in northwestern Bolivia and southern Peru, and the PULSE (PerU Lithosphere and Slab Experiment) array, which
consisted of a further 38 stations deployed in Peru, primarily over the southern part of the Peruvian flat slab. In Appendix B we used finite frequency teleseismic P- and S-wave tomography to image the mantle between 8° and 21°S from 95 to 660 km depth. Since previous seismic studies in northwestern Bolivia and southern Peru have been limited in their ability to image the deeper mantle under the region, our work offers new constraints on the location and geometry of the Nazca slab and the effect of the changing slab dip on the slab and surrounding mantle.

North of 14°S, we image the Nazca slab to the east of its predicted location from the Slab1.0 global subduction zone contours (Hayes et al., 2012) where it steepens inboard of the Peruvian flat slab region. This indicates that they slab remains flat further inland than previously thought. Our results show that when the slab steepens and resumes subduction into the mantle, it does so at a very steep angle (~70°) from about 150 km depth to the top of the mantle transition zone at 410 km depth, where we observe thickening of the slab in the MTZ similar to that seen further south, which is discussed in Appendix A. This region of slab steepening corresponds with the predicted edge of the subducted Nazca Ridge (Hampel, 2002), implying that the geometry of the Nazca slab is at least locally being influenced by the presence of the ridge. We hypothesize the existence of a slab tear along the north side of the Nazca Ridge at 10°S between 71° and 73°W (Fig 3) although due to limitations of our resolution to the north of the subducted Nazca Ridge, further investigation of this hypothesized tear will rely on future work. We offer slab contours for the Nazca slab between 8° and 20°S from 100 to 660 km depth (Fig. 3a) based on a combination of our results, intermediate depth seismicity from the PDE catalog (Engdahl et al., 1998), the Slab1.0 global subduction zone model (Hayes et
al., 2012), and initial constraints on the location of the top of the Peruvian flat slab north of 14°S from receiver functions (Bishop et al., 2013). The geometry and location of the change in slab dip from 100 to 200 km depth just to the south of the Nazca Ridge is defined using earthquake relocations from work by Schneider & Sacks (1987).

The sub-slab mantle immediately under the projection of the subducted Nazca Ridge is dominated by a high amplitude low velocity anomaly (Anomaly E, Fig. 3b) which we have interpreted as either the result of limited partial melting due to sub-slab flow or as asthenospheric mantle at shallow depths due to the existence of thinned oceanic lithosphere under the Nazca Ridge. The shallow mantle to the south of the Peruvian flat slab where the Nazca slab subducts at a normal angle is characterized by fast anomalies under the Eastern Cordillera (Anomaly C, Fig. 3b), which are bounded to the east by low velocity anomalies at 100 km depth. The presence of these low velocity anomalies suggests that Anomaly C is unlikely to represent undisrupted cratonic lithospheric material which has been underthrust from the east. As was proposed in Appendix A as an alternative hypothesis to explain a similar fast anomaly, Anomaly C could be related to foundering of cold lithospheric material, an interpretation that implies that mantle lithosphere is being removed along much of the central Andes. A second fast anomaly under the Subandean Zone near 15°S (Anomaly D, Fig. 3) is interpreted as a vertically elongated delaminating block of lithosphere.

In Appendix C we focus on the deeper structure of the Nazca slab. By combining the datasets used for the studies discussed in Appendices A and B, we are able to increase our maximum depth of resolution to image the Nazca slab in the MTZ and lower mantle to depths of ~900 km from 6° to 32°S. We use finite frequency teleseismic P-wave
tomography to image along-strike variations in the morphology of the subducting Nazca slab as it interacts with the boundary between the upper and lower mantle. Our images of the Nazca slab in the transition zone and lower mantle provide new information about the deformation of the slab.

Significant thickening of the slab in the MTZ is observed in north of 16°S (Fig. 4) which we argue is due to temporary stagnation and possibly folding or buckling of the slab in the MTZ in response to a decrease in the sinking velocity of the slab in the lower mantle associated with the viscosity increase across the boundary between the upper and lower mantle. Little to no thickening of the slab anomaly is observed in the MTZ south of 16°S, indicating that deformation of the slab varies along-strike. Similar along-strike variations are observed in the lower mantle. The slab anomaly in the lower mantle to the north of 16°S has higher amplitudes and appears to be more coherent than the slab anomaly to the south where less thickening of the slab anomaly was observed in the MTZ (Fig. 4). This shift in slab morphology in the MTZ and lower mantle appears to correspond to a change in the dip of the slab as it enters the MTZ at 410 km depth. Our results indicate that along-strike variations in the subduction geometry have a significant effect on the deformation and the fate of the slab at depth.
Figure 1. Map of South America with seismic station locations for individual networks used and the three study regions (Appendices A-C) marked by labelled boxes. Note that not all stations shown where available for the first study (Appendix A). Boundaries of geomorphic provinces (dashed lines) are modified from Tassara (2005). WC: Western Cordillera, AP: Altiplano, EC: Eastern Cordillera, SA: Subandean Zone, PN: Puna, SB: Santa Barbara System, SP: Sierras Pampeanas. Red triangles mark location of Holocene volcanoes (Siebert & Simkin, 2002). Slab contours (gray) are from the Slab1.0 global subduction zone model (Hayes et al., 2012). Plate motion vector from Somoza & Ghidella (2012). Earthquake data are from 1973 to 2012 (magnitude > 4.0); U.S. Geological Survey – National Earthquake Information Center (NEIC) catalog.
Figure 2. 3-D diagram of the resolved subducting Nazca slab and prominent mantle low velocity anomalies from tomographic models in Appendix A. The isosurfaces for this model are obtained by tracing the most coherent low velocity anomalies (less than negative 3%) and slab related (greater than positive 3%) coherent fast anomalies in the tomographic model. Slab geometry above 200 km is determined entirely from Slab1.0 (Hayes et al., 2012). Anomalies B, C, D, and G are discussed in the text. Boundaries of geomorphic provinces as in Figure 1.
Figure 3. A) Slab contours determined from tomographic models in Appendix B. Red contours are continuous slab contours while yellow contours are disrupted by the hypothesized tear north of the Nazca Ridge. Red triangles mark location of Holocene volcanoes (Siebert & Simkin, 2002). Gray outline marks projection of the subducted Nazca Ridge from Hampel (2002). B) 3-D diagram of the resolved subducting Nazca slab and prominent mantle low velocity anomalies inferred from tomographic models in Appendix B. The isosurfaces for this diagram are obtained by tracing the most coherent low velocity anomalies (less than negative 3%) and slab related and other (greater than positive 3%), coherent fast anomalies in the tomographic model. Geomorphic provinces (fine dashed lines) are the same as in Figure 1. Heavy black outline marks projection of the subducted Nazca Ridge from Hampel (2002). Anomalies C, D, and E are discussed in the text.
Figure 4. Trench perpendicular cross sections through the tomography model from Appendix C. Dashed lines mark the edge of resolution. Yellow dots are earthquake locations from the EHB catalog (Engdahl et al., 1998). Solid black line marks the top of the Nazca slab from the Slab1.0 model (Hayes et al., 2012).
REFERENCES


APPENDIX A: IMAGING THE NAZCA SLAB AND SURROUNDING MANTLE TO 700 KM DEPTH BENEATH THE CENTRAL ANDES (18° TO 28°S)

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ABSTRACT

The central Andes in South America is an ideal location to investigate the interaction between a subducting slab and the surrounding mantle to the base of the mantle transition zone (MTZ). We used finite-frequency teleseismic P-wave tomography to image velocity anomalies in the mantle from 100 - 700 km depth between 18° and 28°S in the central Andes by combining data from eleven separate networks deployed in the region between 1994 and 2009. Deformation of the subducting Nazca slab is observed in the MTZ, with regions of both thinning and thickening of the slab that we suggest are related to a temporary stagnation of the slab in the MTZ. Our study also images a strong low velocity anomaly beneath the Nazca slab in the MTZ, which is consistent with either a local thermal anomaly or a region of hydrated material in the MTZ. The shallow mantle (<165 km) under the Eastern Cordillera is generally fast, consistent with proposed underthrusting of the Brazilian cratonic lithosphere or a string of localized lithospheric foundering. Several discontinuous low velocities anomalies are observed beneath parts of the Altiplano and Puna Plateaus including two strong low velocity anomalies in the upper mantle under the Los Frailes Volcanic Field and the southern Puna Plateau, consistent with proposed asthenospheric influx following lithospheric delamination.

INTRODUCTION

The mantle beneath the central Andes of South America has been investigated using both local (e.g., Myers et al., 1998; Graeber and Asch, 1999; Schurr et al., 2006; Koulakov et al., 2006) and teleseismic (e.g. Dorbath et al., 1993; Engdahl et. al, 1995;
Heit et al., 2008) tomography but each study was focused on a region defined by the individual footprint of the seismic array used. This has limited the depth of resolution of these teleseismic tomography studies in this region. More recently Bianchi et al., (2013) combined teleseismic and local data to image part of the central Andes. Global tomography studies (e.g., Bijwaard et al., 1998; Zhao 2004; Li et al., 2008; Fukao et al., 2001, 2009) provide images beneath South America down to lower mantle depths, however global tomography studies cannot provide detailed images of upper mantle structure. Multiple temporary seismic networks and several permanent stations have been deployed in the central Andes, which allows us to combine data from these temporally and spatially distinct networks in a single study and make it possible to obtain a more detailed regional scale model of the mantle. We used finite-frequency teleseismic P-wave tomography to image velocity anomalies in the mantle down to the base of the mantle transition zone at 660 km between 18° and 28°S in the central Andes. Nonuniform distribution and spacing of denser networks in the broad span of the studied region results in coarser resolution in the shallower parts of the model but allows for regional scale anomalies to be imaged and interpreted down to greater depths than smaller local studies.

THE CENTRAL SOUTH AMERICAN ANDES

The central Andes are characterized by significant along-strike variations in magmatism, upper crustal shortening, crustal thickness, and lithospheric structure (Allmendinger et al., 1997; Kay and Coira, 2009). The Central Andean Plateau (CAP) is generally defined as the area above 3 km elevation and is one of the largest high plateaus in the world, second only to the Tibetan Plateau (Figure 1A). The central portion of the
CAP includes two internally drained segments, the Altiplano to the north, a high, relatively flat plateau, and the Puna to the south, a region with a higher elevation and more topographic relief (Isacks, 1988; Whitman et al., 1996). The Altiplano has an average elevation of ~3.5 km while the elevation of the Puna averages ~4.5 km. The edges of the plateau are defined to the west by the modern volcanic arc, the Western Cordillera, and to the east by the Eastern Cordillera, an inactive fold and thrust belt where shortening occurred from ~40 Ma to 15 Ma (McQuarrie et al., 2005; Oncken et al., 2006). Deformation in the Subandean zone east of the Eastern Cordillera began after ~10 Ma in Bolivia and is characterized by a thin-skinned fold and thrust belt north of 24°S (Allmendinger et al., 1997; Oncken et al, 2006). South of 24°S, deformation in the backarc shifts from the thin-skinned deformation of the Subandean zone to thick-skinned deformation in the Santa Barbara System and finally to the basement cored uplifts of the Sierras Pampeanas (Kley and Monaldi, 2002). Receiver function studies (Yuan et al., 2000, 2002; Beck and Zandt, 2002; Wolbern et al., 2009) show a variable crustal thickness across the central Andes, with ~35 km thick crust in the forearc thickening to 70 km in the Western Cordillera. The crust under the Altiplano varies in thickness from 60 to 70 km with an average of ~65 km which then thickens locally to 75 km in the Eastern Cordillera. The crust under the Puna Plateau to the south appears to be slightly thinner, with an average crustal thickness of about 60 km, in spite of the higher average elevation of the Puna relative to the Altiplano. Under the Subandean zone the crust is ~40 km thick thinning to 30-35 km out on the Chaco Plain beyond the eastern limits of Andean deformation.
The subducting Nazca plate age is ~45 Ma at the trench as it enters the subduction zone beneath the central Andes (Müller et al., 2008), and is converging with the South American plate at 61 mm/yr relative to a stable South American reference frame (Norabuena et al., 1999). Subduction of the Nazca plate under the central Andes occurs at an angle of ~30° to a depth of about 300 km as defined by the slab seismicity (Figure 1B; Cahill and Isacks, 1992; Hayes et al., 2012) and is bracketed by the Peruvian flat slab region to the north and the Pampean flat slab region to the south. While there are few earthquakes between 300 and 500 km some earthquakes do occur near the base of the MTZ between 500 and 660 km (Figure 1B). Local earthquake tomography studies between 21° and 24°S indicate that seismic velocities in the forearc are fast, possibly resulting from the presence of plutonic remnants of earlier volcanic arcs (Graeber and Asch, 1999; Koulakov et al., 2006; Schurr et al., 2006). Generally the thick crust of the Altiplano and Puna Plateaus has been imaged with low average crustal P and S-wave velocities (Swenson et al., 2000; Ward et al., 2013). The presence of a low velocity zone in the upper crust of the Altiplano, the Altiplano Low Velocity Zone, is observed in several studies and it is attributed to the presence of a small amount of partial melting of the crust at ~20 km depth (Yuan et al., 2000; Beck and Zandt, 2002; Ward et al., 2013). Further south in the Puna there is a major low velocity zone (~15-20 km) associated with the Altiplano-Puna Volcanic Complex (APVC) known as the Altiplano-Puna Magma Body (APMB) that is thought to be the source region for the extensive 11-1 Ma ignimbrite deposits at the surface (Chmielowski et al., 1999; Zandt et al., 2003; De Silva et al., 2006; Ward et al., 2013). Anomalously low velocities have been observed in upper mantle under the southern Puna Plateau between 25° and 27°S (e.g. Bianchi et al., 2013;
Wolbern et al., 2009; Schurr et al., 2006; Koulakov et al., 2006) and combined with geochemical analyses of the large ignimbrite deposits of the southern Puna volcanic field (e.g. Kay and Kay, 1993; Kay et al., 1994; Kay and Coira, 2009) have been used as evidence for the influx of asthenosphere following lithospheric delamination in the region. Underthrusting of Brazilian cratonic lithosphere has been proposed to explain anomalously high velocities under the Subandean zone (Myers et al., 1998; Beck and Zandt, 2002; Phillips et al., 2012). Similar observations of high velocity anomalies near 16°S (Dorbath et al., 1993; Dorbath and Granet, 1996) and analyses of focal mechanisms further north at ~11°S (Dorbath et al., 1991) are also interpreted as evidence for cratonic underthrusting along much of the central Andes. Continuous removal of mantle lithosphere under the Altiplano/Eastern Cordillera boundary is suggested to accommodate the underthrusting of the Brazilian craton between 18° and 20°S (McQuarrie et al., 2005; Myers et al., 1998; Beck and Zandt, 2002).

Tectonics of the upper plate – Development of the Altiplano-Puna Plateau

The tectonic development of the plateau is debated, with different studies arguing for different uplift histories and mechanisms. Original estimates of trench perpendicular shortening appeared to be unable to fully explain the thickness of the Altiplano crust (Kley and Monaldi, 1998) without along-strike contributions to crustal thickening although more recent studies dispute this claim (e.g. Lamb, 2011). The timing of plateau uplift is uncertain with structural reconstructions implying that plateau development was controlled by gradual shortening and thickening of the crust between 40-30 and 10 Ma (e.g. McQuarrie et al., 2005; Elger et al., 2005; Oncken et al., 2006; Lamb, 2011) while
paleoelevation studies (e.g. Ghosh et al., 2006; Garzione et al., 2008) argue for a more recent (<10 Ma), rapid uplift of the plateau resulting from large-scale delamination of the lithosphere. Based on the geochemistry of the volcanic rocks delamination of the lithosphere has been proposed as an explanation for the existence of high plateaus, particularly the Puna Plateau (Kay and Kay, 1993; Kay et al., 1994; Kay and Coira, 2009). Evidence for delamination from seismic studies exists in several regions of the central Andes. Seismic tomography studies have imaged high velocity anomalies in the mantle hypothesized to be delaminated blocks of lithosphere (Schurr et al., 2006; Asch et al., 2006; Koulakov et al., 2006; Bianchi et al., 2013) around 23°S while other studies hypothesize localized lithospheric removal based on observations of anomalously low velocities below the crust which are interpreted as asthenospheric mantle at shallow depths (Myers et al., 1998; Heit et al., 2008).

_Tectonics of the lower plate – Slab structure and the fate of the slab_

Although the subducting Nazca slab has been imaged in small-scale teleseismic tomography studies and in global tomography studies, few studies imaged the slab on a regional scale until recently (e.g., Pesicek et al., 2012; Bianchi et al., 2013). Global tomography studies show the Nazca slab in the central Andes descending into the lower mantle, although some studies indicate at least temporary stagnation in the mantle transition zone (e.g., Bijwaard et al., 1998; Zhao 2004; Li et al., 2008; Fukao et al., 2001, 2009; Zhao et al., 2013). Unlike the stagnation of the Pacific slab under Japan, where the slab is imaged extending horizontally for several hundred kilometers in the transition zone with little evidence for it resuming subduction into the lower mantle (Fukao et al.,
2009), the Nazca slab appears to resume subduction into the lower mantle after a period of thickening and stagnation in the MTZ. Receiver function studies by Liu et al. (2003) and modeling studies by Quinteros and Sobolev (2013) indicate a correlation between abnormal MTZ thicknesses and probable resistance to subduction of the Nazca plate into the lower mantle which agrees with the images of the temporarily stagnated Nazca slab from global tomography studies discussed above. In the lower mantle, the high velocity anomaly associated with the slab becomes more amorphous (Bijwaard et. al, 1998; Fukao et al., 2001; Li et al., 2008; Zhao et al., 2013). South of our study area a nested regional-global tomography study by Pesicek et al. (2012) imaged a discontinuous Nazca slab at ~38°S, providing evidence for vertical tearing of the slab along the zone of weakness provided by the subduction of the Mocha Fracture Zone. Engdahl et al. (1995) used teleseismic and regional P phases to image the Andean subduction zone between 5° and 25°S down to ~1400 km depth. Their results indicate that the Nazca slab penetrates into the lower mantle throughout their study area, in agreement with global models. In the southern part of their study area, thickening of the slab in the MTZ is observed, indicating at least temporary stagnation of the slab before it penetrates into the lower mantle. A teleseismic tomography study by Heit et al. (2008) along a line at 21°S had difficulties resolving the high velocity slab anomaly without it being shifted vertically relative to the predicted location of the slab based on the observed slab seismicity from the EHB catalog (Engdahl et al., 1998) and used the earthquake location data to constrain the location of the slab in their final model. Similarly, Bianchi et al. (2013) used regional and global tomography data to constrain the location of the Nazca slab in their study which images the lower crust and upper mantle down to depths of ~300 km under the
Puna Plateau between 24° and 29°S. They observe some variations in the amplitude of the slab anomaly, with a distinct low velocity anomaly seen dividing the slab at ~26°S between 100 and 200 km depth. They note that this feature is not well constrained and may be the result of vertical smearing of low velocity anomalies in the crust down into the mantle, resulting in the discontinuous slab anomaly.

