The transition zone beneath West Argentina-Central Chile using P-to-S converted waves

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Key Points:

- We observe depressed 410 with sharp velocity jump and frequency-dependent topography of the 660 with a broad velocity gradient.
- Attributes of the transition zone discontinuities and its velocity structure indicate near pyrolitic mantle at transition zone depths.
- Presence of a negative amplitude precursory signal to P660s (P590s) is explained by basalt accumulation at the base of the transition zone.

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Abstract

We investigate the mantle transition zone beneath the Chile-Argentina flat-subduction region by means of P-to-S conversions at mantle discontinuities from teleseismic events recorded at 103 seismic stations. From the analysis of receiver functions, we obtain clear converted phases from the 410 and 660 discontinuities and we identify a robust precursory signal to $P660s$, of negative amplitude, that we name $P590s$. We observe little frequency dependence in the amplitude of the $P410s$ converted phase; while the $P660s$ is less visible towards higher frequencies. The 410 is on average deeper than 410 km **by (10 ± 1) km in the higher frequency bands**, and **it is** relatively sharp, being consistent with a 10% velocity jump over less than 20 km. The observed 660 depth varies with frequency; it is deeper by up to **(18 ± 2) km** for lower frequencies and close to reference at higher frequencies, being consistent with a 13% broad velocity gradient over **30-40** km, probably caused by a composite of multiple phase transitions. The transition zone thickness is controlled by the frequency-dependent depth variability of the 660. Our findings of relative depth, width and velocity jump of the detected discontinuities, combined with tomographic images of the mantle transition zone cannot be explained by thermal variations alone. Compositional constraints from mineral physics show that a near pyrolitic mantle is consistent with the ratio of the estimated velocity jumps. However, the negative $P590s$ phase in this region could be signal from the velocity reduction due to basalt accumulation at the base of the transition zone.

1 Introduction

The upper-mantle has a layered structure characterized by a number of seismic discontinuities which correspond to sharp jumps in seismic wave speed. Their properties (discontinuity depths, topography, the sharpness and the size of the velocity and density jumps) are of importance, not only in the interest of improving seismic models of Earth's structure, but because their underlying mineral phase transitions give direct insight into the thermal, dynamic and chemical state of the mantle. The transition zone (TZ) separates the upper mantle from the lower mantle and is bounded by two discontinuities lying at depths around 410 and 660 km, which correspond to increases in seismic wave speed of 3.4% v_S and 2.6% v_P at 410 km and of 6.7% v_S and 4.7% v_P at 660 km in *prem* (Dziewonski & Anderson, 1981). These are global features and we will refer to them as the 410 and 660.

The seismic velocity profile of the upper mantle and the transition zone can be explained to the first-order by a mantle of homogeneous pyrolitic composition (chemically equilibrated assemblage of olivine, pyroxene, and garnet with the bulk proportions of 4:2:1, (Ringwood, 1991)). The seismic discontinuities at the TZ are primarily related to isochemical phase transformations in the olivine system (Mg_2SiO_4 -system). At a depth of 410 km the olivine transforms to wadsleyite (Katsura & Ito, 1989), at 520 km the wadsleyite transforms to ringwoodite and at 660 km ringwoodite dissociates into bridgmanite ((Mg, Fe)-perovskite) and magnesiowustite, also known as the post-spinel transition (Ringwood, 1970; Ito & Takahashi, 1989). This picture is supported by mineral physics experimental data, predicting that both 410 and 660 phase transitions are expected to occur in sharp depth intervals (< 20 km) at pressures that correspond to the depth of these discontinuities (Frost, 2003; Litasov & Ohtani, 2005). Such sharp discontinuities are also consistent with many seismic observations which predict a prevailing velocity increase occurring over a depth range of 10-20 km (Paulssen, 1988; Benz & Vidale, 1993; Vidale et al., 1995; J. Collier & Helffrich, 1997; Landes et al., 2006; Schmandt, 2012).