**DATA**

The data used in this study were collected by 284 short period and broadband seismic stations belonging to eleven separate networks deployed at various time periods between 1994 and 2009 (Figure 1A; Table 1). The networks include nine temporary deployments with varying numbers of stations as well as five stations from the GeoForschungsZentrum’s (GFZ) global temporary GEOFON network. The dataset also includes data from permanent stations of the Plate Boundary Observatory network, which were deployed before late 2009. Two permanent stations from the Global Seismograph Network and the Global Telemetered Seismograph Network, LVC and LPAZ respectively, are also included.

Arrival time data was used for direct P phase arrivals for 250 teleseismic events with magnitudes greater than 5.0 between 30° and 90° away from the study region. PKIKP phase arrival time data from an additional 58 global events with a similar magnitude limit between 155° and 180° away from the study region were also used (Figure 2A). Arrivals were picked using the multi-channel cross correlation technique described by VanDecar and Crosson (1990) and modified by Pavlis and Vernon (2010). A total of 12126 direct P and 2671 PKIKP arrivals were picked in three frequency bands.
for the broadband stations with corner frequencies of 0.2 to 0.8 Hz, 0.1 to 0.4 Hz, and 0.04 to 0.16 Hz. Due to the limited frequency content of data from short-period stations, arrival times are picked in a single higher-frequency band with corner frequencies of 0.5 to 1.5 Hz. The distribution of the number of rays per frequency band is dominated by the 0.2 to 0.8 Hz band, with 38% of total rays occurring in that frequency band. Rays are well distributed between the other three frequency bands, with 18% of total rays in the 0.1 to 0.4 Hz band, 17% in the 0.04 to 0.16 Hz and 27% in the 0.5 to 1.5 Hz band. The azimuthal distribution of the arrivals used in this study shows a bimodal distribution of earthquakes, with the majority of events occurring in either the Middle America subduction zone or the South Sandwich subduction zone (Figure 2B). Residual travel times were calculated relative to IASP91 (Kennett and Engdahl, 1991) and then demeaned for each event to calculate the relative travel time residuals.

Crustal thickness estimates from Tassara and Echaurren (2012), were incorporated to calculate travel time corrections for the relative residuals in order to compensate for crustal heterogeneity (Schmandt and Humphreys, 2010). Velocity estimates for the central Andean crust are sparse, with only general averages existing for the entire crust so a homogenous layered crustal velocity model corresponding to the average velocity of the Altiplano crust was used to calculate the crustal corrections for stations in the Altiplano and Puna rather than a more elaborate crustal velocity model (Zandt et al., 1996; Swenson et al., 2000; Dorbath and Masson, 2000). Crustal corrections for stations in the forearc were calculated using a faster velocity model based on local tomography studies (Graeber and Asch, 1999; Koulakov et al., 2006; Schurr et al., 2006).
METHOD

The original tomography technique of Aki et al. (1977) assumes that seismic waves sample only along the infinitesimally thin ray path representing arrivals of infinite frequency. While this is convenient for practical purposes when formulating the tomographic problem, the data we use and the arrivals we pick have limited frequency content leading to the need for a more realistic approximation. This is achieved by the finite-frequency approximation, which allows us to define frequency dependent volumes around the geometrical ray path that are sampled by each arrival (Hung et al., 2000; Dahlen et al., 2000). This zone of sampling around the theoretical ray path is defined by the first Fresnel zone, whose width is dependent on both frequency and distance along the ray path. The radius of the Fresnel zone increases with distance from the source or receiver, resulting in theoretical 3-D “banana-doughnut” sensitivity kernels (Dahlen et al., 2000). The details of the algorithm used in this study to approximate the sensitivity kernels are discussed in Schmandt and Humphreys (2010). Due to differential sensitivity within the Fresnel zone, with zero sensitivity right along the theoretical ray path, sampling at each model layer is determined by the radius of the Fresnel zone at that depth and frequency and the “doughnut”-shaped sensitivity kernel that defines where the greatest sensitivity is within the Fresnel zone. Since the zone of sampling for each model layer and for each arrival is determined by the sensitivity kernels, multiple nodes can be sampled in each layer by a single arrival rather than the single node that lies along the geometrical ray path. The use of finite-frequency tomography improves sampling of the model space and recovery of the amplitude of velocity anomalies in the modeled region over ray theory based techniques. In a global tomography study, Montelli et al. (2004)
noted a 30-50 percent increase in the amplitude of velocity perturbations recovered by a finite-frequency tomographic inversion over a ray-theory based tomographic inversion.

The frequency-dependent relative travel time residuals are inverted using LSQR algorithm of Paige and Saunders (1982) in order to obtain velocity perturbations within the modeled volume. This algorithm aims to obtain the minimum energy/length model that satisfactorily explains the observed data. In the inversion process we also incorporated station and event terms in order to compensate for receiver side perturbations that are not taken into account in the crustal corrections and event side velocity perturbations that remain outside the modeled volume. Since a homogenous-layered velocity model is used to calculate the crustal corrections, due to the lack of detailed information about the velocity structure in the study region, the station terms address local variations in the velocity structure from the homogenous velocity model as well as errors in the a priori crustal thicknesses. Event terms are calculated to account for differences in the mean arrival time related to variations in mean velocity structure under the specific subset of stations that record a given event. Since this study incorporates data from different arrays which are not all deployed contemporaneously, different subsets of stations record different events. This potentially introduces errors in calculated mean travel time residuals related to variations in the velocity structure in different parts of the study region which are also addressed by the calculation of the event terms. The inverse problem is regularized using norm and gradient damping (to account for depth dependent theoretical ray path location uncertainty) as well as model smoothing, details of which are discussed in Schmandt and Humphreys (2010). A tradeoff analysis (Figure 3) was performed between the variance reduction and the Euclidean model norm to choose the
overall damping and smoothing weights. Using the chosen damping \( (6) \) and smoothing \( (5) \) parameters results in a variance reduction of 75.0%.

The model space was parameterized into a series of nodes in a non-uniform grid which increases in size with depth and distance from the center of the model. The nodes are spaced 40 km apart in the densest sampled portions of the model with node spacing increasing to 55 km at the edges of the model space. The vertical distribution of nodes increases from 35 km in the shallowest parts of the model to 55 km in the deepest parts of the model. Node spacing within each horizontal layer also dilates with depth, with the nodes in the center part of the model increasing from 40 km spacing in the uppermost model layers to 56 km spacing in the lowermost model layers. Node spacing on the outer edges of the model increases from 55 km in the uppermost model layers to 78 km in the lowermost model layers.

Sampling of the model space was determined by calculating normalized hit quality maps (Supplemental Material, Figure S1). The calculation of the hit quality maps relies on the idea that better sampling of a node is achieved with the intersection of rays from multiple azimuths (Schmandt & Humphreys, 2010). A node which has been sampled by multiple rays from all four of the geographical quadrants is assigned a hit quality of 1 while a node which is not sampled at all has a hit quality of 0. The hit quality maps for our dataset indicate that most nodes are reasonably well sampled down to \(~660\) km depth although in deeper layers overall hit quality decreases as spreading of the rays with depth limits the number of crossing rays in the model space. Sampling of the uppermost layer (60 km depth) is strongly dependent on the distribution of stations as noted by Biryol et al. (2011) and absorbs any errors in crustal corrections and has
therefore been ignored in any interpretations. Similarly, the lowermost layer in the model (715 km) has also been removed from any interpretations because variations in lower mantle structure not accounted for in the global model used to calculate travel times outside of the model are absorbed into the lowermost layer. The well-sampled parts of the forearc in the uppermost model layers are limited to between 18° and 24°S because of a lack of stations in the forearc beyond those latitudes which were deployed contemporaneously with the networks used in this study. The extent of the well-sampled region in the eastern part of the study area is highly variable because of the distribution of stations and does not extend out into the stable foreland beyond the easternmost zone of Andean-related deformation.

*Synthetic resolution tests*

In addition to calculating the hit quality maps, we also performed a series of synthetic tests on our dataset to investigate our resolution in our model. Initial synthetic tests used a “checkerboard” defined by alternating fast and slow velocity anomalies defined in 8 node cubes which span two model layers (Figure 4). Since node spacing dilates with depth, the location and size of the anomalies in the synthetic input also change with depth. Fast and slow anomalies for the synthetic input are defined as +5 and -5 percent anomalies respectively. These anomalies are separated vertically by neutral background layers, which have no velocity anomalies. About 60% of the input amplitude of the anomalies is recovered by the inversion. The checkerboard tests show good lateral resolution throughout much of the model space (Figure 4). Lateral smearing increases and amplitude recovery decreases towards the edges of the model space. Neutral layers
(60, 95, 200, 240, 365, 410, 555, and 605 km depth) show evidence of vertical smearing as low amplitude velocity anomalies are being incorrectly resolved in the neutral layers. Additional synthetic recovery tests were performed to check on the ability of our inversion to resolve the subducting slab since the Nazca slab is expected to be a prominent feature in our study region. A synthetic slab based on the Slab1.0 global subduction zone model (Hayes et al., 2012) with a +5 percent velocity anomaly was input into the inversion to test how well the model is able to resolve a continuous, dipping, high velocity anomaly corresponding to the Nazca slab (Figure 5). The synthetic slab anomaly is ~70-100 km thick with variations resulting from the dilation of the grid-spacing with depth. Recovery of the slab anomaly is affected by the vertical smearing along the ray paths, particularly in the shallower parts of the model. Cross sections through the synthetic model show the slab being shifted vertically along the ray paths in the shallower parts of the model, resulting in the shift of the resolved slab anomaly to the east in the shallowest layers. The resolved slab anomaly is more diffuse in the uppermost parts of the model, with the vertical smearing resulting in lower amplitude (+1-2 percent) positive velocity anomalies spread out in the model space rather than a distinctly resolved slab (Figure 5). The resolved slab anomaly becomes more distinct with depth. In deeper parts of the model, ~60% of the input amplitude (+3 percent) is recovered and little distortion of the shape of the input slab anomaly is observed, indicating that the slab should be well-resolved below ~200 km.
Incorporating a priori slab data into the model

In order to improve our image of the uppermost mantle above the slab where the synthetic slab recovery tests (Figure 5) indicate that the slab is not always well imaged, we constrain the location of the slab in our model using a priori information. Using the same method as Heit et al. (2008), we subtract the effect of the Nazca slab from the observed data using a priori information about the slab location to calculate theoretical residuals for rays travelling through a +4% fast synthetic slab model with the same geometry as the synthetic model shown in Figure 5, which is based on the Slab1.0 global subduction zone model (Hayes et al., 2012). The synthetic slab model used by Heit et al. (2008) had the same amplitude (+4%) as our synthetic slab but was based entirely on EHB earthquake locations. Since our model images the mantle under the Andes to a depth beyond the intermediate depth seismicity, the use of the global subduction zone model was necessary since it offers some constraints on the location of the slab below the deepest intermediate depth earthquakes. The ability to remove the effect of the slab from the observed data is a result of the linear nature of the tomographic inversion, and is discussed in detail by Heit et al. (2008). The final model superimposes the results of the “no-slab” inversion with the model of the slab, resulting in the tomograms seen in Figures 6 and 7. This model is used for interpretations of anomalies above 200 km where synthetic tests indicated that the vertical smearing of the slab anomaly is the most prominent. We recognize that there is some uncertainty about the strength of the slab anomaly that could impact our final results above the slab in the upper 150 km. The results of a second tomographic inversion (Figures 8 and 9) which does not use a priori data to constrain the location of the slab are used to interpret anomalies below ~200 km.
where the synthetic slab recovery tests (Figure 5) indicate that vertical smearing of the slab is less dominant.

RESULTS

The final results from the teleseismic tomography inversions of the two models are displayed as horizontal depth slices (Figures 6 and 8) and vertical cross sections (Figures 7 and 9). The tomograms displayed in Figures 6 and 7 use the a priori slab information to constrain the location of the subducting slab and are used to interpret anomalies in the upper 200 km while Figures 8 and 9 show tomograms that do not constrain the location of the slab and are used to interpret anomalies below 200 km. Additional depth slices from both models are displayed in the Supplemental Material, Figures S4 and S5. Model resolution is best in the central part of the model, and decreases towards the edges of the model space. As discussed previously, resolution in our model is reasonably good down to roughly 660 km depth. Below that depth, hit quality decreases and anomalies are more poorly resolved. Resolution in the shallowest depth slices is strongly controlled by the location of the stations as the ray paths become near vertical beneath the stations, and vertical smearing has a more pronounced effect and must be taken into account when interpreting anomalies in the shallowest parts of the model space. Horizontal resolution is good throughout the model, allowing us to interpret lateral differences in mantle structure with more confidence. We note that all major labeled anomalies discussed in the following sections (Anomalies A through G) are present in both sets of inversions (Figures 7, 8, 9, and 10), which attests to their robustness.
Imaging the Nazca slab

The most prominent anomaly observed in our tomograms without incorporating any constraints on the location of the slab (Figures 8 and 9) is the trench-parallel fast anomaly that appears to migrate east with depth, which is interpreted to be the subducting Nazca slab. The amplitude of this fast anomaly varies between +1 and +3 percent. The slab anomaly is discontinuous and has lower amplitudes at shallower depths, but becomes more prominent below ~200 km. The fast anomaly that corresponds to the slab is discontinuous between 100 and 165 km between 24° and 28°S (Figure 9). Slow anomalies are resolved in place of the subducting slab even though the synthetic slab resolution tests indicate that at least partial resolution of the slab should occur in this part of the model. Synthetic tests discussed previously (Figure 5) indicate that the slab-shaped anomaly is being shifted vertically along the ray paths, resulting in a diffuse, low amplitude, discontinuous slab anomaly in the shallowest part of the model. Similar difficulties in resolving the subducting slab were observed by Heit et al. (2008) for a teleseismic tomography study along a dense line of stations across the Andes at 21°S and were attributed to limitations in the ray path configuration. The vertical smearing of the Nazca slab anomaly along the ray paths complicates our ability to interpret anomalies in the uppermost parts of our model, hence our additional efforts to constrain the subducting slab in the upper mantle as shown in Figures 6 and 7.

In the deeper parts of the model, the Nazca slab anomaly is well-resolved in our synthetic tests without needing to use a priori information to define the location of the subducting slab. For interpretations in the deeper part of our model, we use the inversion
of the observed data without the imposition of an a priori slab (Figures 8 and 9), allowing us to make observations about the shape of the subducting Nazca slab below 200 km.

The shallow mantle (90-200 km)

The final results from the teleseismic tomography inversion using the a priori slab data to constrain the location of the Nazca slab are shown in Figures 6 and 7. Fast anomalies with amplitudes of between +1 and +2% are observed in the shallowest depth layers in the northern part of the study area (18° to 22°S) under the eastern edge of the Altiplano and Eastern Cordillera (Anomaly A and A’, Figures 6 and 7). The fast anomaly is observed to extend under the Subandean Zone at 20°S but any determination of the extent of this fast anomaly under the Subandean Zone is hindered by the lack of stations in the easternmost part of the study area to the north or south of this latitude. The fast anomalies continue under the Eastern Cordillera between 22° and 26°S, although the amplitude of the anomalies is decreased and the edges are less distinct (Figure 7). An exception is the prominent, circular high velocity feature located at 25°S between the Eastern Cordillera and the Santa Barbara System (Anomaly A’, Figures 6 and 7). This local anomaly remains strong to ~200 km depth where it merges with the slab.

Two separate slow anomalies are observed in the northern part of the study. A slow anomaly with amplitude of -2% is observed at 18°S under the Altiplano (Anomaly B, Figure 6). The lateral extent of this anomaly is unknown, since the limited distribution of stations in this part of the Andes results in a narrow region which is well-resolved. A second, higher amplitude (-2.5 to -3%) slow anomaly is observed at 21°S between 66° and 67°W (Anomaly C, Figure 6). This anomaly is limited to the east by the presence of
the fast anomalies discussed above and has a distinct western boundary at ~67°W where
the amplitude of the resolved anomaly decreases abruptly.

A distinct slow anomaly is observed at shallow depths under the Puna between
~24.5° and 27.5°S (Anomaly D, Figures 6 and 7). While the amplitude of this anomaly
varies between -1.0 and -1.5 %, its lateral extent is well-defined by the fast anomalies
observed to the south and east. The western edge of this slow anomaly is defined by the
location of the subducting slab. To the southeast of the Puna, fast velocity anomalies are
observed with amplitudes of between +1 and +2% (Anomaly E, Figure 6). The
northwestern edges of these fast anomalies are distinct although the eastern and southern
extent of these anomalies is unknown as they fall outside of our study area. A second low
velocity anomaly (Anomaly F, Figure 7) is observed below the slab under the Puna with
amplitudes of -1 to -2 %.

The deeper mantle (~280-660 km)

Results for the teleseismic tomography inversion which do not use the a priori
slab information to define the location of the subducting slab are shown in Figures 8 and
9. Since synthetic slab recovery tests (Figure 5) indicate that the slab anomaly should be
well-resolved in the deeper parts of the model without the extensive vertical smearing
effect noted in the shallow model layers, we can use the inversion of the observed data
without constraining the location of the Nazca slab to make observations about the shape
of the slab in the deeper parts of the model. The slab anomaly is resolved down to 660
km, with along-strike variations in amplitude and width being observed in the tomograms
below 300 km depth (Figure 8). A distinct region of thinning of the slab anomaly is
observed at 24° to 25°S at depths greater than 300 km (Figures 8 and 9). Other regions of thinning are observed in the mantle transition zone (MTZ) at 20°S (Figure 8). The fast slab anomaly appears to thicken in other parts of the MTZ, particularly at 18° and 26°S. Together these anomalies give the slab a laterally periodic thinning and thickening appearance that is especially noticeable below 500 km (Figure 8). We do not have resolution to be certain whether this segmented appearance is the result of vertical tears or stress induced deformation as the slab penetrates the mantle transition zone.

A high amplitude (-3%) slow anomaly is observed below the subducting Nazca slab in the mantle transition zone (410-660 km depth) between 22° and 28°S and is especially prominent south of 24°S (Anomaly G, Figures 8 and 9). Although other velocity anomalies are observed under the subducting Nazca slab, the amplitude of this anomaly makes it distinctive.

**DISCUSSION**

*The shallow mantle (90-200 km)*

Figures 6 and 7 show results for our model where we incorporate the slab in order to better constrain the upper 200 km of the model. In general we observe relatively slow velocities under the Altiplano and Puna at depths of 95 and 130 km and faster velocities to the east under parts of the Eastern Cordillera and Subandean zone (Figure 6). We interpret the lateral change from slow to fast P-wave velocities as the edge of the continental craton underthrusting from the east. The presence of fast anomalies under the Eastern Cordillera (Anomaly A, Figure 6) suggests that the craton extends under parts of the Eastern Cordillera and is consistent with results from previous studies (Beck and
Zandt 2002; Dorbath et al., 1993). If correct this suggests that western extent of
underthrusting of the craton is limited mostly to the Eastern Cordillera and does not
underthrust the entire high plateau as is suggested further north in southern Peru by
Phillips et al. (2012). Another possible interpretation of the high-velocity anomalies
under the Eastern Cordillera is that it represents cold downwellings of foundering
lithospheric material. The cylindrical shaped anomaly at 25°S (Anomaly A’, Figures 6
and 7) is especially suggestive of delaminating lithosphere and has been interpreted as
such by Beck et al. (2014, this volume). We observe a low amplitude slow anomaly at
21°S that partially overlaps and continues to the south of the Los Frailes volcanic field in
the Eastern Cordillera (Anomaly C, Figure 6). Heit et al. (2008) noted a low velocity
anomaly in the same location although their anomaly is more laterally extensive. A
similar low velocity anomaly was observed by Myers et al. (1998) with regional P-wave
tomography slightly north of the location of our anomaly and also correlated with the Los
Frailes volcanic field. A fast P-wave anomaly under the central Altiplano (above ~130
km) in the northern part of the study area (north of 21°S) was observed previously by
Myers et al. (1998) and interpreted as intact mantle lithosphere under this part of the
Altiplano. This anomaly is not obvious in our results probably due to the difference in
resolution between our study and the local tomography study of Myers et al. (1998).

In general, we do not observe a prominent low velocity anomaly under the
Altiplano-Puna Volcanic Complex (APVC) near 23°S. The anomalies observed under the
APVC are low amplitude, slow anomalies, which are not nearly as distinct as those
observed under the Los Frailes and the Puna volcanic fields (Anomalies C and D, Figure
6). This is consistent with results from other studies (as discussed in Sobolev et al., 2006)
which note that the dominant low velocity anomaly in the region under the APVC is in the crust rather than the upper mantle, unlike both the Los Frailes and the Puna volcanic fields.

We observe low velocities at depths of 95 and 130 km beneath the Puna Plateau (Anomaly D, Figures 6 and 7). Slow P-wave velocities in the upper mantle beneath the southern Puna Plateau have been observed in other seismological studies (e.g. Bianchi et al., 2013; Wolbern et al., 2009; Schurr et al., 2006; Koulakov et al., 2006) and interpreted as evidence for recent lithospheric removal. The local P-wave tomography studies show more heterogeneity in the uppermost mantle. Evidence for lithospheric thinning has been seen in shear wave attenuation studies (Whitman et al., 1992) and receiver function studies (Heit et al., 2007) and influx of asthenosphere to the base of the crust following delamination of the continental lithosphere has been proposed to explain both the geophysical observations and the geochemical characteristics of the large ignimbrite deposits in the southern Puna (Kay and Kay, 1993; Kay et al., 1994; Kay and Coira, 2009). The location of the -1 to -2% velocity anomaly observed between 24° and 28°S under the southern Puna in our results is consistent with asthenosphere in the upper mantle under the Puna Plateau. Local tomography studies (e.g. Bianchi et al., 2013; Schurr et al., 2006; Koulakov et al., 2006) have observed small fast anomalies in the upper mantle beneath the Puna which have been interpreted as delaminated blocks of lithosphere. In general, our results are neither as detailed nor do they show as much small scale heterogeneity as the local P-wave travel time tomography models due to the regional scale of our study and therefore we cannot robustly resolve the presence of such small scale features. However, Anomaly A’ (Figure 6), which was not clearly resolved in
previous studies, could represent the larger scale delaminated material that was replaced by the influx of asthenosphere.

In the shallowest depth layer (95 km) fast velocity anomalies are observed to the southeast of the Puna Plateau (Anomaly E, Figure 6). The edges of this anomaly are distinct, and correspond to the surface boundary of the Sierras Pampeanas. This corresponds to the surface location of the Precambrian Pampean Terrane proposed by Ramos et al. (1986) which is thought to comprise the crystalline basement of the Sierras Pampeanas. Therefore, we could be imaging the edge of the thicker mantle lithosphere of the Pampean Terrane. A fast anomaly in this area was also observed by Bianchi et al. (2013) and interpreted similarly. This is also start of the transition from the normal dipping slab to a near horizontal or flat slab. Hence, the fast velocities at a depth of 130 km are likely beginning to sample the downgoing slab as it flattens to the south. The slow anomaly under the slab (Anomaly F, Figure 7) is similar to an anomaly imaged by Bianchi et al. (2013) who interpreted it as a sub-slab region of hot asthenospheric mantle. However, this anomaly has a limited along-strike extent.

The deeper mantle (~280-660 km)

Limited regional-scale imaging of the Nazca slab between 250 and 700 km depth has been done prior to this study. Engdahl et al. (1995) used teleseismic travel time tomography to image the Nazca slab down to depths of 1400 km, concluding that the slab penetrates into the lower mantle after a period of temporary stagnation in the transition zone where the slab thickens before resuming subduction into the lower mantle. Local tomography studies in this region use earthquakes in the slab and usually only image the
upper 150-200 km above the slab and often do not image the slab itself (e.g. Schurr et al., 2006; Koulakov et al., 2006).

The subducting Nazca slab is clearly imaged in our model between 280 and 660 km depth (Figures 8 and 9). A few clusters of deep earthquakes between 500 and 600 km depth are observed in our study area. These deep earthquakes are located within the observed slab anomaly rather than towards the edges, which agrees with the idea that deep earthquakes occur within the cold cores of subducting slabs (Kirby et al., 1996). The amplitude of the observed deep slab in our tomograms varies with both latitude and depth. Some of the variation in amplitude of the deep slab is probably due to ray path distribution as similar amplitude variations are observed in the synthetic slab recovery tests. Variations in slab thickness are also observed in our tomograms. However, similar variations in slab thickness are not reproduced in the output of the synthetic slab recovery tests (Figure 5). This suggests that the observed variability in the thickness of the slab in our resulting model is not being controlled by ray path distribution or the geometry of our model, and thus probably represents real variations in the subducting slab. The subducting Nazca slab shows distinct variations along strike (Figure 8), with localized zones of both thickening and thinning in the mantle transition zone (MTZ).

Since we do not have resolution below the 660 km discontinuity, we cannot rule out the possibility that the slab stagnates in the transition zone east of our image solely based on our study. However, global tomography models show the Nazca slab penetrating into the lower mantle (e.g., Bijwaard et al., 1998; Li et al., 2008; Zhao, 2004; Fukao et al., 2001; 2009; Zhao et al., 2013). Our results show localized thickening of the Nazca slab in the MTZ between 16° and 18°S and south of 25°S (Figures 8 and 9).
Similar thickening of subducting slabs during penetration into the lower mantle has been observed in global models (e.g. Bijwaard et al., 1998; Li et al., 2008) and has been interpreted as evidence for temporary stagnation of a subducting slab in the MTZ before the slab subducts into the lower mantle. Our images of the deeply subducted Nazca slab also show evidence of varying degrees of thinning in several places, possibly resulting from a changing stress state along strike as the slab deforms in the MTZ before resuming subduction into the lower mantle (Figure 10). At these depths the resolution of the model is very uniform and does not show any lateral variations that could be perceived as lateral variations in slab thickness, indicating the robustness of these slab thickness variations. The most extreme areas of thinning occur at 20° and 24°S and appear to correlate with along-strike changes in seismicity between 150 and 300 km depth (Figure 1B). Assuming the National Earthquake Information Center (NEIC) earthquake catalogue is uniform at the magnitude 4 level, there is a distinct decrease in intermediate depth seismicity south of ~24°S (Figure 1B). While it is difficult to directly relate changes in the stress state of the slab in the MTZ with changes in seismicity patterns between 150 and 300 km depth, it is possible that these regions of thinning represent tears in the subducting slab that separate different stress regimes, which could affect the pattern of intermediate depth seismicity.

The high amplitude low velocity anomaly below the slab between 22° and 28°S in the MTZ is well-resolved in our model (Anomaly G, Figures 8 and 9). This anomaly can be explained either by a thermal anomaly or by volatiles in the MTZ. An electrical conductivity study further to the south near 33°S also indicates the presence of a potentially water-rich or hot anomaly in the sub-slab mantle from transition zone depths.
(Burd et al. 2013). A thermal anomaly would locally increase temperature and therefore lower the seismic velocity. Similarly, the presence of increased water in the MTZ could also lower the seismic velocity. From the seismic velocities alone we cannot distinguish between these two causes. A similar low velocity anomaly has been observed in the sub-slab mantle in other subduction zones (e.g., Obayashi et al., 2006; Zhao, 2004; Zhao et al., 2013). A low velocity zone directly above the 410-km discontinuity under the subducting Pacific slab in the Honshu subduction zone has been interpreted as a local thermal anomaly (e.g., Obayashi et al., 2006; Bagley et al., 2009; Zhao, 2004; Zhao et al., 2013). Bagley et al. (2009) and Obayashi et al. (2006) calculated that the observed low velocity anomalies can be explained by approximate temperature anomalies of 150°C or 200°C respectively. The exact origin of these thermal anomalies is uncertain, although most studies suggest that they are associated with upwellings of hot lower mantle material either as a result of local-scale convection induced by the subduction of the nearby slab or as a result of the presence of small scale plumes (Zhao, 2004; Bagley et al., 2009; Morishige et al., 2010). Wolbern et al. (2009) imaged a depressed 410 km discontinuity and an elevated 660 km discontinuity just to the west of our low velocity anomaly at 67.5° to 68°W. They argue that this thinning of the mantle transition zone is consistent with a local thermal anomaly in the sub-slab mantle but were unable to constrain the extent of the anomaly due to poor ray coverage at that depth.

It is also possible that water is the cause of the sub-slab anomaly observed in our tomograms. Schmerr and Garnero (2007) observed a depressed 410 km discontinuity east of the Nazca slab using SS phases and argued that it cannot be explained solely by a thermal origin. Several studies suggest that subduction zones can carry water into the
MTZ (e.g., Bercovici and Karato, 2003; Smyth and Jacobsen, 2006). The hydrated material is chemically buoyant and should remain on top of the MTZ, potentially melting. A similar process was theorized by Bercovici and Karato (2003) as part of their transition zone water filter model of mantle convection which argues for the hydration of the MTZ. As the trench migrates west and the Nazca slab pushes the sub-slab mantle west it will disrupt the 410 km discontinuity and could drag hydrated mantle material that has accumulated near the 410 km discontinuity into the MTZ. A region of hydrated material in the MTZ formed in this way could explain our low velocity zone. The depressed 410 discontinuity observed by Schmerr and Garnero (2007) is much broader in the east-west direction than the width of the slab suggesting that there is hydrated material both above and below the slab. In addition, the 410 discontinuity is significantly disrupted beneath the slab (Schmerr and Garnero, 2007). Hence, it is possible that our low velocity anomaly is consistent with some component of hydrated material in the MTZ beneath the slab.

CONCLUSION

Until recently, limited regional scale seismic imaging has been done in the central Andes. Many studies of the central Andes have either looked at a small region under a single seismic network, limiting their maximum depth of resolution, or imaged the region as part of a global study, limiting their ability to resolve smaller scale variations. Our regional scale P-wave tomographic images of the central Andes show significant along-strike variation in upper mantle structure. Fast anomalies under the Eastern Cordillera between 16° and 26°S provide evidence for underthrusting of the Brazilian cratonic
lithosphere under the Subandean Zone and Eastern Cordillera along much of the central Andes or possibly a “curtain” of delaminating lithospheric material. Slow velocity anomalies underneath the Puna Plateau are related to the presence of partial melt in the shallow mantle, while fast velocity anomalies to the southeast correspond to the mantle lithosphere of the Precambrian Pampean terrane.

Detailed imaging of the structure of the Nazca slab in the deeper parts of the upper mantle has not been done in this part of the central Andes prior to this study. We have imaged previously unseen structure in the subducting Nazca slab at the base of the upper mantle and in the transition zone. Variations in the width of the slab anomaly in the deeper parts of the model show evidence for deformation of the slab between 300 and 660 km (Figure 10). Thickening of the slab in the mantle transition zone is observed in several places, in agreement with the idea that the Nazca slab stagnates at least temporarily in the transition zone before resuming subduction into the lower mantle. A sub-slab low velocity anomaly in the mantle transition zone between 22° and 28°S is similar to those seen in other subduction zones, and is interpreted as either a local thermal anomaly or a region of hydrated material in the mantle transition zone.

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REFERENCES


Norabuena, E.O., Dixon, T.H., Stein, S., and Harrison, C., 1999, Decelerating Nazca-

Obayashi, M., Sugioka, H., Yoshimitsu, J., and Fukao, Y., 2006, High temperature
anomalies oceanward of subducting slabs at the 410-km discontinuity: Earth and

Oncken, O., Hindle, D., Kley, J., Elger, K., Victor, P., and Schemmann, K., 2006,
Deformation of the central Andean upper plate system – Facts, fiction, and
constraints for plateau models, in Oncken, O., et al., eds., The Andes—Active
3–27.

1, p. 43-71.

Pavlis, G., and Vernon, F., 2010, Array processing of teleseismic body waves with the
USArray: Computers & Geosciences, v. 36, p. 910-920.

subducting slab structure in the region of the 2010 M8.8 Maule earthquake (30-
40°S), Chile: Geophysical Journal International, v. 191, p. 317-324, doi:
10.1111/j.1365-246X.2012.05624.x.

Phillips, K., Clayton, R.W., Davis, P., Tavera, H., Guy, R., Skinner, S., Stubailo, I.,
Audin, L., and Aguilar, V., 2012, Structure of the subduction system in southern

Quinteros, J., and Sobolev, S.V., 2013, Why has the Nazca plate slowed since the Neogene?: Geology, v. 41, no. 1, p. 31-34.


Tassara, A., 2005, Interaction between the Nazca and South American plates and formation of the Altiplano-Puna Plateau: Review of a flexural analysis along the


the Altiplano and Puna Plateaus in the Central Andes: Geophysical Journal

Yuan, X., Sobolev, S.V., Kind, R., Oncken, O., Bock, G., Asch, G., Schurr, B., Graeber, F., Rudloff, A., Hanka, W., Wylegalla, K., Tibi, R., Haberland, C., Rietbrock, A.,
Giese, P., Wigger, P., Rowr, P., Zandt, G., Beck, S., Wallace, T., Pardo, M., and
Comte, D., 2000, Subduction and collision processes in the Central Andes

Yuan, X., Sobolev, S.V., and Kind, R., 2002, Moho topography in the central Andes and
402.

and Silver, P.G., 1996, Anomalous crust of the Bolivian Altiplano, central Andes:
Constraints from broadband regional seismic waveforms: Geophysical Research

Zandt, G., Leidig, M., Chmielowski, J., Baumont, D., and Yuan, X., 2003, Seismic
detection and characterization of the Altiplano-Puna Magma Body, central Andes:

Zhao, D., 2004, Global tomography images of mantle plumes and subducting slabs:

Zhao, D., Yamamoto, Y., and Yanada, T., 2013, Global mantle heterogeneity and its
influence on teleseismic regional tomography: Gondwana Research, v. 23, p. 595-
616.
### Tables

Table 1. Seismic networks incorporated into this study

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<th>Network Name</th>
<th>Sensor type</th>
<th>No. of stations</th>
<th>Reference</th>
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<tr>
<td>Global Seismograph Network</td>
<td>Broadband</td>
<td>1 (LVC)</td>
<td>Butler et al., 2004</td>
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<tr>
<td>Global Telemetered Seismograph Network</td>
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<td>Broadband</td>
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<td>GFZ, 2013</td>
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<td>Chmielewski et al., 1999</td>
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<td>Intergrated Plate Boundary Observatory Chile</td>
<td>Broadband</td>
<td>14</td>
<td>Sodoudi et al., 2011</td>
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FIGURES

Figure 1. A) Map showing seismic station locations for individual networks used in the study and topography of the central Andes. Yellow outlines mark locations of Los Frailes Volcanic Field (LFVF), Altiplano-Puna Volcanic Complex (APVC), and Puna Volcanic Field (PVF) (modified from Kay et al., 2010). Boundaries of geomorphic provinces (dashed lines) are modified from Tassara (2005). WC: Western Cordillera, AP: Altiplano, EC: Eastern Cordillera, SA: Subandean Zone, PN: Puna, SB: Santa Barbara System, SP: Sierras Pampeanas. Red triangles mark location of Holocene volcanoes (Siebert and Simkin, 2002). B) Map of seismicity (magnitude > 4) and Nazca slab depth contours. Slab contours are from the Slab1.0 global subduction zone model (Hayes et al. 2012). Earthquake data are from 1973 to 2012; U.S. Geological Survey – National Earthquake Information Center (NEIC) catalog.
Figure 2. A) Global map centered on our study region (black square) showing the location of events used in this study. Darker gray circles mark events used for direct P arrivals while open circles mark events used for PKIKP arrivals. B) Plot showing the azimuthal distribution for all rays used in the study. Ray distribution is strongly controlled by the location of plate boundaries where the earthquakes are generated.

Figure 3. Plot showing the tradeoff analysis between the variance reduction and the Euclidean model norm (L2) performed to choose preferred overall damping (D1-D10) and smoothing (S1 –S10) weights. The black star shows the damping (D6) and smoothing (S5) parameters used in this study.
Figure 4. Horizontal depth slices for the checkerboard tests for every other model layer. Input for neutral layers (0% velocity deviation) is not shown. Output for neutral layers (60, 200, 365, 555 km depth) is shown in the left column. Resolution of velocity anomalies in neutral layers shown here indicates that some vertical smearing is occurring. On the right is the input and output for the layers with the checkerboard anomalies. The checkerboard tests show that for shallower layers, resolution is controlled by station distribution as expected. Deeper model layers indicate that while the input amplitude cannot be completely resolved, lateral changes in anomalous velocity resolve with little horizontal smearing. Resolution is lost towards the edges of the model region. Checkerboard tests for additional model layers and cross sections through the model are in Supplemental Material, Figures S2 and S3.
Figure 5. Cross-section results for our synthetic slab recovery tests. The geometry of our input slab model is based on the Slab1.0 contours (green line, Hayes et al., 2012). A) E-W oriented cross sections through the synthetic input (left) and output (right) model. Decreased amplitude recovery and vertical smearing of the recovered slab anomaly are observed in the upper 200 km of the model. In general, the amplitude recovery increases
with depth. B) N-S oriented cross sections through the synthetic input (left) and output (right) model. The recovered slab anomaly at shallower depths (cross section at 66°W) is more diffuse than at deeper depths (cross section as 64°W).