Temperature anomalies in the mantle move the phase changes to different pressures (depths) according to their respective Clapeyron slopes (Bina & Helffrich, 1994). Due to the opposite sign in the Clapeyron slopes of the olivine-related phase changes responsible for the 410 and 660, the depth changes of these discontinuities are anti-correlated

in response to a thermal anomaly. While the 410 becomes shallower in colder regions and deeper in hotter ones, the depth changes of the 660 are opposite. As a consequence, in an olivine-dominant composition, the TZ becomes thicker near subducted slabs and thinner beneath high-temperature regions (J. D. Collier & Helffrich, 2001; Helffrich, 2000; Lawrence & Shearer, 2006; Vidale & Benz, 1992). The presence of other transforming components and variation in composition may introduce further complexities.

The mineralogical models most commonly used for comparisons to seismic observations assume the mantle as a chemical equilibrium assemblage (EA) of major elements, consisting of either a homogeneous pyrolitic composition (Ringwood, 1966; Ringwood, 1991) or a chemical stratification with homogeneous and equilibrated composition in each layer (e.g., Anderson & Bass, 1986). Yet EA models are incompatible with the experimentally measured small chemical diffusivity of mantle materials in the solid-state, which implies that over the age of Earth the mantle could hardly become equilibrated (e.g., Xu et al., 2008). Therefore, as opposite end-member, the mechanically mixed (MM) mantle model is considered, in which basalt and its depleted residual harzburgite - the two fractions of differentiated lithosphere introduced at subduction zones- are stirred by convection but do not equilibrate over geological timescales (Allègre & Turcotte, 1986). Geodynamic models suggest MM composition are important to consider for much of the mantle, and degrees of segregation are expected with basalt accumulating at the bottom of the mantle (Brandenburg & van Keken, 2007) and potentially also at the bottom of the TZ (e.g., Ballmer et al., 2015). Xu et al. (2008) compute the seismic velocities of EA and MM mantle according to a self-consistent thermodynamic model developed by Stixrude and Lithgow-Bertelloni (2005, 2011) demonstrating that, for identical bulk composition, EA and MM have different mineral proportions and therefore different seismic properties and seismological attributes of the discontinuities (Xu et al., 2008). Seismic velocities and density for the two cases differ mainly at the bottom of the TZ, where they exhibit a complex structure. For EA models the velocity profiles show high velocity gradients below 660 km -with increasing width as the basalt fraction increases- due to the gradual transformation of garnet to bridgmanite. This feature is not present in the MM models, which rather show two sharp features, a large velocity jump at 660 km (due to the ringwoodite dissociating) and a small one at 745 km (due to the garnet dissociating). The seismic structures around the 660 vary even more when considering temperature changes in the EA and MM models (Xu et al., 2008). In the EA model the 410 shallows and reduces in amplitude with increasing basalt, as is also seen in experiments (Katsura et al., 2004). In the MM model the 410 does not change depth, but only weakens in amplitude with increasing basalt. Basalt enrichment in both EA and MM models increases seismic velocities above the 410 and decreases them within the mantle transition zone. In the transition zone, when compared to radial seismological models, the MM provides a better match to the seismic velocity profiles than the EA end-member, because the former is faster and has a steeper radial velocity gradient (Xu et al., 2008). In addition, if a MM model were representative of the mantle seismic velocities, these would be primarily sensitive to temperature because the predicted MM velocity profiles are nearly independent of bulk composition. Careful analyses of the depth, width and velocity jump(s) at each discontinuity can untangle the effects of composition and temperature, which are distinct for EA and MM end-member.