Figure 6. Horizontal depth slices for 95, 130, and 165 km from the tomography model using a priori information to constrain the slab as discussed in the text. The short dashed line represents the edge of the well-resolved region of the model, defined as regions with hit quality greater than 0.2 (Biryol et. al., 2011). Geomorphic provinces (green lines) are the same as in Figure 1A. Stars mark locations of seismic stations used in the study. Red triangles mark location of Holocene volcanoes (Siebert and Simkin, 2002). Black dots are earthquake locations from the EHB catalog (Engdahl et. al., 1998). Solid black line marks edge of fast velocity anomaly (Anomaly A and A’) interpreted as possible edge of cratonic lithosphere or delaminated lithosphere. Labeled anomalies (A, A’, B, C, D, and E) are discussed in the text. Additional depth slices (280-660 km depth) from this model are in the Supplemental Material, Figure S4.
Figure 7. East-west oriented cross sections through the tomography model shown in Figure 6. Location of the Nazca slab is constrained using a priori information as in Figure 6. Dashed lines are the same as in Figure 6. Black dots are earthquake locations from the EHB catalog (Engdahl et. al., 1998). Solid black line marks the top of the Nazca slab from Slab1.0 model (Hayes et. al., 2012). Labeled anomalies (A, A’, C, D, F, and G) are discussed in the text.
Figure 8. Horizontal depth slices from 280 to 660 km from the tomography model without a priori constraints on the location of the subducting slab. The short dashed line represents the edge of the well-resolved region of the model, defined as regions with hit quality greater than 0.2 (Biryol et. al., 2011). Stars mark locations of seismic stations used in the study. Red triangles mark location of Holocene volcanoes (Siebert and Simkin, 2002). Black dots are earthquake locations from the EHB catalog (Engdahl et. al., 2007).
al., 1998). Solid black lines are slab contours from Slab1.0 model (Hayes et al., 2012).
Labeled anomaly G is discussed in the text. Additional depth slices (95-165 km depth) from this model are in the Supplemental Material, Figure S5.

Figure 9. East-west oriented cross sections through the tomography model without a priori constraints on the location of the subducting slab as in Figure 8. Dashed lines are the same as in Figure 8. Black dots are earthquake locations from the EHB catalog (Engdahl et al., 1998). Solid black line marks the top of the Nazca slab from the Slab1.0
model (Hayes et. al., 2012). Labeled anomalies (A, A’ C, D, F, and G) are discussed in the text.

Figure 10. 3-D diagram of the resolved subducting Nazca slab and prominent mantle low velocity anomalies inferred from our tomographic models. The isosurfaces for this model are obtained by tracing the most coherent low velocity anomalies (less than negative 3%) and slab related (greater than positive 3%), coherent fast anomalies in the tomographic model. Slab geometry above 200 km is determined entirely from the Slab1.0 model (Hayes et. al., 2012). Anomalies B, C, D, F, and G labeled as in previous figures. Boundaries of geomorphic provinces as in Figure 1.
SUPPLEMENTAL MATERIAL

Figure S1. Normalized hit quality plots. Black stars on 95 km depth layer are station locations. Hit quality for a node is based on the number and azimuthal distribution of rays that sample that node. Good hit quality is indicated by red shading while poor hit quality is indicated by blue shading. Hit quality is strongly controlled by station distribution in
uppermost depth slices with decreasing dependence on station distribution as depth increases.

Figure S2. Horizontal slices for checkerboard tests for every other model layer. Input for neutral layers (0% velocity deviation) is not shown. Output for neutral layers (95, 240,
410, 605 km depth) is shown in the left column. Resolution of velocity anomalies in neutral layers shown here indicates that vertical smearing is occurring. Checkerboard tests show that for shallower layers, resolution is controlled by station distribution as expected. Deeper model layers indicate that while the input amplitude cannot be completely resolved, lateral changes in anomalous velocity resolve with little horizontal smearing. Resolution is lost towards the edges of the model region.
Figure S3. Cross sections through checkerboard tests. Irregular distribution of anomalies with depth in synthetic input (left) is due to the dilation of the node spacing with depth. A) E-W oriented cross sections through the synthetic input (left) and output (right) model. B) N-S oriented cross sections through the synthetic input (left) and output (right) model.
Figure S4. Horizontal depth slices from 280 to 660 km from the tomography model. Location of the Nazca slab is constrained using a priori information as discussed in the text. Dashed lines as in Figure 7. Stars mark locations of stations. Red triangles mark location of volcanoes. Black dots are earthquake locations from the EHB catalog (Engdahl et. al., 1998). Labeled anomaly G is discussed in the text.
Figure S5. Horizontal depth slices for 95, 130, and 165 km from the tomography model without a priori constraints on the location of the subducting slab. Geomorphic provinces (green lines) same as in Figure 1a. Dashed lines as in Figure 7. Solid black lines are slab contours from Slab1.0 model (Hayes et. al., 2012). Labeled anomalies (A, A’, B, C, D, and E) are discussed in the text.
APPENDIX B: IMAGING THE TRANSITION FROM FLAT TO NORMAL SUBDUCTION: VARIATIONS IN THE STRUCTURE OF THE NAZCA SLAB AND UPPER MANTLE UNDER SOUTHERN PERU AND NORTHEASTERN BOLIVIA

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SUMMARY

Two arrays of broadband seismic stations were deployed in the north central Andes between 8° and 21°S, the CAUGHT array over the normally subducting slab in northwestern Bolivia and southern Peru, and the PULSE array over the southern part of the Peruvian flat slab where the Nazca Ridge is subducting under South America. We apply finite frequency teleseismic P- and S-wave tomography to data from these arrays to investigate the subducting Nazca plate and the surrounding mantle in this region where the subduction angle changes from flat north of 14°S to normally dipping in the south. We present new constraints on the location and geometry of the Nazca slab under southern Peru and northwestern Bolivia from 95 to 660 km depth. Our tomographic images show that the Peruvian flat slab extends further inland than previously proposed along the projection of the Nazca Ridge. Once the slab re-steepens inboard of the flat slab region, the Nazca slab dips very steeply (~70°) from about 150 km depth to 410 km depth; below this the slab thickens and deforms in the mantle transition zone. We tentatively propose a ridge-parallel slab tear along the north edge of the Nazca Ridge between 130 and 350 km depth, although additional work is needed to confirm the existence of this feature. The sub-slab mantle directly below the inboard projection of the Nazca Ridge is characterized by a prominent low velocity anomaly. We interpret this as either the result of limited partial melting due to sub-slab flow or as asthenospheric mantle at shallow depths due to the existence of thinned oceanic lithosphere under the Nazca Ridge. South of the Peruvian flat slab fast anomalies are imaged in an area confined to the Eastern Cordillera and bounded to the east by well resolved low velocity anomalies. These low velocity anomalies at depths greater than 100 km suggest that thick
mantle lithosphere associated with underthrusting of cratonic crust from the east is not present. In northwestern Bolivia a vertically elongated fast anomaly under the Subandean Zone is interpreted as a block of delaminating lithosphere.

INTRODUCTION AND BACKGROUND

The Nazca-South America subduction zone in northern Bolivia and southern Peru is characterized by a change in the upper plate structure, from a narrower orogen in the north to the broad Central Andean Plateau (CAP), typically defined as the region above 3 km elevation, in the south. The CAP is one of the largest high plateaus in the world, second only to the Tibetan plateau. The northern part of the CAP, the Altiplano plateau, is an internally drained basin characterized by relatively flat, although high, topography with an average elevation of ~3.5 km (Isacks, 1988; Whitman et al., 1996). The edges of the Altiplano are defined by the modern active volcanic arc in the west, the Western Cordillera, and a fold and thrust belt in the east, the Eastern Cordillera (Fig. 1a). Deformation occurred in the Eastern Cordillera between ~40 and 15 Ma (McQuarrie et al., 2005; Oncken et al., 2006) before shifting to the modern, thin-skinned fold and thrust belt, the Subandean Zone, where deformation began after 10 Ma (Allmendinger et al., 1997; Oncken et al., 2006).

The development of the Altiplano plateau is debated, with different studies arguing for different mechanisms of plateau uplift. The timing of plateau uplift is also debated, with some studies arguing for gradual uplift of the plateau due to shortening and thickening of the crust between 40-30 and 10 Ma based on structural reconstructions (e.g., McQuarrie et al., 2005; Elger et al., 2005; Oncken et al., 2006; Lamb, 2011) while
paleoelevation studies (e.g. Ghosh et al., 2006; Garzione et al., 2008) argue for a rapid period of uplift since 10 Ma resulting from large scale delamination of the lithosphere. Work by Saylor & Horton (2014) using volcanic glasses indicates that uplift of the plateau may have been spatially and temporally non-uniform. This leads them to conclude that the uplift of the plateau could be linked to locally varying crustal shortening or localized removal of mantle lithosphere rather than uniform shortening along the entire CAP or large scale delamination of the lithosphere. Local delamination of the lithosphere has been proposed in several regions of the Andes using both geochemical (e.g., Kay & Kay, 1993; Kay et al., 1994; Kay & Coira, 2009) and seismic evidence. Seismic observations include both the presence of fast velocity anomalies in the mantle, hypothesized to be delaminated blocks of lithosphere (Schurr et al., 2006; Asch et al., 2006; Koulakov et al., 2006; Bianchi et al., 2013; Beck et al., 2015), and anomalously low velocities at the base of the crust, which are interpreted as asthenospheric mantle at shallow depths (Myers et al., 1998; Heit et al., 2008). Receiver function studies from the central Andes south of our study region (20° to 24°S) indicate that the crustal thickness varies across the Andes (Beck & Zandt, 2002; Yuan et al., 2000, 2002; Wölbern et al., 2009) going from a minimum in the forearc and foreland (~35 km thick) to maxima of up to 70 km thick in the Western and Eastern Cordilleras. The average crustal thickness observed in the Altiplano is ~65 km. Similar variations in crustal thicknesses are observed in receiver function studies between 8° and 21°S using data from the CAUGHT and PULSE deployments (Ryan et al., 2012; Bishop et al., 2013). Fast seismic velocities have been observed in the forearc crust in both local earthquake tomography studies between 21° and 24°S (Graeber and Asch, 1999;
Koulakov et al., 2006; Schurr et al., 2006), and in ambient noise tomography along much of the central Andes (Ward et al., 2013, 2014), possibly resulting from the presence of plutonic remnants of earlier volcanic arcs. Generally the thick crust of the Altiplano plateau has been imaged with low average crustal P and S-wave velocities (Swenson et al., 2000; Ward et al., 2013, 2014). Anomalously high velocities in both the crust and uppermost mantle (<150 km depth) have been observed under the Subandean zone and have been attributed to underthrusting of Brazilian cratonic lithosphere (Myers et al., 1998; Beck & Zandt, 2002; Phillips et al., 2012; Scire et al., 2015). Similar observations of high velocity anomalies near 16°S (Dorbath et al., 1993; Dorbath & Granet, 1996) and analyses of focal mechanisms further north at ~11°S (Dorbath et al., 1991) are also interpreted as evidence for cratonic underthrusting along much of the central Andes.

Continuous removal of mantle lithosphere under the Altiplano/Eastern Cordillera boundary is suggested to accommodate the underthrusting of the Brazilian craton between 18° and 20°S (McQuarrie et al., 2005; Myers et al., 1998; Beck & Zandt, 2002).

The subducted Nazca plate in northern Bolivia and southern Peru undergoes a change in slab dip from the Peruvian flat slab region in the north to the normal steep slab region south of 14°S (Fig. 1b). Subduction of the Nazca plate occurs at an angle of ~30° to a depth of about 300 km under the central Andes as defined by slab seismicity (Cahill & Isacks, 1992; Hayes et al., 2012) and tomography (Scire et al., 2015). North of ~14°S, the Nazca slab subducts at a near horizontal angle for several hundred kilometers inland before resuming a more normal angle of subduction (Cahill & Isacks, 1992; Hayes et al., 2012). The Nazca plate is ~45 Ma old when it enters the subduction zone beneath the central Andes (Müller et al., 2008), and is converging with the South American plate at
8.5 cm/yr relative to a stable South American reference frame (Somoza & Ghidella, 2012). The Nazca slab has been imaged in a nested regional-global tomography study south of our study area by Pesicek et al. (2012) who imaged a discontinuous slab at ~38°S, providing evidence for vertical tearing of the slab along the zone of weakness provided by the subduction of the Mocha Fracture Zone. Bianchi et al. (2013) used regional and global tomography to constrain the subducting Nazca slab to depths of ~300 km under the Puna Plateau between 24° and 29°S. They observed some variations in the amplitude of the slab anomaly as well as a disruption of the slab anomaly at ~26°S between 100 and 200 km depth by a low velocity anomaly which they note may be the result of vertical smearing of low velocity anomalies in the crust down into the mantle, resulting in the discontinuous slab anomaly. In a P-wave tomography study along a line at 21°S Heit et al. (2008) noted difficulties resolving the Nazca slab at depths shallower than ~300 km without it being shifted vertically relative to the predicted location of the slab based on the observed slab seismicity from the EHB catalog (Engdahl et al., 1998) and used the earthquake location data to constrain the location of the slab in their final model. Scire et al. (2015) also had difficulties resolving the Nazca slab above ~250 km depth and used earthquake location data to constrain the location of the slab in order to interpret anomalies in the upper 250 km of the mantle. Both Scire et al. (2015) and Engdahl et al. (1995) observed increases in the thickness of the Nazca slab in the mantle transition zone (MTZ), consistent with at least temporary stagnation of the slab in the MTZ. Similar observations of changes in the thickness of the Nazca slab in the MTZ are observed in global tomography studies (e.g., Bijwaard et al., 1998; Zhao 2004; Li et al., 2008; Fukao et al., 2001, 2009; Zhao et al., 2013). Both Engdahl et al. (1995), who used
teleseismic and regional P phases to image the Andean subduction zone between 5° and 25°S down to ~1400 km depth, and global tomography studies (e.g., Bijwaard et al., 1998; Zhao 2004; Li et al., 2008; Fukao et al., 2001, 2009; Zhao et al., 2013) observe that the Nazca slab continues subduction into the lower mantle although the high velocity anomaly associated with the slab becomes more amorphous in the lower mantle.

The flat slab segment in Peru was first identified by Barazangi & Isacks (1976, 1979) although the exact geometry was a matter of some debate (e.g., Snoke et al., 1977; Hasegawa & Sacks, 1981; Boyd et al., 1984). Flat slab subduction is thought to be influenced by the subduction of overthickened oceanic lithosphere although other factors, including the relative motion of the upper plate, also appear to contribute (Gutscher et al., 2000; van Hunen et al., 2002; Espurt et al., 2008; Martinod et al., 2013). However, debate exists about the importance of the subduction of overthickened oceanic lithosphere relative to other factors in the development of flat slab subduction (e.g., Gerya et al., 2009; van Hunen et al., 2004; Skinner & Clayton, 2013; Phillips & Clayton, 2014). Regions of flat slab subduction are characterized by the absence of volcanism. This appears to be due to the displacement of the asthenospheric mantle wedge by the slab, resulting in a “cold” mantle wedge in which melting cannot occur at the typical distance from the trench.

The Nazca slab is estimated to flatten to the north of ~14°S at about 100 km depth for several hundred kilometers inland (Cahill & Isacks, 1992; Hayes et al., 2012; Phillips & Clayton, 2014) although some uncertainty in the exact geometry of the Peruvian flat slab exists due to the limited slab seismicity (Fig. 1b). The Western Cordillera is an active volcanic arc up to around 14°S where volcanism is shut off by the Peruvian flat
slab segment (McGeary et al., 1985). The southern end of the flat slab segment near 14°S is bounded by the Nazca Ridge, a bathymetric high on the Nazca plate thought to have been formed by the Easter-Salas hotspot (see Ray et al., 2012 and references therein for a complete discussion of the origin of the Nazca Ridge). The oceanic crustal thickness of the Nazca Ridge is estimated to be between 15 and 18 km from Rayleigh wave dispersion curve modeling (Woods & Okal, 1994) and offshore seismic reflections studies (Hampel et al., 2004). The Peruvian flat slab extends ~1500 km to the northwest, steepening near the Grijalva Fracture Zone at ~5°S. The subduction of a large oceanic plateau, termed the Inca Plateau, which corresponds to the Marquesas Plateau in the western Pacific is proposed as an explanation for the lateral extent of the Peruvian flat slab (Gutscher et al., 1999b). Subduction of this hypothetical plateau may support flat subduction in the northern section of the Peruvian flat slab, although Skinner & Clayton (2013) argue that the Inca Plateau is too far east to support the flat slab.

Increased plate coupling is also considered to be one of the consequences of shallow subduction, with increased upper plate seismicity observed above the Pampean flat slab segment (Gutscher, 2002). The subduction of the Nazca Ridge has an observable effect on the structure of the upper crust, with the Fitzcarrald Arch in the Amazonian foreland in Peru as an example (Espurt et al., 2007; Regard et al., 2009; Espurt et al., 2010). Evidence for uplift associated with the subducting Nazca Ridge is also observed along the Peruvian coastline (Saillard et al., 2011; Macharé & Ortlieb, 1992; Hsu, 1992). Reconstructions of the migration history of the Nazca Ridge initially indicated that the Nazca Ridge entered the trench at 8 Ma at 8°S (von Huene et al., 1996). A more recent reconstruction by Hampel (2002) using comparisons with the Tuamotu Plateau, the
conjugate feature of the Nazca Ridge in the western Pacific, concludes that the Nazca Ridge entered the trench at ~11°S at about 11 Ma. The oblique angle of the Nazca Ridge to the trench and the plate convergence direction results in migration of the ridge south along the margin although estimates of the rate of southward migration vary between studies. Arc volcanism appears to have shut off at ~4 Ma above the present day location of the Nazca Ridge (Rosenbaum et al., 2005), indicating that the subducting slab flattened in this region around the end of the Miocene.

Although some regions of flat slab subduction have been extensively studied, particularly the Pampean flat slab segment in central Chile (e.g., Porter et al., 2012; Gans et al., 2011; Marot et al., 2014; Burd et al., 2013) the effect of along strike changes in slab dip on the deformation pattern of the subducting plate are not completely understood. As the angle of subduction changes along strike, the subducting plate is forced to bend and deform to accommodate the change in subduction angle. This change in subducting plate geometry has the potential to result in thinning or even tearing of the subducting slab between regions of steep and flat subduction. Mantle flow patterns would be affected by the geometry of the subducting slab, particularly if tearing of the slab occurs. Eakin & Long (2013) observe trench-normal fast shear wave splitting directions under the Peruvian flat in the sub-slab mantle, which shift to trench-oblique fast splitting directions to the south under the normally dipping slab. Changes in splitting directions are also observed in the mantle above the flat slab on to the north and south of the Nazca Ridge (Eakin et al., 2014). This implies that the flat slab is influencing mantle deformation and flow both above and below the Nazca slab. Slab tears are postulated further north in the Andes beyond the northern edge of the Peruvian flat slab where the
Carnegie Ridge is subducting near 1°S (Gutscher et al., 1999a). Evidence for slab tearing related to Carnegie Ridge subduction includes variations in volcano geochemistry and upper mantle focal mechanisms. However, studies above the Pampean flat slab in central Chile indicate that while mantle flow patterns are being affected by the shallowing of the slab, the slab bends in the mantle as the slab geometry shifts from normal to flat subduction rather than tearing (Anderson et al., 2004). A slab tear has been hypothesized on the southern edge of the Peruvian flat slab as a result of the abrupt change in slab dip (e.g., Barazangi and Isacks, 1979), but later studies based on slab seismicity (e.g., Hasegawa and Sacks, 1981; Schneider and Sacks, 1987; Kumar, et al. 2014) and receiver functions (Phillips & Clayton, 2014) indicate that the change in slab dip south of the Nazca Ridge is accommodated by bending of the slab rather than a tear above 200 km depth. Recent work using receiver functions (Bishop et al., 2013) and earthquake surface wave tomography (Knezevic Antonijevic et al., 2014) have proposed the possibility of a shallow, trench parallel tear above 100 km depth to the north side of the ridge.

We deployed two arrays of seismic stations, the CAUGHT array in northwestern Bolivia and southern Peru, and the PULSE array which was deployed over the southern part of the Peruvian flat slab, to investigate the structure of the mantle and the subducting Nazca plate in this region (Fig. 1a). We use finite frequency teleseismic P- and S-wave tomography to image the mantle from depths of 95 km to the base of the mantle transition zone between 8° and 21°S to explore variations in upper mantle structure and to try to better constrain the geometry of the subducting Nazca slab. This study offers further insights into how the Altiplano and the Andean system interacts with the mantle below by presenting new tomography images of the Andean mantle above the normally dipping
slab under the central Andes. In addition, constraining the Nazca slab under northwestern Bolivia and southern Peru in the transition between flat and normal subduction allows us to investigate the effects of flat subduction on the deformation and geometry of the subducting slab.

**DATA**

The dataset assembled for this study was collected from 99 broadband seismometers deployed in northwestern Bolivia and southeastern Peru between 2010 and 2013 (Fig. 1a). The CAUGHT deployment includes 48 stations in Bolivia and Peru. The PULSE deployment consists of an additional 38 stations deployed in Peru, primarily over the southern part of the Peruvian flat slab. An additional 8 stations from the PeruSE deployment (Phillips et al., 2012), including 4 stations along the coast of Peru, and 3 stations from the permanent Plate Boundary Observatory network (Sodoubi et al., 2011) in Chile were included to improve coverage of the study area. Also included are 2 permanent stations, LPAZ and NNA, from the Global Telemetered Seismic Network and the Global Seismograph Network respectively.

Arrival times were picked for direct P phases for 150 earthquakes with magnitudes greater than 5.0 between 30° and 90° away from the study region. Additional arrivals were picked for PKIKP phases for 85 earthquakes with a similar magnitude limit between 155° and 180° away from the study region (Fig. 2a). S-wave data were rotated to the radial and tangential components and direct S-wave arrivals were picked on the tangential component for 144 earthquakes with magnitudes greater than 5.0 between 30° and 90° away from the study region (Fig. 2c). Similar to our previous study (Scire et al.,
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2015), first arrivals were picked using the multi-channel cross correlation technique described by VanDecar & Crosson (1990) and modified by Pavlis & Vernon (2010). 14,124 direct P and 6,889 PKIKP arrivals were picked in three frequency bands with corner frequencies 0.2 to 0.8 Hz, 0.1 to 0.4 Hz, and 0.04 to 0.16 Hz. Distribution of the picks in the three frequency bands is dominated by the 0.2 to 0.8 Hz band, with 48.93% of all picks in the 0.2 to 0.8 Hz band, 27.10% in the 0.1 to 0.4 Hz band, and 23.97% in the 0.04 to 0.16 Hz band. 6,014 direct S arrivals were also picked in three frequency bands with corner frequencies 0.1 to 0.7 Hz, 0.04 to 0.16 Hz, and 0.01 to 0.09 Hz. The S arrivals are fairly evenly split between the 0.01 to 0.09 Hz band and the 0.04 to 0.16 Hz band with 48.84% of all picks occurring in the 0.01 to 0.09 Hz band and 40.90% in the 0.04 to 0.16 Hz band. Relatively few high quality S arrivals were observed in the highest frequency band with only 10.26% of all picks occurring in the 0.1 to 0.7 Hz band.

Back azimuthal distribution of incoming rays for both the P- and S-waves (Figs 2b and 2d) is uneven, with the majority of rays coming from direct P and direct S phases from earthquakes either in the Middle America subduction zone to the northwest or the South Sandwich Islands subduction zone to the southeast. This bimodal distribution is further emphasized by the limited number of usable events to the east of our study region on the Mid-Atlantic Ridge. Travel time residuals were calculated relative to IASP91 (Kennett & Engdahl, 1991). The residuals were then demeaned for each event to determine the relative travel time residuals.

Travel times were corrected for crustal variations using Moho depth estimates from receiver functions from Bishop et al. (2013) and Ryan et al. (2012). Limited information about crustal velocity variations in our study region exists, so crustal
corrections were made using a homogenous layered velocity model corresponding to the average velocity of the central Andean crust (Zandt et al., 1996; Swenson et al., 2000; Dorbath & Masson, 2000). Crustal corrections for stations in the forearc were calculated using a faster velocity model based on local tomography studies from further south (Graeber & Asch, 1999; Koulakov et al., 2006; Schurr et al., 2006) and ambient noise tomography which covers part of our study region (Ward et al., 2013, 2014).

METHOD

This study uses a finite-frequency teleseismic tomography algorithm discussed in detail by Schmandt & Humphreys (2010) and used in our previous study (Scire et al., 2015). The finite-frequency approximation defines frequency dependent volumes around the theoretical ray path that are sampled by each ray (Hung et al., 2000; Dahlen et al., 2000). Unlike the ray theory approximation discussed by Aki et al. (1977) which assumes that sampling occurs only along the infinitesimally thin theoretical ray path, our approach defines the sampled area around the geometrical ray path by the first Fresnel zone, whose width is dependent on both the frequency and the distance along the ray path. Within the Fresnel zone, sensitivity varies with zero sensitivity along the theoretical ray path. The sampling in each model layer is determined by the width of the Fresnel zone, which increases with distance from the source or receiver, and the differential sensitivity within the Fresnel zone, resulting in theoretical “banana-doughnut” sensitivity kernels (Dahlen et al., 2000). A detailed discussion of the specific algorithm used to calculate the sensitivity kernels can be found in Schmandt & Humphreys (2010). Multiple nodes in each model layer are sampled by these sensitivity kernels for a single arrival, increasing
sampling of the model space over tomography algorithms that assume that sampling occurs only along the theoretical ray path.

The model space is parameterized into a series of nodes in a non-uniform grid that dilates with increasing depth and distance from the center of the model as in Scire et al. (2015). In the shallowest layers of the model, the horizontal nodes are spaced 35 km apart in the center of the model where sampling is densest, increasing to 56 km apart at the edges of the model. Vertical node spacing increases with depth, from 35 km at 60-95 km depth to 55 km at depths greater than 555 km. The horizontal node spacing also increases with depth, with nodes in the center of the model increasing from 35 km spacing in the shallowest model layers to 49 km in the deepest model layers. Node spacing on the outer edges of the model increases from 56 km in the shallowest model layers to 78 km in the deepest model layers. The sampling of the model space was determined by calculating normalized hit quality maps in which each node is assigned a hit quality value between 0 and 1 that is a function of both the number of rays sampling that node as well as the azimuthal distribution of those rays (Figs S1 and S2; Biryol et al., 2011; Schmandt & Humphreys, 2010; Scire et al., 2015). Nodes that are not sampled are assigned a hit quality of 0 while nodes that are sampled by multiple rays from all four geographical quadrants are assigned a hit quality of 1. As in Scire et al. (2015), the uppermost model layer (60 km) and the lowermost model layer (715 km) are not interpreted because these layers absorb effects from anomalies outside of the model space that are not accounted for in the crustal corrections or the global mantle velocity model, respectively.

The inverse problem is regularized using norm and gradient damping as well as model smoothing to account for uncertainties in the location of the ray paths, details of
which are discussed in Schmandt & Humphreys (2010). We also incorporate station and event terms in order to compensate for perturbations outside of the model space. Since our crustal thickness corrections are calculated using a homogenous layered velocity model, the station terms absorb local perturbations in the velocity structure that are not accounted for in the crustal corrections as well as any errors in the a priori crustal thickness. The event terms account for variations in mean velocity structure under the subset of stations that record a specific event. We use the LSQR algorithm of Paige & Saunders (1982) to invert the frequency-dependent relative travel time residuals in order to obtain velocity perturbations within the modeled volume. This algorithm aims to obtain the minimum energy/length model that satisfactorily explains the observed data. We performed a tradeoff analysis between the variance reduction and the Euclidean model norm to choose the overall damping and smoothing weights (Fig. S3). Using the chosen damping (6) and smoothing (5) parameters for both the P- and S-wave inversions results in variance reductions of 83.5% and 82.7% respectively.

Synthetic resolution tests were performed to determine the ability of our model to resolve mantle structures. A synthetic “checkerboard” input was created with alternating fast and slow velocity anomalies with amplitudes of +5% and -5% respectively for P-waves, defined in 8-node cubes (Fig. 3a), and +9% and -9% respectively for S-waves, defined in 27-node cubes (Fig. 3b). Larger “checkers” were used for the S-wave synthetic tests due to the decreased resolution of the S-wave inversion caused by the more limited number of rays used in the inversion. Resolution of the S-wave inversion is also affected by the frequency bands in which the S-wave arrivals were picked. The Fresnel zones for the S-wave arrivals are larger than for the P-wave arrivals due to the lower frequency
content of the S-wave arrivals since Fresnel zone radius is inversely proportional to frequency (Hung et al., 2000). The size and location of the input anomalies varies with depth due to the depth dependent dilation of the node spacing. In order to determine the extent of any lateral or vertical smearing, the input anomalies are separated by neutral 2-node cubes that contain no velocity anomalies. The output from the checkerboard tests shows that lateral resolution is good in the center of the model space, with smearing of anomalies increasing towards the edges of the model space due to the decrease in the number of crossing rays at the edges of the model space (Fig. 3). Some vertical smearing of anomalies is observed, resulting in low amplitude anomalies being erroneously resolved into neutral background layers in both the P- and S-wave resolution tests. The synthetic tests also indicate that there is a loss of amplitude in the output of the checkerboard tests. Only ~60% of the input amplitude is recovered in the center of the model space, with amplitude recovery decreasing towards the edges of the model. Similar loss of recovered amplitude occurs with depth, with lower amplitude recovery observed in the center of the model in the deepest depth slices.

A second set of synthetic tests were performed to determine the ability of the inversion to resolve the subducting Nazca slab, since it is one of the most prominent structures that we expect to observe. The Slab1.0 global subduction zone model (Hayes et al., 2012) was used to create a synthetic slab anomaly that is between ~70-100 km thick with amplitudes of +5% to input into the model in order to test the ability of the P-wave model to resolve the flat slab in the northern part of the study region and the transition to the steeply dipping slab in the south (Fig. 4). As was observed by Scire et al. (2015), recovery of the slab anomaly above 200 km is complicated by vertical smearing effects.
and decreased amplitude recovery. In the northern part of the study region, the recovered amplitude of the shallow, flat anomaly corresponding to the Peruvian flat slab is even more diminished, indicating that we probably cannot image the flat slab segment due to an insufficient number of crossing rays in the shallowest model layers. Below 200 km depth, the recovered slab anomaly is more distinct and increases in amplitude. In the deeper parts of the model, the recovered slab location corresponds well to the input slab location, indicating that we should be able to image the Nazca slab down to 660 km depth. Similar patterns are observed for slab recovery tests for the S-wave model for a synthetic slab with amplitude of +9% although less amplitude is recovered for the S-wave model especially on the edges of the model.

RESULTS

The results from this study are displayed as horizontal depth slices in Figs 5, 6, and 7, and vertical cross sections in Figs 8 and 9. Additional depth slices and cross sections are included in the Supplemental Material (Figs S5, S6, and S7). The model is resolved down to 660 km depth. In the shallowest depth slices, resolution is strongly controlled by the location of the stations and vertical smearing is more prominent. Horizontal resolution is good throughout the model, allowing for the interpretation of lateral differences in mantle structure. At all depths, model resolution decreases towards the edges of the model space. In the northern part of the study area, the shallowest depth slice that we are able to interpret, 95 km, is already within the lithosphere of the Peruvian flat slab, preventing us from making any observations of anomalies above the flat slab.
Therefore, any anomalies observed in our results in this region are mostly below the flat slab.

*The Nazca slab*

The subducting Nazca slab is the most prominent anomaly we observe in our tomograms. The Nazca slab is observed in both the P- and S-wave tomograms as a fast anomaly with variable amplitude that migrates northeast with depth (Figs 5, 6, and 7). Due to the complications in resolving shallow structure discussed previously, identifying the Nazca slab in the shallower parts of the model (particularly 95 to 130 km depth) is difficult (Fig. 5). In particular, resolution of the slab anomaly between 130 and 165 km depth below the active volcanic arc in the southern part of our study area is complicated by the presence of a high amplitude low velocity anomaly just east of the arc. Difficulty resolving the slab anomaly in the region where the subduction angle changes from steep to shallow near 14°S is also noted at 95 and 130 km depth possibly due to interference from high amplitude low velocity anomalies both northwest (below) and southeast (above) of the slab. At the northern edge of our study area, a fast anomaly that corresponds to the location of intermediate depth earthquakes between 95 and 165 km depth is noted although its connection to the slab anomaly to the south is outside of our resolution (Anomaly A, Fig. 5). The slab anomaly becomes more prominent and more continuous at greater depths, allowing us to make observations about the subducting Nazca slab down to 660 km depth. For the depth slices at 280 km and 365 km the shape of the Nazca plate is quite different than the Slab1.0 contours (Fig. 6). North of ~16°S the strike of the slab is more north-south than the contours and there is a sharp eastward
deflection in the slab strike prior to resuming a more north-south orientation following the contours southward. This “kink” is especially clear in the S-wave tomogram at 280 km depth but is less clear in the corresponding P-wave image. We note that this part of the slab has very little seismicity, possibly explaining the difference from the seismicity constrained Slab1.0 contours. The slab anomaly shows along-strike variations in amplitude in our tomograms (Figs 6 and 7). A bend in the slab anomaly is observed in the mantle transition zone (MTZ) between 14° and 18°S corresponding to the location of the Bolivian orocline (Fig. 7). North of the bend in the slab anomaly, the fast slab anomaly appears to thicken in the MTZ.

The shallow mantle (95-200 km)

The uppermost mantle south of 14°S above the normally dipping Nazca slab is highly variable. In the southernmost part of our study area, a -2% slow anomaly in the P-wave tomograms is observed under the western Altiplano (Anomaly B, Figs 5 and 8). The corresponding anomaly in the S-wave results is much higher amplitude (-6%). This anomaly is confined to the west by the subducting slab and to the northeast by a variable amplitude fast anomaly (Anomaly C, Figs 5, 8, and 9). The shape and size of this fast anomaly (Anomaly C) is highly variable although it appears to be confined to the Eastern Cordillera and Subandean Zone. Slow anomalies are noted to the northeast of this fast anomaly, indicating that its lateral extent is limited. A second fast (+3%) anomaly is noted in the P-wave tomograms under the Subandean Zone around 15°S (Anomaly D, Figs 5, 6, 8, and 9). This anomaly extends vertically in the mantle from ~95 km depth to 200 km depth.
The shallow mantle in the northern part of our study region is dominated by high amplitude slow anomalies in the mantle below the Peruvian flat slab (Anomaly E, Figs 5, 6, 8, and 9). These high amplitude anomalies extend vertically through several depth slices, particularly the southern part of the low velocity anomaly which appears to be confined to the area under the projection of the Nazca Ridge and extends from 95 to 280 km depth. A second high amplitude slow velocity anomaly is observed just to the north (Anomaly F, Figs 5 and 8). While similar in amplitude to Anomaly E, particularly at 95 km depth, this second slow anomaly is confined to the uppermost depth slices, 95 km and 130 km. The eastern edge of our resolution does not extend past the region where the Nazca slab begins to subduct steeply into the mantle inboard of the Peruvian flat slab so no observations of the mantle above the slab can be made.

The deeper mantle (200-660 km)

High amplitude low velocity anomalies in the S-wave tomograms (-5 to -6%) are seen in the MTZ between 410 and 660 km depth above the subducting Nazca slab at the eastern edge of the study area with some continuation of the low velocity anomaly into the upper mantle above 410 km depth (Anomaly G, Figs 7 and 8). Similar low velocity anomalies are seen in the P-wave tomograms above the slab in the MTZ although the amplitude of these anomalies is higher in the shallower MTZ (-3% at 410 km depth) and decreases with depth to -2% at 660 km depth.
DISCUSSION

Geometry of the subducting Nazca slab

The fast slab anomaly that we image in our tomograms allows us to make some observations about the geometry of the slab. We use our results to offer some modifications to the Slab1.0 global subduction zone model, which is primarily constrained by slab seismicity in our study region (Hayes et al., 2012). The slab contours in Fig. 10 were determined by tracing the center of the fast slab anomaly in the P-wave tomograms associated with the subducting slab at depths below 200 km on the depth slices. The location of the subducting slab above 200 km is constrained using a combination of our results, the location of intermediate depth seismicity from the PDE catalog (Engdahl et al., 1998), the Slab1.0 global subduction zone model (Hayes et al., 2012) and initial constraints on the location of the top of the Peruvian flat slab north of 14°S from receiver function work by Bishop et al. (2013). Work by Schneider and Sacks (1987) using relocated earthquakes is used to define the geometry and location of the change in slab dip from 100 to 200 km depth just to the south of the Nazca Ridge between 14° and 16°S.

Transitioning from flat to normal subduction and the effect of the Nazca Ridge

Intermediate depth seismicity marks the region where the slab dip changes from shallow subduction along the subducted Nazca Ridge to normal subduction under Bolivia to the south. Although this bend in the slab is not resolved at 95 or 130 km depths, probably due to interference from high amplitude low velocity anomalies at these depths, slab earthquakes mark the location of the slab at these depths and, when combined with
the location of the steepening slab inboard of the Nazca Ridge, indicate that the transition from shallow to normal subduction is marked by rapid bending of the slab in the shallow mantle directly south of the ridge. Seismicity from the PDE catalog (Fig. 5; Engdahl et al., 1998) combined with work by Schneider and Sacks (1987) allows us to roughly constrain the location of the subducting slab at 95, 130, and 165 km depth near 14°S and are taken into account in the slab contours in Fig. 10. By 200 km depth, the slab is imaged as a continuous fast anomaly, which bends in the mantle between the shallowly subducting Nazca Ridge and the normally subducting slab to the south.