The main objective of this work is investigating the seismic properties of the mantle transition zone discontinuities at the regional scale, beneath an area that includes the Chile-Argentina flat subduction, exploiting data collected from past temporary seismic deployments. In particular, we estimate depth, magnitude and width of velocity jumps at TZ depths that help constrain the characteristics of the transition zone in terms of thermal state and/or mineralogical composition. To our knowledge, this is the first study of such a kind performed in this region. We particularly explore the frequency-dependency of the observations, to assess the width of the discontinuities as an additional constraint.

Figure 1. Tectonic setting of the study area (marked with a rectangle), showing volcanoes since Holocene with gray triangles and seismicity with coloured circles. Contours (black) of the Nazca slab from the Slab1.0 model of Hayes et al. (2012) and predicted path of the Juan Fernández hotspot chain (white dashed) at present from Yañez et al. (2001).

2 Study Area

The study region (Fig. 1) is located at the western margin of South America (between 24°S and 39°S) and includes an area known as the Chile-Argentina flat subduction zone (between 28°S and 32°S). Unlike what happens in most subduction zones, in this region the Nazca Plate subducts beneath the South American Plate with an angle close to the horizontal in the shallow-most mantle. The flat-slab geometry correlates with the inland prolongation of the subducting Juan Fernández Ridge (JFR in Fig. 1) (Yañez et al., 2001) and a hotspot seamount chain on the Nazca Plate. The region is characterized by the absence of uplift and of volcanism since the Miocene. Towards the South, the Nazca Plate has a normal subduction with a dip of about 30°. The change in subduction angle is evident in the contours of the Wadati-Benioff zone of the Nazca Plate (black contours in Fig. 1 from Hayes et al. (2012)).

Beneath this region, the tomographic images from global-scale models show high velocity anomalies interpreted as subducted lithosphere sinking down into the mantle, either entering the TZ and spreading horizontally as a stagnant slab (Fukao & Obayashi, 2013) or penetrating into the lower mantle (Li et al., 2008) at the eastern edge of our study area.

Different seismic experiments have been conducted in this area mainly to constrain the lithospheric structure, which seems to play an important role in the mechanisms that explain flat subduction (Gutscher et al., 2000). The exact origin of the flat subduction is still under debate, but **one** potential model shows that a moderately overthickened crust provides the additional buoyancy for the Nazca slab to remain flat (Gutscher et al., 2000; Gutscher, 2002). In the flat-slab region and towards the South, the analyses of the shallow structure using P-wave receiver functions (Gilbert et al., 2006; Gans et al., 2011), S-wave receiver functions (Heit et al., 2008), Pn apparent phase velocities (Fromm et al., 2004), and gravity modelling (Yañez et al., 2001) show a consistent presence of

Figure 2. a) Distribution of the 103 broad seismic stations across four networks (see legend) used in this study. b) Worldwide distribution of the 197 teleseismic events used in the analysis and piercing point distribution for P410s (orange crosses) and P660s (gray crosses) for the 1342 RFs that has passed the quality controls at $f_2 = 0.08$ Hz. Black lines show slab contours (Hayes et al., 2012).

subduction of overthickened oceanic crust of about 13-20 km thick. Recent increase of seismic data have led to high resolution relative-time teleseismic tomography (Portner et al., 2017) and shear-wave splitting (Lynner et al., 2017) studies, which find evidence that the structure of the flat Nazca slab in this region is quite heterogeneous at upper mantle depths, suggesting in particular that there is a hole or tear in the slab.

The complex tectonic setting has motivated a wealth of seismic experiments in this region. The current data coverage makes it an excellent laboratory to explore and quantify the relative contributions of thermal and compositional heterogeneities to the mantle transition zone discontinuity structure, which is still rather unexplored.