At the northern edge of our study region, near 10°, the high amplitude fast anomaly (Anomaly A; Figs 5 and 9) appears to correspond to the slab earthquakes at 130 and 165 km depth; we interpret this anomaly as the Nazca slab. This anomaly is offset from the observed slab anomaly under the projection of the Nazca Ridge indicating that the slab has either undergone abrupt bending just north of the subducted Nazca Ridge or a small, ridge parallel tear in the slab exists along the north edge of the Nazca Ridge (Fig. 10). Although we are unable to directly image the slab in the region where the postulated tear exists, the lack of intermediate depth seismicity in the region combined with the offset in the slab anomaly leads us to conclude that a tear is possible on the north side of the ridge, unlike the southern side of the ridge where bending in the slab is marked by a limited region of intermediate depth seismicity. The offset in the slab anomaly north of the subducting Nazca Ridge decreases with depth until 365 km where a continuous slab anomaly is imaged, leading us to conclude that any possible tear is limited in extent (Fig. 10). Since we cannot directly image the region along the north side of the Nazca Ridge due to limitations in our resolution, confirmation of the existence of the hypothesized tear
will require future work in the area. Due to the uncertainty in our interpretation, we offer a set of slab contours which assume that the slab north of the Nazca Ridge is continuous (Fig. 10, Supplemental Material, Table S1).

Along the subducted projection of the Nazca Ridge (north of 14°S) the slab anomaly between 165 km depth and the top of the mantle transition zone, 410 km depth, appears to be further inland than is indicated by the Slab1.0 global subduction zone model contours (Fig 1b; Hayes et al., 2012). This indicates that the Peruvian flat slab extends further inland than was previously thought along the projection of the Nazca Ridge (Figs 6, 7, and 8). Since the slab inboard of the subducting Nazca Ridge is aseismic (Fig. 1b) and limited geophysical work has been done previously in this area, this segment of the Nazca slab is not well-constrained in the Slab1.0 global subduction zone model. Our results are in agreement with work by James & Snoke (1990), who used underside wide-angle reflections from the upper surface of the mostly aseismic part of the Nazca slab in this area to locate the top of the slab (Figs 5, 6, 7, and 8). This indicates that the dip of the Nazca slab once it re-steepens inboard of the flat slab region is very steep (between 70° to 75°) below depths of about 150 to 200 km. This region of slab steepening corresponds to the southeast edge of the projection of the subducted Nazca Ridge based on the conjugate feature, the Tuamotu Plateau (Fig. 6; Hampel, 2002), implying that the presence of the ridge is locally influencing the geometry of the flat slab.

The low velocity anomaly under the southern Peruvian flat slab (Anomaly E, Figs 5, 6, 8, and 9) could be associated with decompression melting of the sub-slab asthenosphere as mantle material moves north through the hypothesized tear in the slab along the north edge of the Nazca Ridge. This sub-slab low velocity anomaly is observed
at similar depths as the postulated slab tear (~95-280 km) and is limited to the region immediately beneath the subducted Nazca Ridge. The amplitude of this anomaly is much higher in the S-wave results, consistent with the presence of partial melt (Nakajima et al., 2001). Another possible explanation for the low velocity anomaly in the sub-slab mantle is the presence of relatively hot, asthenospheric mantle at anomalously shallow depths due to the existence of thinned oceanic lithosphere under the Nazca Ridge. A region of decreased elastic thickness associated with the Nazca Ridge has been proposed by Tassara et al. (2007), who suggest that it is related to weakening of the oceanic lithosphere due to interactions with known hotspots. Anomalously thin oceanic lithosphere is also observed in association with the Hawaiian hotspot (Li et al., 2004) leading us to conclude that similar thinning of the oceanic lithosphere under the Nazca Ridge is possible. Dating of some of the lavas that make up the Nazca Ridge indicates that much of the ridge is between 5-13 Myr younger than the seafloor onto which the ridge was emplaced (Ray et al., 2012). This implies that lithospheric cooling of the Nazca plate could have been disrupted by the Easter-Salas hotspot, which was located to the east of the axis of the East Pacific Rise during at least some of the formation of the Nazca Ridge, leading to a thinner lithosphere under the Nazca Ridge than expected considering the ~45 Ma age of the oceanic plate (Müller et al., 2008). The northern shallow sub-slab slow velocity anomaly (Anomaly F, Figs 5 and 8) could be related to a trench parallel tear at a depth of 60-80 km to the north of the ridge which has been postulated by Bishop et al. (2013) and Knezevic Antonijevic et al. (2014); however this tear is shallower than we are able to resolve (<100 km depth) and therefore we are unable to speculate further about the hypothesized tear and its relationship with the observed slow velocity anomaly.
Variations in the central Andean shallow mantle (95-200 km)

Between 95 and 130 km depth, we observe slow velocities under the western Altiplano in the southern part of our study area (Anomaly B, Figs 5 and 8) and a fast, variable anomaly under the Eastern Cordillera and Subandean Zone (Anomaly C, Figs 5 and 8). The low velocity anomaly, Anomaly B, is particularly prominent in the S-wave inversion and interferes with our ability to resolve a continuous, steeply dipping Nazca slab anomaly under Bolivia even though the location of the steeply dipping slab is marked by intermediate depth seismicity. This anomaly is consistent with the presence of a region of partial melting of the mantle above the subducting Nazca slab, which serves as the source region for the arc volcanism observed in the Western Cordillera. The higher amplitude of this anomaly in the S-wave tomograms in comparison to the P-wave tomograms is consistent with the presence of melt, which more strongly affect S-wave velocities (Nakajima et al., 2001). Similar anomalies associated with arc source regions have been seen in elsewhere the Andes in both teleseismic (e.g. Heit et al., 2008) and local (e.g. Schurr et al., 2006) tomography studies.

The fast anomaly (Anomaly C) that we observe under the Eastern Cordillera and Subandean Zone between 14° and 18° S in both the P- and S-wave inversions is bounded to the northeast by a well resolved low velocity anomaly (Fig. 5). Although other studies in the central Andes (e.g. Beck & Zandt, 2002; Dorbath et al., 1993; Myers et al. 1998; Phillips et al., 2012; Scire et al., 2015) have postulated underthrusting of Brazilian cratonic lithosphere as far west as the Eastern Cordillera, the presence of the low velocity anomaly inboard of Anomaly C suggests that this fast anomaly is unlikely to represent undisrupted cratonic lithosphere. As a possible alternative to underthrusting of Brazilian
cratonic lithosphere, Scire et al. (2015) proposed that a similar elongated fast anomaly under the Eastern Cordillera between 16° and 25°S could be related to foundering of cold lithospheric material. This interpretation implies that lithospheric material is being removed along much of the eastern edge of the Andes. The vertically elongated fast anomaly which descends from 95 to 200 km depth near 15°S under the Subandean Zone (Anomaly D; Figs 5, 6, 8, and 9) is similar to anomalies which have been observed in other parts of the Andes, particularly under the Puna Plateau to the south. Previous studies have interpreted these anomalies as delaminating blocks of lithosphere (e.g., Schurr et al., 2006; Koulakov et al., 2006; Bianchi et al., 2013; Scire et al., 2015; Beck et al., 2015). The presence of anomalously slow velocities at depths below 95 km in the mantle adjacent to the Anomaly D could be indicative of upwelling asthenosphere resulting from piecemeal delamination of the mantle lithosphere. Similar low velocities in the shallow mantle have been interpreted as asthenospheric mantle at shallow depths under the central Andes (e.g., Myers et al., 1998; Heit et al., 2008). Since our resolution is limited to depths below 95 km, we are unable to image the top of the hypothesized delaminating block or where it is connected to the Andean crust if such a connection exists. Therefore we are limited in our ability to interpret the timing of the hypothesized delamination event and its possible effects on the crust although future studies using methods with better resolution in the shallow mantle and crust should clarify the interpretation.
Deformation of the slab in the mantle transition zone

As previously discussed, the slab anomaly is easily identified and appears to be continuous below 200 km depth although it is further inland in the northern part of our study region than indicated by the Slab1.0 global subduction zone contours (Figs 6 and 7). A cluster of deep earthquakes (~600 km depth) is observed in the northern part of our study region. Additional isolated deep earthquakes occur between 500 and 600 km depth to the south of the large cluster. The location of those earthquakes falls within our fast slab anomaly, in agreement with the idea that deep focus earthquakes occur within the cold cores of slab in the MTZ (Kirby et al., 1996). In the region where we have resolution northeast of the slab anomaly between 8° and 14°S, we do not see any evidence that the Nazca slab has stagnated for any significant distance to the east in the MTZ. This agrees with global tomography studies which observe a diffuse fast anomaly in the lower mantle which corresponds with the location of the subducting Nazca slab and indicates that the Nazca slab does not experience long-term stagnation in the MTZ and instead continues subducting into the lower mantle (e.g., Bijwaard et al., 1998; Li et al., 2008; Zhao, 2004; Fukao et al., 2001; 2009; Zhao et al., 2013). Our results do show a region of localized widening of the Nazca slab anomaly in the MTZ between 8° and 14°S, similar to observations made further south (Scire et al., 2015). Thickening of subducting slabs in the MTZ has been observed in global tomography studies and has been interpreted as evidence of temporary stagnation in the MTZ, resulting in thickening of the slab before subduction resumes into the lower mantle (e.g. Bijwaard et al., 1998; Li et al., 2008) or folding of the slab to accommodate a decreasing sinking velocity of the slab in the lower mantle (e.g., Ribe et al., 2007). The location of this region of thickening corresponds to
the very steep (~70°) dip of subduction due to the effect of the Peruvian flat slab on the inboard location of the Nazca slab and is possibly influenced by the Nazca slab intersecting the base of the MTZ at a near vertical angle. Between 14° and 17°S, the Nazca slab anomaly bends and thins in the MTZ. This is particularly apparent in the P-wave tomography, where the slab anomaly changes strike from almost due east to a north-south orientation in the lower MTZ (Fig. 9). This change in strike of the slab in the MTZ appears to correspond to the location of the Bolivian orocline and possibly represents the translation of the orocline from the surface down to mantle transition zone depths.

The high amplitude, low velocity anomaly observed in the MTZ above the Nazca slab east of 67°W between 10° and 17°S (Anomaly G, Figs 7 and 8) is similar to anomalies seen further south (Scire et al., 2015). Scire et al. (2015) argued that low velocity anomalies in the MTZ could be explained either by local thermal anomalies or by localized regions of hydration in the MTZ since both local increases in temperature or water content in the MTZ would lower seismic velocities. However, we are unable to distinguish between these two possible causes with seismic velocities alone. Low velocity anomalies are observed in the sub-slab mantle in other subduction zones (e.g., Obayashi et al., 2006; Zhao, 2004; Zhao et al., 2013). A low velocity anomaly directly above the 410 km discontinuity was observed under the subducting slab in the Honshu subduction zone and was interpreted as a local thermal anomaly (Bagley et al., 2009; Obayashi et al., 2006; Morishige et al., 2010). The other possible cause of the low velocity anomaly in the MTZ, a region of locally elevated hydration, is suggested further south by Schmerr & Garnero (2007), who observed a depressed 410 km discontinuity which they argued
could not be explained by a solely thermal anomaly. Some studies suggest that subducting slabs can carry water into the MTZ (e.g., Bercovici & Karato, 2003; Smyth & Jacobsen, 2006), and direct evidence for hydration of MTZ minerals has been observed recently in a sample of ringwoodite discovered as an inclusion in a diamond (Pearson et al., 2014). Therefore, a region of locally enhanced hydration in the MTZ is a possible explanation for our observed low velocity anomaly.

CONCLUSION

Previous seismic studies in southern Peru and northern Bolivia have been limited in their ability to image the deeper mantle under the region. Our teleseismic tomography study, while limited in its ability to resolve structures in the shallowest mantle, shows new P- and S-wave tomograms of the subducting Nazca slab down to the base of the MTZ, including the transition from shallow, flat slab subduction in the north to more normal, steeply dipping subduction in the south (Fig. 11). We propose new slab contours for the Nazca slab between 8° and 20°S from 100 to 660 km depth based on a combination of our results and work in the region by other studies (Fig. 10; Supplemental Material Table S1). We tentatively suggest a possible slab tear along the north side of the Nazca Ridge at 10°S between 71° and 73°W (Fig. 10) although due to limitations of our resolution to the north of the Nazca Ridge, confirmation of the existence of this tear will rely on future work. The high amplitude low velocity anomaly we observe under the Nazca Ridge (Anomaly E, Fig. 11) could be explained either by the flow of sub-slab mantle through the proposed tear or the presence of asthenospheric mantle at anomalously shallow depths due to disruption of the formation of lithosphere by the
Easter-Salas hotspot. Our results show a vertically elongated fast anomaly under the Subandean Zone between 95 and 200 km depth (Anomaly D, Fig. 11). Considering both the elongated vertical geometry of the fast anomaly and the adjacent low velocity anomalies in both the P- and S-wave models, we propose that this corresponds to a block of delaminating lithosphere. North of 14°S, we image the Nazca slab to the east of its predicted location from the Slab1.0 global subduction zone model where it steepens inboard of the Peruvian flat slab region. This indicates that the slab remains flat further inland than was previously thought, and that when it steepens and resumes subduction into the mantle, it does so at a very steep (~70°) angle. This region of slab steepening corresponds with the predicted edge of the subducted Nazca Ridge (Hampel, 2002), implying that the geometry of the Nazca slab is at least locally being influenced by the presence of the ridge. When this very steeply dipping slab hits the 660 km discontinuity at the base of the MTZ, we observe local thickening of the Nazca slab between 8° and 14°S, consistent with temporary stagnation of the slab in the MTZ before it resumes subduction into the lower mantle. We also observe high amplitude low velocity anomalies above the slab east of 67°W between 10° and 17°S in the MTZ, which we have interpreted as representing either local thermal anomalies or regions of locally increased hydration.

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REFERENCES


Burd, A.I., Booker, J.R., Mackie, R., Pomposiello, C. & Favetto, A., 2013. Electrical conductivity of the Pampean shallow subduction region of Argentina near 33 S:


Figure 1. A) Map showing seismic station locations for CAUGHT and PULSE networks as well as additional stations used in the study and topography of the central Andes.
Figure 2. A) Global map centered on our study region (black square) showing the location of P-wave events used in this study. Darker gray circles mark events used for direct P arrivals while open circles mark events used for PKIKP arrivals. B) Plot showing the azimuthal distribution for all rays used for P-waves in the study. Ray distribution is strongly controlled by the location of plate boundaries where the earthquakes are generated. C) Global map centered on our study region (black square) showing the location of S-wave events used in this study. D) Plot showing the azimuthal distribution for all rays used for S-waves in the study. Ray distribution is strongly controlled by the location of plate boundaries where the earthquakes are generated.
Figure 3. Horizontal depth slices for the checkerboard tests for a selection of model layers. Input for neutral layers (0% velocity deviation) is not shown. Output for neutral layers is shown in the left column. A) Checkerboard tests for P-wave inversion. The checkerboard tests show that for shallower layers, resolution is controlled by station distribution as expected. Deeper model layers indicate that while the input amplitude cannot be completely resolved, lateral changes in anomalous velocity resolve with little
horizontal smearing. Resolution is lost towards the edges of the model region. Resolution of velocity anomalies in neutral layers shown here (left column) indicates that some vertical smearing is occurring. B) Checkerboard tests for S-wave inversion. Larger input “checkers” are used for S-waves due to decreased resolution of S-wave inversion relative to P-wave inversion. Amplitude recovery for the S-wave inversion decreases more rapidly with depth than for the P-wave inversion. Resolution of velocity anomalies in neutral layers shown here (left column) indicates that some vertical smearing is occurring. Checkerboard tests for additional model layers are in Supplemental Material, Fig. S4.
Figure 4. Cross-section results for our synthetic slab recovery tests. Cross section locations are shown in Figure 1. The geometry of our input slab model is based on the Slab1.0 contours (black line, Hayes et al., 2012). Synthetic input (left), recovered Vp (center), and recovered Vs (right) models are shown. Decreased amplitude recovery and vertical smearing of the recovered slab anomaly is observed in the upper 200 km of the model. In general, the amplitude recovery increases with depth.
Figure 5. Horizontal depth slices for 95, 130, and 165 km from the tomography model for both Vp (left) and Vs (right). The short dashed line represents the edge of the well-resolved region of the model, defined as regions with hit quality greater than 0.2 (Biryol et al., 2011). Geomorphic provinces (green lines) are the same as in Figure 1A. Stars mark locations of seismic stations used in the study. Red triangles mark location of Holocene volcanoes (Siebert & Simkin, 2002). Yellow dots are earthquake locations from the EHB catalog (Engdahl et al., 1998). Solid black lines are slab contours from
Slab1.0 model (Hayes et al., 2012). Orange dots mark reflection points from James & Snoke (1990). Heavy black outline marks projection of the subducted Nazca Ridge from Hampel (2002). Labeled anomalies (A, B, C, D, E, and F) are discussed in the text.

Figure 6. Horizontal depth slices for 200, 280, and 365 km from the tomography model for both Vp (left) and Vs (right). Dashed lines are the same as in Figure 5. Stars mark
locations of seismic stations used in the study. Red triangles mark location of Holocene volcanoes (Siebert & Simkin, 2002). Yellow dots are earthquake locations from the EHB catalog (Engdahl et al., 1998). Solid black lines are slab contours from Slab1.0 model (Hayes et al., 2012). Orange dots mark reflection points from James & Snoke (1990). Heavy black outline marks projection of the subducted Nazca Ridge from Hampel (2002). Labeled anomalies (A, D, E, and G) are discussed in the text.
Figure 7. Horizontal depth slices for 410, 505, and 605 km from the tomography model for both Vp (left) and Vs (right). Dashed lines are the same as in Figure 5. Stars mark locations of seismic stations used in the study. Red triangles mark location of Holocene volcanoes (Siebert & Simkin, 2002). Yellow dots are earthquake locations from the EHB catalog (Engdahl et al., 1998). Solid black lines are slab contours from Slab1.0 model (Hayes et al., 2012). Orange dots mark reflection points from James & Snoke (1990). Labeled anomaly G is discussed in the text.
Figure 8. Trench perpendicular cross sections through the tomography model for both Vp (left) and Vs (right). Cross section locations are as shown in Figure 1. Dashed lines are the same as in Figure 5. Yellow dots are earthquake locations from the EHB catalog (Engdahl et al., 1998). Solid black line marks the top of the Nazca slab from the Slab1.0 model (Hayes et al., 2012). Orange dots mark reflection points from James & Snoke (1990). Labeled anomalies (B, C, D, E, F, and G) are discussed in the text.

Figure 9. Trench parallel cross sections through the Vp tomography model. Cross section locations are as shown in Figure 1. Dashed lines are the same as in Figure 5. Yellow dots are earthquake locations from the EHB catalog (Engdahl et al., 1998). Solid black line marks the top of the Nazca slab from the Slab1.0 model (Hayes et al., 2012). Orange dots mark reflection points from James & Snoke (1990). Labeled anomalies (A, C, D, and E) are discussed in the text. Cross sections through the Vs tomography model shown in the Supplemental Material, Fig. S7.
Figure 10. A) Slab contours determined from this study. Red contours are continuous slab contours determined using constraints discussed in the text (Supplemental Material, Table 1). Yellow contours are disrupted by the hypothesized tear north of the Nazca Ridge. Red triangles mark location of Holocene volcanoes (Siebert & Simkin, 2002). Gray outline marks projection of the subducted Nazca Ridge from Hampel (2002). B) 3-D mesh of continuous slab contours without hypothesized tear. Mesh surface and slab contours were created by using a linear interpolation between data points from our results (black dots). C) 3-D mesh of slab contours with hypothesized tear on the north side of the Nazca Ridge.
Figure 11. 3-D diagram of the resolved subducting Nazca slab and prominent mantle low velocity anomalies inferred from our tomographic models. The isosurfaces for this diagram are obtained by tracing the most coherent low velocity anomalies (less than negative 3%) and slab related (greater than positive 3%), coherent fast anomalies in the tomographic model. Geomorphic provinces (fine dashed lines) are the same as in Figure 1A. Heavy black outline marks projection of the subducted Nazca Ridge from Hampel (2002). Anomalies A, C, D, and E labeled as in previous figures.
Figure S1. Normalized hit quality plots for P-wave tomography model. Black stars on 95 km depth layer are station locations. Hit quality for a node is based on the number and
azimuthal distribution of rays that sample that node. Good hit quality is indicated by red shading while poor hit quality is indicated by blue shading. Hit quality is strongly controlled by station distribution in uppermost depth slices with decreasing dependence on station distribution as depth increases.
Figure S2. Normalized hit quality plots for S-wave tomography model. Black stars on 95 km depth layer are station locations. Hit quality for a node is based on the number and azimuthal distribution of rays that sample that node. Good hit quality is indicated by red shading while poor hit quality is indicated by blue shading. Hit quality is strongly controlled by station distribution in uppermost depth slices with decreasing dependence on station distribution as depth increases.

Figure S3. Plot of tradeoff analysis for the P-wave inversion between the variance reduction and the Euclidean model norm (L2) performed to choose preferred overall damping (D1-D10) and smoothing (S1 –S10) weights. The black star shows the damping (D6) and smoothing (S5) parameters used in this study.
Figure S4. Horizontal depth slices for the checkerboard tests for a selection of model layers. Input for neutral layers (0% velocity deviation) is not shown. Output for neutral layers is shown in the left column. A) Checkerboard tests for P-wave inversion. The checkerboard tests show that for shallower layers, resolution is controlled by station distribution as expected. Deeper model layers indicate that while the input amplitude cannot be completely resolved, lateral changes in anomalous velocity resolve with little horizontal smearing. Resolution is lost towards the edges of the model region. Resolution
of velocity anomalies in neutral layers shown here (left column) indicates that some vertical smearing is occurring. B) Checkerboard tests for S-wave inversion. Larger input “checkers” are used for S-waves due to decreased resolution of S-wave inversion relative to P-wave inversion. Amplitude recovery for the S-wave inversion decreases more rapidly with depth than for the P-wave inversion. Resolution of velocity anomalies in neutral layers shown here (left column) indicates that some vertical smearing is occurring.

Figure S5. Horizontal depth slices for 240 and 320 km from the tomography model for both Vp (left) and Vs (right). Dashed lines are the same as in Figure 5. Stars mark locations of seismic stations used in the study. Red triangles mark location of Holocene volcanoes (Siebert & Simkin, 2002). Yellow dots are earthquake locations from the EHB catalog (Engdahl et al., 1998). Solid black lines are slab contours from Slab1.0 model (Hayes et al., 2012). Heavy black outline marks projection of the subducted Nazca Ridge from Hampel (2002). Labeled anomalies (A, D, and E) are discussed in the text.
Figure S6. Horizontal depth slices for 455, 555, and 660 km from the tomography model for both Vp (left) and Vs (right). Dashed lines are the same as in Figure 5. Stars mark locations of seismic stations used in the study. Red triangles mark location of Holocene volcanoes (Siebert & Simkin, 2002). Yellow dots are earthquake locations from the EHB.
catalog (Engdahl et al., 1998). Solid black lines are slab contours from Slab1.0 model (Hayes et al., 2012). Labeled anomaly G is discussed in the text.

Figure S7. Trench parallel cross sections through the Vs tomography model. Cross section locations are as shown in Figure 1. Dashed lines are the same as in Figure 5. Yellow dots are earthquake locations from the EHB catalog (Engdahl et al., 1998). Solid black line marks the top of the Nazca slab from the Slab1.0 model (Hayes et al., 2012). Orange dots mark reflection points from James & Snoke (1990). Labeled anomalies (A, C, D, and E) are discussed in the text.
APPENDIX C: THE DEFORMING NAZCA SLAB IN THE MANTLE TRANSITION ZONE AND LOWER MANTLE: CONSTRAINTS FROM TELESEISMIC TOMOGRAPHY ON THE DEEPLY SUBDUCTED SLAB BETWEEN 6° AND 32°S

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ABSTRACT

We present new tomographic images of the Nazca slab under South America from 6° to 32°S from 95 km to lower mantle depths (895 km). By combining data from fourteen separate networks in the central Andes we are able to use finite frequency teleseismic P-wave tomography to image the Nazca slab from the upper mantle into the mantle transition zone (MTZ) and the upper part of the lower mantle. We image the Nazca slab penetrating into the lower mantle after undergoing some deformation in the MTZ. Our tomographic images show that there is significant along-strike variation in the morphology of the Nazca slab in the MTZ and the lower mantle. Thickening of the slab in the MTZ is observed north of the Bolivian, possibly related to buckling or folding of the slab in response to the increased viscosity of the lower mantle which decreases the sinking velocity of the slab. South of the orocline, the slab appears to continue into the lower mantle with only minor deformation in the MTZ. In the lower mantle, a similar difference in morphology is observed. North of 16°S, the slab anomaly in the lower mantle appears to be more coherent and penetrates more steeply into the lower mantle. To the south the slab appears to be flattening just below the 660 km discontinuity and displays some along-strike segmentation of the slab anomaly at depths greater than 800 km. This change in slab morphology in the MTZ and lower mantle appears to correspond to the change in the dip of the slab as it enters the MTZ, from steeply dipping in the north to more moderately dipping in the south.
INTRODUCTION

The Nazca-South America subduction zone is a long-lived plate boundary system which shows significant along-strike variations in subduction angle. Subduction of the Nazca plate under the central Andes of South America occurs at an angle of ~30° down to depths of about 300 km along much of the margin, with two regions of flat or shallow subduction occurring under Peru, north of 14°S, and under central Chile and Argentina, south of 30°S (Cahill and Isacks, 1992; Hayes et al., 2012). North of 14°S, the Nazca slab subducts at a shallow angle for several hundred kilometers inland (Cahill and Isacks, 1992; Hayes et al., 2012; Bishop et al., 2013) before resuming subduction at a very steep angle (~70°) down to the bottom of the mantle transition zone (MTZ) (Scire et al., 2015b). We combined data from fourteen separate networks deployed in the central Andes at various times from 1994 to 2013 to image the subducting Nazca slab down to 895 km depth between 6° and 32°S (Fig. 1). Previous studies by the authors (Scire et al., 2015a; 2015b) used subsets of this dataset to examine variations in upper mantle structure down to 660 km depth. By combining the two datasets we are able to increase the depth to which we can resolve velocity anomalies in the mantle due to the increased aperture of the array at the surface. In this paper we will focus on the geometry of the Nazca slab in the MTZ and lower mantle, providing new constraints on the interaction between the subducting Nazca slab and the transition into the lower mantle.

Tomographic images of the deeply subducted Nazca slab

Variations of the width of the Nazca slab in the MTZ have been observed using teleseismic tomography, indicating that some deformation of the Nazca slab is occurring
in the MTZ, possibly related to resistance to continued subduction through the 660 km
discontinuity into the lower mantle (Scire et al., 2015a, 2015b). Increases in slab
thickness in the MTZ have also been observed in global tomography studies and have
been interpreted as evidence for temporary stagnation of some subducting slabs in the
MTZ, resulting in thickening and deformation of the slab in the MTZ before subduction
continues into the lower mantle below the 660 km discontinuity (e.g. Bijwaard et al.,
1998; Li et al., 2008). Receiver function studies (Liu et al., 2003) and modeling studies
(Quinteros and Sobolev, 2013) indicate a correlation between an abnormally thick MTZ
and resistance to subduction of the Nazca plate into the lower mantle, in agreement with
the images of the temporarily stagnated Nazca slab in the tomography studies discussed
previously. The imaged morphology of subducting slabs in the MTZ varies globally, with
some slabs penetrating into the lower mantle with only minor broadening in the MTZ
while other subducting slabs flatten out and stagnate in the MTZ for long periods (e.g.,
Bijwaard et al., 1998; Li et al., 2008; Fukao et al., 2009). In the Nazca-South America
subduction zone, global tomography studies observe a diffuse fast anomaly in the lower
mantle which corresponds to the location of the subducting Nazca slab, indicating that
while deformation of the Nazca slab may occur in the MTZ, the slab continues to subduct
into the lower mantle and does not experience long-term stagnation in the MTZ (e.g.,
Bijwaard et al., 1998; Li et al., 2008; Zhao, 2004; Fukao et al., 2001; 2009; Zhao et al.,
2013). Similarly, Engdahl et al. (1995) observed an amorphous fast anomaly in the lower
mantle associated with the subducting Nazca slab using teleseismic and regional P phases
to image the Nazca-South America subduction zone between 5° and 25°S down to ~1400
km depth. Using teleseismic traveltime tomography, Rocha et al. (2011) imaged the
Brazilian mantle on the eastern side of South America between 38° to 58°W. At the very westernmost edge of their study area (57-58°W), they image a fast anomaly from 700 to 1300 km depth which they interpret as being the Nazca slab in the lower mantle although its geometry it difficult to resolve due to vertical smearing on the edge of their model space. South of 30°S, Pesicek et al. (2012) used a nested global-regional tomography model to image the Nazca subduction zone down to 1500 km depth. They image a detached segment of Nazca slab below ~200 km depth to the south of 38°S, which they interpret as being a result of a tear corresponding to the location of the Mocha Fracture Zone. North of the hypothesized tear, Pesicek et al. (2012) image the Nazca slab in the transition zone but the slab passes out of the eastern edge of their model space before reaching the lower mantle.

DATA AND METHODS

The dataset used in this study consists of 384 short period and broadband seismometers deployed in the central Andes as part of fourteen separate networks between 1994 and 2013 (Fig. 1). The networks include thirteen temporary deployments with varying numbers of stations. The dataset also includes data from permanent stations of the Plate Boundary Observatory network which were deployed before 2013. Two permanent stations from the Global Seismograph Network, LVC and NNA, and one permanent station from the Global Telemetered Seismograph Network, LPAZ, are also included. Arrival times were picked for direct P phases for 402 earthquakes with magnitudes greater than 5.0 between 30° and 90° away from the study region (Fig. 2a). Additional arrivals were picked for PKIKP phases for 144 earthquakes with a similar
magnitude limit between 155° and 180° away from the study region. As in previous studies by the authors (Scire et al., 2015a; 2015b) arrivals were picked using the multi-channel cross correlation technique described by VanDecar and Crosson (1990) and modified by Pavlis and Vernon (2010). A total of 27,435 direct P arrivals were picked for the broadband stations in three frequency bands (0.04 to 0.16 Hz, 0.1 to 0.4 Hz, and 0.2 to 0.8 Hz) and for the short period stations in one frequency band (0.5 to 1.5 Hz) due to the limited frequency content of data from the short period stations. An additional 10,055 PKIKP arrivals were picked in the same three frequency bands for the broadband stations. Distribution of the picks in the three frequency bands is dominated by the 0.2 to 0.8 Hz frequency band, with 44% of all picks in the 0.2 to 0.8 Hz frequency band, 24% in the 0.1 to 0.4 Hz frequency band, and 21% in the 0.04 to 0.16 Hz frequency band. The 0.5 to 1.5 Hz frequency band contains the fewest picks, 11% of the total number of picks, due to the limited duration of the deployments of most of the short period stations and the lack of PKIKP picks for these stations. The azimuthal distribution of incoming rays is uneven, with the majority of events occurring in either the Middle America subduction zone to the northwest or the South Sandwich subduction zone to the southeast (Fig. 2b).

Travel time residuals were calculated relative to IASP91 (Kennett and Engdahl, 1991) and then demeaned for each event to calculate the relative travel time residuals. Travel times were corrected for crustal variations using Moho depth estimates from receiver functions from Bishop et al., (2013) and Ryan et al., (2012) and crustal thickness estimates from Tassara and Echaurren (2012) in order to compensate for crustal heterogeneity (Schmandt and Humphreys, 2010). Crustal corrections were made using a homogenous layered velocity model corresponding to the average velocity of the central
Andean crust (Zandt et al., 1996; Swenson et al., 2000; Dorbath and Masson, 2000). Crustal corrections for stations in the forearc were calculated using a faster velocity model based on local tomography studies (Graeber and Asch, 1999; Koulakov et al., 2006; Schurr et al., 2006) and ambient noise tomography which covers our study region (Ward et al., 2013, 2014).

This study uses a finite-frequency teleseismic tomography algorithm discussed in detail by Schmandt and Humphreys (2010) and used in previous studies by the authors (Scire et al., 2015a; 2015b). The finite-frequency approximation defines the sampled area around the geometrical ray path by the first Fresnel zone, whose width is dependent on both the frequency and the distance along the ray path (Hung et al., 2000; Dahlen et al., 2000). The sampling in each model layer is determined by the width of the Fresnel zone, which increases with distance from the source or receiver, and the differential sensitivity within the Fresnel zone, resulting in theoretical “banana-doughnut” sensitivity kernels (Dahlen et al., 2000). A detailed discussion of the specific algorithm used to calculate the sensitivity kernels can be found in Schmandt and Humphreys (2010).

The model space is parameterized into a series of nodes in a non-uniform grid which dilates with increasing depth and distance from the center of the model. In the shallowest layers of the model, nodes are spaced 35 km apart in the center of the model where sampling is densest, increasing to 56 km apart at the edges of the model. Vertical node spacing increases with depth, from 35 km at 60-95 km to 60 km at greater than 715 km depth. The horizontal node spacing in the shallowest layers varies from 35 to 56 km and also increases with depth such that the horizontal node spacing varies from 52 to 83 km at a depth of 940 km. The uppermost model layer (60 km) and the lowermost model
layer (940 km) are removed from any interpretation because these layers absorb effects from anomalies outside of the model space which are not corrected for in the crustal corrections or the global mantle velocity model respectively. In addition, station and event terms are incorporated in order to compensate for perturbations outside of the model space. Since a homogenous layered velocity model is used to calculate the crustal corrections, the station terms absorb local perturbations in the velocity structure as well as any errors in the a priori crustal thickness. The event terms account for variations in mean velocity structure under the subset of stations that record a specific event.

The LSQR algorithm of Paige and Saunders (1982) is used to invert the frequency-dependent relative travel time residuals to calculate velocity perturbations within the modeled volume by obtaining the minimum energy/length model that satisfactorily explains the observed data. The inverse problem is regularized using norm and gradient damping as well as model smoothing to account for uncertainties in the location of the ray paths, details of which are discussed in Schmandt and Humphreys (2010). A tradeoff analysis was performed between the variance reduction and the Euclidean model norm to choose the overall damping and smoothing weights (Supplemental Material, Fig. S1). Using the chosen damping (5) and smoothing (4) parameters results in a variance reduction of 82.3%.

**Sampling in the mantle transition zone**

Sampling of the model space is determined by calculating normalized hit quality maps in which each node is assigned a hit quality value between 0 and 1 that is a function of both the number of rays sampling that node as well as the azimuthal distribution of
those rays (Fig. 3; Biryol et al., 2011; Schmandt and Humphreys, 2010). The calculation of the hit quality maps relies on the idea that better sampling of a node is achieved with the intersection of rays from multiple azimuths. A node that has been sampled by multiple rays from all four of the geographical quadrants is assigned a hit quality of 1 while a node that is not sampled at all has a hit quality of 0. Comparisons of the hit quality maps for the combined data set used in the current study with the maps for the subsets of data used previously by the authors (Scire et al., 2015a, 2015b) shows that combining the two smaller datasets has resulted in a substantial increase in hit quality, particularly at mantle transition zone depths (Fig. 3). This indicates that our combined data should be better able to resolve velocity anomalies in the MTZ as well as in the lower mantle due to the increased sampling at these depths.

Resolution and slab recovery

Synthetic resolution tests were performed to determine the ability of our model to resolve structures in the MTZ and lower mantle. A synthetic “checkerboard” input was created with alternating fast and slow velocity anomalies with amplitudes of +5% and -5% respectively defined in 8-node cubes (Fig. 4). The size and location of the synthetic input anomalies changes with depth due to the dilation of the node spacing with depth and distance from the center of the model. These anomalies are separated by neutral two-node cubes, which have no velocity anomalies, in order to assess the extent of any vertical or lateral smearing. The output from the checkerboard tests shows that lateral resolution is good in the center of the model space with increased lateral smearing and decreased amplitude recovery observed towards the edges of the model space (Fig. 4). On
the eastern edge of the model space in the deepest layers where we expect the Nazca slab to be located in the MTZ and lower mantle, amplitude recovery is decreased and lateral smearing has increased from the center of the model space but ~30% of the input anomaly is still being recovered. Lateral smearing at the eastern edge of the model space in the lower mantle depth slices is dominantly to the southeast due to large number of rays coming from the South Sandwich subduction zone (Fig. 2b). Neutral layers (505 and 660 km) show evidence of vertical smearing with low amplitude anomalies being erroneously resolved into neutral background layers.

Additional synthetic tests were performed to assess the ability of our inversion to resolve the subducting Nazca slab in the MTZ and lower mantle. The Slab1.0 global subduction zone model (Hayes et al., 2012) was used to create a synthetic slab anomaly that is between ~70-100 km thick with amplitudes of +5% to input into the model in order to test the ability of the inversion to resolve the Nazca slab (Fig. 5). Since the Slab1.0 model ends between 600 and 660 km depth due to limited information about the structure of the Nazca slab in the lower mantle, the synthetic Nazca slab was taken from the 660 km depth slice and projected into the lower mantle depth slices (715-940 km). The recovered slab anomaly from 300-900 km depth is continuous with amplitudes of ~60% of the input amplitude. A slight decrease in amplitude recovery is noted with increasing depth but the continuity of the recovered slab anomaly remains at all depths. Limited broadening of the recovered slab anomaly is observed, indicating that variations in the width of the slab anomaly in our results probably represent real variations in the thickness of the Nazca slab. As in the checkerboard tests (Fig. 4), some smearing of the slab anomaly is observed at the eastern edge of the model space in the lower mantle depth
slices (e.g., 895 km) although the amplitude of the smeared slab anomaly decreases rapidly (Fig. 5).