3 Data and Method

3.1 Data Selection

In this study, we combine data available from 103 stations (Fig. 2 a) across four two-years experiments: 18 stations from CHARGE (Beck et al., 2000, network code YC); 43 from SIEMBRA (Beck & Zandt, 2007, network code ZL), 30 from PUDEL (Heit et al., 2007, network code 2B) and 12 from ESP (Gilbert, 2008, network code XH). Other stations from these experiments were not used in this analysis either due to lack of data access or to poor data quality. Through the IRIS Data Management Center, we access the waveforms of 964 events of Mw between 5.5 and 7, and from epicentral distances between 30° and 95° . We immediately exclude data with a signal-to-noise ratio of the P phase on the vertical component (filtered between 0.02-0.08 Hz) of less than 1.4. This results in a data set of 9200 event-station pairs.

3.2 Teleseismic Receiver Functions and Processing

We use teleseismic Pds phases, i.e. the phases resulting from the direct P wave converting to an S wave at seismic discontinuity at a depth (denoted by d in Pds). These

Figure 3. Example ray-path for a P -to- S conversion at the 410 and 660. The right panel is an enlargement of the box beneath the station in the left panel.

phases are shown in Fig. 3 for the 410 and 660 discontinuities. Because of their polarization and their almost vertical incidence, Pds conversions are recorded on the radial (R) component of teleseismic earthquakes. They arrive in the coda of the P phase along with multiple reflections and scattered waves. The abundance of other phases and the weak amplitude of the Pds phases make direct detection in individual seismograms difficult. In addition, Pds phases are expected to be coherent with the waveform of the main P arrival for conversion at discontinuities which are thinner than one half of the P -wavelength (Richards, 1972; Paulssen, 1988; Bostock, 1999). In order to extract them by the waveform similarity, we use deconvolution of the P phase in the vertical component (P_Z) from the P -wave coda in the radial (P_R), which is known as the receiver function technique (RF) (Phinney, 1964; Vinnik, 1977; Langston, 1979; Ammon, 1991). Since the teleseismic P phase approximates the source, the above deconvolution eliminates the source component and isolates the P -to- S conversion peaks from discontinuities beneath the station. Moreover, the differential travel times between P and Pds provide a measure of the one-way S travel time between the surface and the discontinuity. Thus, using an adequate velocity model, relative travel times can be translated to discontinuity depths.

Deconvolution of P_Z from P_R is equivalent to spectral division in the frequency domain:

$$\frac{P_R(\omega)\bar{P}_R(\omega)}{P_Z(\omega)\bar{P}_Z(\omega)} \quad (1)$$

where the upper bar denotes the complex conjugate. To stabilize the spectral division, we use the water-level deconvolution (Clayton & Wiggins, 1976). We apply the deconvolution using five different frequency bands, all with lower cut-off frequency at $f_1 = 0.02$ Hz and higher cut-off frequency at $f_2 = 0.08, 0.12, 0.2, 0.32$ and 0.64 Hz, respectively. A detailed description of the processing for computing the RFs, and the automatic and visual quality control are given in the Text S1. After quality control, we retain 1342 RFs for $f_2 = 0.08$ Hz, 862 RFs for $f_2 = 0.12$ Hz, and 872 RFs for $f_2 = 0.2, 0.32, 0.64$ Hz. Fig. 2b shows the piercing point distribution for P410s and P660s for the RFs that have passed the quality controls at $f_2 = 0.08$ Hz.

3.3 Detection of P -to- S Conversions in Stacked RFs

For signal detection, we use the phase weighted stack (PWS) (Schimmel & Paulssen, 1997) and the linear stack (LS). The PWS is a non-linear stack which uses the coher-

ence of the instantaneous phase as a weight in the linear stack to enhance the signal-to-noise ratio (SNR). The stacking is performed with a linear move-out correction:

$$S(s_j, t) = \frac{1}{N} \sum_{i=1}^N RF_i(t + s_j(\Delta_i - \Delta_{REF}))c_i \quad (2)$$

$$c_i = \left| \frac{1}{N} \sum_{k=1}^N e^{i\phi_k(t + s_j(\Delta_i - \Delta_{REF}))} \right|^\nu \quad (3)$$