RESULTS AND DISCUSSION

We present our results in horizontal depth slices (Figs 6 and 7) and vertical cross sections (Fig. 8). Since detailed discussions of the shallow mantle have previously been published for subsets of this dataset (Scire et al., 2015a, 2015b), we will focus our discussion on the deformation of the Nazca slab in the mantle transition zone (MTZ) and the lower mantle. The Nazca slab is observed as a trench parallel fast anomaly in the MTZ (Fig. 6). The fast anomaly continues into the lower mantle as a fairly continuous anomaly at 715 and 775 km depth (Fig. 7). Resolution of the Nazca slab is complicated by the lateral smearing along the eastern edge of the model space as depth increases which was noted in the synthetic tests discussed previously (Figs 4 and 5). The fast Nazca slab is still observable but the preferentially southeast lateral smearing must be taken into account when interpreting the geometry of the slab at 835 and 895 km depth. A high amplitude slow anomaly (-3 to -4%) is observed in the MTZ above (to the east) of the slab anomaly north of 18°S (Fig. 6). This anomaly appears to continue vertically into the upper mantle (Fig. 8, cross sections D, E, F).

*Imaging the Nazca slab*

The fast slab anomaly in the MTZ appears to be mostly continuous (Figs 6 and 8). Some along-strike changes in slab width are observed, particularly to the north of 16°S where the slab anomaly is much broader than to the south. The slab anomaly broadens
with depth in the MTZ, with a much thicker slab being observed in the northern part of our study region at 660 km depth than at 410 km depth. The slab anomaly bends in the MTZ between 15° and 17°S, paralleling the change of the strike of the trench at the surface associated with the Bolivian orocline. South of this change in strike the slab anomaly is mostly continuous with a consistent thickness other than south of 26°S where some minor thickening of the slab anomaly is observed (Fig. 6).

Our observations of the Nazca slab in the lower mantle (Figs 7 and 8) are consistent with global tomography studies, many of which image a diffuse fast anomaly in the lower mantle associated with the subducting Nazca slab (e.g., Bijwaard et al., 1998; Li et al., 2008; Zhao, 2004; Fukao et al., 2001; 2009; Zhao et al., 2013). However, we are able to offer higher resolution images of the slab and along-strike variations in the slab geometry in the lower mantle. The slab anomaly generally broadens in the lower mantle. While the slab anomaly is fairly continuous at 715 and 775 km depth, segmentation of the slab anomaly is observed below 800 km depth (Fig. 7). This segmentation occurs primarily south of the bend in the slab anomaly associated with the Bolivian orocline. While some of the eastward smearing of the slab anomaly in the southern part of our study region is probably due to the ray path distribution as discussed previously, the along-strike segmentation of the slab does not appear to be an effect of the decreasing resolution with depth since similar segmentation is not observed in the synthetic slab recovery tests (Fig. 5). North of the orocline, the slab anomaly is more coherent with higher amplitudes being resolved than in the southern part of the study region.
The deforming Nazca slab

The thickening of the Nazca slab in the MTZ that we observe in the northern part of our study region and to a lesser extent to the south of 26°S (Fig. 6) is similar to observations made by global tomography studies which interpret the observed widening of the slab in the MTZ as a result of temporary stagnation as the slab resists subduction into the lower mantle (e.g., Bijwaard et al., 1998; Li et al., 2008; Zhao, 2004; Fukao et al., 2001; 2009; Zhao et al., 2013). Since we do not see any evidence for continued stagnation to the east, we conclude that the stagnation is only temporary, unlike other slabs like the Pacific slab under East Asia which remains stagnant in the MTZ for several hundred kilometers (e.g., Zhao 2004; Fukao et al., 2001, 2009; Li et al., 2008; Zhao et al., 2013; Fukao and Obayashi, 2013). Several studies have suggested that folding or buckling of the slab occurs in the MTZ to accommodate a decreasing sinking velocity of the slab in the lower mantle due to the viscosity increase across the 660 km discontinuity (e.g., Ribe et al., 2007; Běhounková and Čížková, 2008; Lee and King, 2011; Garel et al., 2014). This would result in the observed thickening of the slab anomaly in the MTZ towards the northern end of our study region (Figs 6 and 8). Běhounková and Čížková (2008) concluded that buckling or folding of the slab related to the viscosity increase between the upper and lower mantle would explain the imaging of slab as diffuse fast anomalies in the lower mantle rather than the narrower slab anomalies which are observed in the upper mantle. Little to no thickening of the slab anomaly is observed in the MTZ to the south of 16°S, indicating that deformation of the slab in the MTZ varies along-strike (Fig. 8).
Focal mechanisms for events in the MTZ (Fig. 6) indicate that the Nazca slab is generally in a state of downdip compression, consistent with the idea that a viscosity increase occurs across the 660 km discontinuity separating the upper and lower mantle and results in increased resistance to subduction into the lower mantle (e.g., Isacks and Molnar 1971; Vassilou, 1984). Deviations from this broad trend of downdip compression could be explained by other factors which have been hypothesized to contribute to deep seismicity including transformational faulting associated with phase changes of olivine in cold subducting slabs in the MTZ (e.g., Wiens et al., 1993; Kirby et al., 1996; Guest et al., 2003, 2004) and thermal shear instability (e.g., Ogawa, 1987; Kanamori et al., 1998; Karato et al., 2001). The fault planes of the focal mechanisms change orientation along strike, following the strike of the slab as it bends in the MTZ. Seismicity in the transition zone in our study region is mostly clustered into two distinct regions, with a cluster of events occurring north of 12°S where we observed broadening of the slab anomaly in the MTZ and a second cluster occurring south of 26°S where the slab anomaly also broadens, although to a lesser extent than to the north. Work by Myhill (2013) indicates that such bands of denser deep seismicity could be associated with the localization of strain in the hinge zones of folds in the slab in the MTZ. If, as previously discussed, these zones of thickening of the slab in the MTZ are related to buckling or folding of the slab in response to increased resistance to subduction resulting from the viscosity increase across the upper-lower mantle boundary, then the distribution of the associated deep seismicity could similarly be affected by the deformation of the slab in the MTZ. An isolated region of very deep seismicity (greater than 625 km depth) in the MTZ corresponds to the location of the 9 June 1994 Mw 8.3 deep Bolivia earthquake which occurred at ~640 km
depth at ~14°S (International Seismological Centre, 2012; Fig. 6). This event was the largest deep-focus earthquake recorded until the more recent 2013 Mw 8.3 Okhotsk earthquake. Slip occurred along a subhorizontal plane (e.g., Kikuchi and Kanamori, 1994; Beck et al., 1995) perpendicular to the observed strike of the slab anomaly in our tomograms. Prior to the deep Bolivia earthquake limited deep-focus seismicity was observed in this region between the two clusters of deep seismicity to the north and south which were discussed previously. The focal mechanism of the 1994 Bolivia earthquake deviates from the more consistent pattern of downdip compression observed in the clusters of deep seismicity to the north and south. Detailed analyses of the fault rupture pattern of this earthquake have indicated that the 1994 Bolivia earthquake was probably caused by a combination of factors (e.g., Ihmlé, 1998; Zhan et al., 2014). Zhan et al. (2014) concluded that the initial rupture occurred in the cold core of the slab due to transformational faulting as metastable olivine underwent a phase change (e.g., Kirby et al., 1996). A large sub-event during the first phase of rupture triggered shear melting allowing the rupture to continue to propagate away from the core of the slab into the warmer slab material (Ogawa, 1987). The alignment of the E-W nodal plane with the strike of the slab anomaly in our results (Fig. 6) agrees with the idea that most of the slip was perpendicular to the strike of the slab as the rupture propagated away from the cold core of the slab into the warmer slab material where the melting point is more easily reached and slip increased due to positive feedbacks during shear melting.

The Nazca slab shows along-strike variations in morphology in the lower mantle (Figs 7 and 8). The high amplitude continuous slab anomaly north of 16°S corresponds to the steeply dipping (~70°) segment of the Nazca slab inboard of the Peruvian flat slab as
well as the thickened slab in the MTZ (Scire et al., 2015b). This corresponds to the region where Engdahl et al. (1995) observed ponding of high velocity material immediately below the MTZ. Where the dip of the Nazca slab is more gradual when entering the MTZ (Fig. 8), less widening of the slab anomaly is observed. The slab anomaly, particularly where less widening of it is observed, also appears to begin flattening to the east after it enters the lower mantle although this could be influenced by lateral smearing on the edges of our model. Since several global tomography studies image the diffuse fast anomaly in the lower mantle associated with the Nazca slab broadening to the east (e.g. Bijwaard et al., 1998; Fukao et al., 2001; Zhao 2004; Li et al., 2008), as does Engdahl et al. (1995), it is possible this apparent flattening is not solely a result of lateral smearing. Rocha et al. (2011) image a fast anomaly that they interpret as the subducting slab at 900 km depth at the edge of their study region under Brazil at 57° to 58°W, which also indicates that at least some of the eastward flattening that we observe is genuine. The along-strike segmentation of the slab in the lower mantle south of the Bolivian orocline also appears to correspond to the more gradual dip angle of the slab as it enters the MTZ. In the north where the incoming dip angle is high and we observed extensive deformation in the MTZ, the slab anomaly remains coherent into the lower mantle unlike further south. This potentially indicates that slab dip angle and the subsequent deformation in the MTZ are influencing the strength and cohesion of the slab in the lower mantle.

_A hydrated mantle transition zone_

Hydration of the mantle transition zone by the transportation of hydrous phases to depth by subducting slabs has been suggested by a number of studies (e.g., Bercovici and
Karato, 2003; Smyth and Jacobsen, 2006; Ohtani et al., 2004; Magni et al., 2014). Direct evidence for hydration of MTZ minerals has been observed recently in a sample of hydrated ringwoodite discovered as an inclusion in a diamond (Pearson et al., 2014). Outer rise normal faulting is thought to allow for hydration of the oceanic mantle lithosphere (e.g., Peacock, 2001; Garth and Rietbrock, 2014), although the degree of mantle lithospheric hydration is uncertain. The high amplitude low velocity anomaly observed in our tomograms in the MTZ to the north of 18°S could be associated with dehydration of the oceanic mantle lithosphere in the MTZ or just below the 660 km discontinuity in the lower mantle, leading to a region of locally elevated hydration in the MTZ or possibly dehydration melting related to the downward flow of hydrous minerals due to the penetration of the Nazca slab into the lower mantle. Schmandt et al. (2014) theorized that downward flow into the lower mantle would result in dehydration melting due to the low water storage capacity of lower mantle minerals (e.g., Bolfan-Casanova et al., 2000). Since the hydrous melt is likely to be slightly less dense than the top of the lower mantle and therefore would rise upwards back into the MTZ (Sakamaki et al., 2006), the presence of this melt in the lower mantle is likely short-lived. Since our low velocity anomaly is confined to the MTZ, we suggest that the low velocity anomaly could be due to the presence of a hydrous melt which has returned to the MTZ after dehydration melting has occurred in the lower mantle. In addition, dehydration melting has been suggested to occur when hydrous wadsleyite moves upwards across the 410-km discontinuity into the olivine stability field in the upper mantle (Bercovici and Karato, 2003). Both Schmerr and Garnero (2007) and Contenti et al. (2012) have observed a disrupted 410-km discontinuity under the central Andes, which both have attributed to
compositional heterogeneity in the upper part of the MTZ in this area. Schmerr and Garnero (2007) suggest that their observation of a deepened 410 km discontinuity could be due to the presence of a hydrated “lens” of wadsleyite while Contenti et al. (2012) suggest that the decreased reflectivity of the 410 km discontinuity could be either due to the effects of hydration of the zone directly above MTZ or the presence of a layer of neutrally buoyant melt. While some studies suggest that this hydrous melt would be neutrally buoyant at the 410 km discontinuity (e.g., Sakamaki et al., 2006; Bercovici and Karato, 2003), Hirschmann et al. (2006) argues that the density contrast between the hydrous melt and solid peridotites is such that the hydrous partial melts could potentially percolate up rather than pooling at 410 km depth. Continued upward flow of hydrous melts from the MTZ could explain the apparent continuation of the low velocity anomaly into the upper mantle (Fig. 8, cross sections D, E, and F) although the effect of this hydrous melt on the upper mantle and potentially even the cratonic lithosphere under which it is rising is unknown.

CONCLUSION

By combining data from fourteen separate networks which were deployed in the central Andes we are able to image the Nazca-South American subduction zone to depths of ~900 km between 6° and 32° S. Our images of the Nazca slab in the MTZ and lower mantle provide new information about the deformation of the slab and its interactions with the boundary between the upper and lower mantle. We observe significant thickening of the slab in the MTZ in the northern part of our study region (north of 16°S) which is due to temporary stagnation and possibly folding or buckling of the slab in the
MTZ in response to the decreasing of the sinking velocity of the slab due to increasing viscosity in the lower mantle. Limited thickening of the slab is observed to the south where the Nazca slab appears to subduct into the lower mantle with little to no deformation in the MTZ. While the Nazca slab penetrates into the lower mantle through the entirety of our study area, significant along-strike variation in the amount of deformation both in and below the mantle transition zone is observed. The slab anomaly changes from being a broad, higher amplitude cohesive anomaly in the north to a more segmented slab anomaly in the south. This variation appears to partially correspond to changes in the dip of the slab where it enters the MTZ. We observe a low velocity zone above the slab in the MTZ to the north of 18°S, which we attribute to the effects of hydration of the MTZ. Our images of the subducting Nazca slab could potentially provide new constraints on possible slab morphologies which could help guide interpretations of future modeling studies exploring the interactions between the slab and the MTZ. Slab morphology and deformation in the MTZ and lower mantle varies strongly in the same subduction zone, indicating that along-strike variations in the plate boundary have a significant effect on the deformation and fate of the slab at depth.

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REFERENCES


Scire, A., Biryol, C., Zandt, G. and Beck, S., 2015a. Imaging the Nazca slab and surrounding mantle to 700 km depth beneath the Central Andes (18° to 28°S), in *Geodynamics of a Cordilleran Orogenic System: The Central Andes of Argentina*


Figure 1. Map showing seismic station locations for individual networks used in the study and topography of the central Andes. Slab contours (gray) are from the Slab1.0 global subduction zone model (Hayes et al. 2012). Earthquake data for deep earthquakes (depth > 375 km) are from 1973 to 2012 (magnitude > 4.0); U.S. Geological Survey – National Earthquake Information Center (NEIC) catalog. Red triangles mark location of Holocene...
volcanoes (Siebert and Simkin, 2002). Plate motion vector from Somoza and Ghidella (2012). Cross sections lines (yellow) are shown for cross sections in Fig. 8.

Figure 2. A) Global map centered on our study region (black square) showing the location of events used in this study. Darker gray circles mark events used for direct P arrivals while open circles mark events used for PKIKP arrivals. B) Plot showing the azimuthal distribution for all rays used in the study. Ray distribution is strongly controlled by the location of plate boundaries where the earthquakes are generated.
Figure 3. Normalized hit quality plots for P-wave tomography model. Hit quality for a node is based on the number and azimuthal distribution of rays that sample that node. Good hit quality is indicated by red shading while poor hit quality is indicated by blue shading. Comparisons with hit quality maps from Scire et al. (2015a) (center) and Scire et al. (2015b) (right) indicate that sampling in the MTZ is substantially improved by combining the two datasets. Hit quality maps for additional depth slices for this study are in Supplemental Material, Fig. S2.
Figure 4. Horizontal depth slices for the checkerboard tests for a selection of model layers. Input for neutral layers (0% velocity deviation) is not shown. Output for neutral layers (505 and 660 km depth) is shown in the left column. Resolution of velocity anomalies in neutral layers shown here indicates that some vertical smearing is occurring. On the right is the input and output for the layers with the checkerboard anomalies. The checkerboard tests show that while amplitude recovery remains good in the center of the model space, lateral smearing increases towards the edge of the model, particularly in the lower mantle (e.g., 835 km depth). Checkerboard tests for additional model layers are in Supplemental Material, Fig. S3.
Figure 5. Depth slices for our synthetic slab recovery tests. The geometry of our input slab model is based on the Slab1.0 contours (Hayes et al., 2012). Synthetic input (left) and output (right) are shown. Lateral smearing of the recovered slab anomaly increases with depth while recovered amplitude decreases slightly.
Figure 6. Horizontal depth slices for the mantle transition zone (410, 455, 505, 555, 605, and 665 km depth) from the tomography model. The short dashed line represents the edge of the well-resolved region of the model, defined as regions with hit quality greater
than 0.2 (Biryol et al., 2011). Stars mark locations of seismic stations used in the study. Red triangles mark location of Holocene volcanoes (Siebert and Simkin, 2002). Yellow dots are earthquake locations from the EHB catalog (Engdahl et al., 1998). Solid black lines are slab contours from Slab1.0 model (Hayes et al., 2012). Focal mechanisms are from the Global Centroid Moment Tensor database (Dziewonski et al., 1981; Ekström et al., 2012).

Figure 7. Horizontal depth slices for the lower mantle (715, 775, 835, and 895 km depth) from the tomography model. Dashed lines are the same as in Fig. 5. Stars mark locations of seismic stations used in the study. Red triangles mark location of Holocene volcanoes (Siebert and Simkin, 2002). Solid black lines are slab contours from Slab1.0 model (Hayes et al., 2012).
Figure 8. Trench perpendicular cross sections through the tomography model. Cross section locations are as shown in Figure 1. Dashed lines are the same as in Figure 5. Yellow dots are earthquake locations from the EHB catalog (Engdahl et al., 1998). Solid black line marks the top of the Nazca slab from the Slab1.0 model (Hayes et al., 2012).
Figure S1. Plot of tradeoff analysis for the P-wave inversion between the variance reduction and the Euclidean model norm (L2) performed to choose preferred overall damping (D1-D10) and smoothing (S1 –S10) weights. The black star shows the damping (D5) and smoothing (S4) parameters used in this study.
Figure S2. Normalized hit quality plots for P-wave tomography model. Hit quality for a node is based on the number and azimuthal distribution of rays that sample that node. Good hit quality is indicated by red shading while poor hit quality is indicated by blue shading.
Figure S3. Horizontal depth slices for the checkerboard tests for a selection of model layers. Input for neutral layers (0% velocity deviation) is not shown. Output for neutral layers (455 and 715 km depth) is shown in the left column. Resolution of velocity anomalies in neutral layers shown here indicates that some vertical smearing is occurring. On the right is the input and output for the layers with the checkerboard anomalies. The checkerboard tests show that while amplitude recovery remains good in the center of the model space, lateral smearing increases towards the edge of the model, particularly in the lower mantle.
Figure S4. Horizontal depth slices for the upper mantle (95 to 365 km depth) from the tomography model. Dashed lines are the same as in Fig. 5. Stars mark locations of seismic stations used in the study. Red triangles mark location of Holocene volcanoes (Siebert and Simkin, 2002). Solid black lines are slab contours from Slab1.0 model (Hayes et al., 2012).