where c_i is the phase stack, s_j refers to a specific relative slowness with respect to the P phase, t is relative time with respect to the P phase, Δ_i is the epicentral distances for RF_i , Δ_{REF} is the reference epicentral distance, ϕ_k is the instantaneous phase for the k -th RF , N is the total number of stacked RF s and the parameter ν controls the weighting of phase similarity and dissimilarity. The LS is defined with $\nu = 0$ ($c_i = 1$), while for the PWS we choose a value of $\nu = 2$. The reference distance, Δ_{REF} , of 50° is chosen for all the slant stacks to account for the distribution of epicentral distances which are mainly concentrated between 35° and 65° .

Signal identification and detection using stacks of RF s is performed in three steps: (1) RF s are stacked in the relative time-slowness domain ($\Delta t, \Delta s$), as shown in Fig. S2 a. The slowness value (Δs_{obs}) for each coherently stacked signal is saved to be used in the next step. (2) The depth stacks (PWS and LS) with linear move-out correction are made using the slowness value obtained for each phase in step 1. The resulting trace for the PWS is shown in Fig. S2 b. (3) To test the robustness of potential signals, the 95 per cent confidence criteria is used. A maximum in the stacked RF obtained in step 2 is considered a signal detection if the mean stacked amplitude is larger than twice the standard error at that time. A more detailed description of signal detection and identification can be found in Text S2.

3.4 Common Conversion Point Stacking

The averaged properties of the discontinuities in the area are obtained by stacking all receiver functions which passed the quality control (see 4.1) and lateral variations of the detected discontinuities are investigated through a common conversion point (CCP) stacking approach. To apply the CCP stack, we first group the RF s into bins of common conversion point areas for two different depth: 410 km and 660 km -which correspond to the global depth of our target discontinuities. The piercing point density is variable over the entire study region. This implies a variable discontinuity visibility, which we account for using adaptive bin sizes depending on the local number of piercing points. The bin radii that we choose are on the order of the Fresnel radius of $P410s$ and $P660s$ at the highest frequency considered in this study ($r_{Fresnel}^{P410s} \sim 0.7^\circ$ and $r_{Fresnel}^{P660s} \sim 0.8^\circ$ at 0.2 Hz). The bins are defined as circular caps of radius $r_1 = 0.5^\circ$ equally spaced every 0.7° . If the number of RF s in the bin is less than 20, we increase the bin radius to $r_2 = 1.5r_1$, or to $r_3 = 2r_1$ if numbers remain low. The variable bin sizes and their fixed distances allow overlapping of RF s between adjacent stacks. Bins at r_3 with still less than 10 RF s are not considered in the analysis; this results in about 160 stacks for $P410s$ and 180 for $P660s$ across the study area. Fig. S3 shows an example of bin size distribution (top panel) and the corresponding number of RF s inside each bin (bottom panel) for each phase and for $f_2 = 0.08$ Hz. Small differences in the data coverage are observed for the higher frequency bands.

For each bin and conversion depth, we first slant stack the RF s and perform a visual quality control on individual stacks (see Text S3). We estimate relative slowness from the stacks that pass the quality check to compute the stacking with move-out correction

using the correct slowness at which each phase is detected. Finally, we measure relative travel time values and standard errors from the PWS at all frequencies.

3.5 Depth Conversion

To constrain absolute depth of 410 and 660, we need regional absolute v_P and v_S models for the study area down to the base of the TZ. Regional scale relative velocity models, like the model in Portner et al. (2017), are difficult to use as the reference model and the method can lead to artefacts (Bastow, 2012). Different 3D global tomographic models are available consistently showing average high P- and S-wave velocity anomalies in the upper-mantle and transition zone below the study area (e.g., Li et al., 2008; Kustowski et al., 2008; Ritsema et al., 2011; Obayashi et al., 2013).

We chose to compute the average 1D P-wave velocity profiles for each CCP bin by modifying *ak135* reference model (Kennett, 1991) to include the P-velocity anomalies, δv_P , of the tomographic model MIT08 (Li et al., 2008). The authors, by means of noiseless synthetic tests, estimate generally less than 70% amplitude recovery in the upper mantle for their tomographic model and an overall variable resolution, partly due to the damping applied in the inversion procedure to suppress the effects of noise in the data (Li et al., 2008). In order to account for possible resolution overestimation given by noiseless synthetic tests, we assume that the amplitudes of anomalies recovered by the inversion could be about 50% than those of unknown real anomalies. Therefore, we compute the corrections using MIT08 anomalies multiplied by a factor of 2. This factor choice, while being arbitrary to a certain extent, likely provides an upper bound for time corrections and puts the average P660s depth in our CCP stack at around 660 km. S-wave velocity profiles are derived from the δv_p values employing a constant factor $\delta v_S/\delta v_P = 1.5$; a factor typically ranging between 1.5 and 2 in the upper mantle (Ritsema & Van Heijst, 2002).

We use TauP tool-kit (Crotwell et al., 1999) to create the modified velocity models for each bin. The corresponding time corrections are computed with respect to *ak135* and are shown in Fig. S5. For our dataset of detected phases, these corrections range between -1 s to 1.9 s for *P410s* and from -1.4 s to 2.8 s for *P660s*. If we did not modify MITP08 anomalies, these values would have range between -0.5 s and 1 s for *P410s* and from -0.7 s and 1.4 s for *P660s*. Finally, the corrected time values are converted to depth using the reference velocity model *ak135* for a reference distance of 50°.

3.6 Thickness of Discontinuities

Seismically, the discontinuities are often approximated as linear velocity gradients, which act as low-pass filters to converted waves, attenuating the high frequency energy. The high corner frequency, f , of these filters is related to the thickness, Δz , of the velocity gradient (Bostock, 1999; Paulssen, 1988) through the following inequality:

$$\Delta z < \frac{\lambda_P}{2} = \frac{v_P}{2f} \quad (4)$$

where λ_P is the wavelength of the incident *P*-wave and v_P is the *P*-wave velocity at the corresponding discontinuity depth. This inequality (4) indicates that *Pds* phases are sensitive to linear velocity gradients occurring in a depth interval thinner than half the wavelength of the incident *P*-wave.

Therefore, to constrain the discontinuity thickness, one needs to find the highest frequency f^{max} for which significant *P*-wave energy is converted. This rule of thumb has been used before to constrain the thickness of mantle discontinuities (e.g., Paulssen, 1988; van der Meijde et al., 2003; Jenkins et al., 2017). We estimate this frequency value from the LS of the entire dataset by analysing the resulting amplitudes of converted phases

across the different frequency bands that we use to compute the RFs. For comparison, we compute synthetic seismograms for velocity gradients of different thickness using reflectivity synthetics (Fuchs & Müller, 1971) and compute RFs and stacks in different frequency bands similar to the real dataset.

Table 1. Estimated slowness values.¹

f_2 (Hz)	Δs_{P410s} (s/deg)	Δt_{P410s} (s)	Δs_{P590s} (s/deg)	Δt_{P590s} (s)	Δs_{P660s} (s/deg)	Δt_{P660s} (s)
0.08	-0.08	46.6	-0.26	63.9	-0.16	71.8
0.12	-0.12	46.1	-0.14	64.9	-0.18	70.4
0.2	-0.1	46.2	-0.1	63.5	-0.18	70.7
0.32	-0.16	46.1	-0.08	62.8	-0.14	70.4
0.64	-0.16	46.0	-0.12	62.7	-0.14	70.7

¹ Extracted from the PWS in the relative time slowness domain (Fig. 4) and corresponding relative travel time.

4 Results

4.1 Stacking the Entire Dataset

We first obtain the average properties of the discontinuities in the area by stacking all receiver functions for different frequency bands (see Fig. 4). Signals are clearer in the PWSs than in the LSs. Nevertheless, both types of stacks show coherently stacked signals at relative travel times and slownesses which coincide with the theoretical values in *ak135* for the converted phases *P410s* (45.2 s, -0.1 s/°) and *P660s* (70.4 s, -0.21 s/°). The slowness values observed for each frequency band are summarized in Table 1. At the lower frequencies ($f_2 = 0.08, 0.12, 0.2$ Hz) a coherently stacked signal is observed arriving before *P660s*, it is detected with a negative slowness of -0.26 s/° at about 64 s (detection at $f_2 = 0.08$ Hz) and a negative amplitude (i.e. opposite sign with respect to *P* phase). Its slowness suggests that this signal corresponds to a converted phase (a multiple would arrive with positive relative slowness) from a heterogeneity/discontinuity at a depth of about 590 km. Its polarity suggests that the conversion occurs when the *P* phase passes from a low velocity region to a higher velocity region; i.e., a velocity decrease with increasing depth. Hereafter, we will call this signal *P590s*.

We compute move-out corrected stacks from Eq. 2 using the slowness values in Table 1 for each coherently stacked signal of negative slowness (i.e. *P410s*, *P590s*, and *P660s*). These stacks are shown in Fig. 5 showing the positive *P410s* and *P660s*, and negative *P590s*. The amplitude of the LS and PWS are normalized in the time interval 30-85 s to allow for a direct visual comparison between both stacking techniques. Again, the figure shows that the LS is overall noisier than the PWS. This difference is even more remarkable for the highest frequencies, where the large-amplitude high frequency noise is less attenuated with the LS. Additionally, the standard error of the PWS is larger than that of the LS due to the larger sensitivity of the PWS to small variations within the dataset. The estimated relative travel times at all frequencies and their corresponding standard errors are summarized in Table 2.

The stacks of Fig. 4 also show a coherently stacked signal at (23.7 ± 0.1) s with zero relative slowness. This signal is only clearly observed at the lower frequency bands. A potential *P210s* and crustal/lithospheric multiples (*PSds*, *PPds*, *PSdp*, where *d* is the Moho or any other lithospheric discontinuity) are expected to arrive at similar relative

Figure 4. Stacks of the entire data set in the relative time-slowness domain using the linear stack (LS, left panel) and the phase weighted stack (PWS, right panel) for $\Delta_{REF}=50^\circ$. Each row corresponds to a different frequency band with lower corner frequency at 0.02 Hz and upper corner frequency, from top to bottom, at 0.08 Hz (1342 RFs), 0.12 Hz (862 RFs), 0.2 Hz (872 RFs), 0.32 Hz (872 RFs), 0.64 Hz (872 RFs). Black crosses correspond to theoretical relative slowness and travel time for P210s, P410s, P510s and P660s in a modified *ak135* velocity model to account for the '510' discontinuity.

travel times with slowness close to the slowness of the P phase. A more suitable approach to separate a potential $P210s$ from multiples would be using S receiver functions (e.g., Farra & Vinnik, 2000), which have the advantage of being free of S-wave multiple reflections (e.g., Yuan et al., 2006; Vinnik et al., 2009). However, this analysis is beyond the scope of this article.

Table 2. Estimated relative travel time values and relative amplitudes of P-to-S conversions.²

f_2 (Hz)	Δt_{P410s} (s)	$P410s/P$ (%)	Δt_{P590s} (s)	$P590s/P$ (%)	Δt_{P660s} (s)	$P660s/P$ (%)	TZT (s)
0.08	46.7 ± 0.1	6.9 ± 0.2	64.0 ± 0.1	6.3 ± 0.2	71.5 ± 0.2	7.3 ± 0.2	24.8 ± 0.2
0.12	46.1 ± 0.1	6.1 ± 0.3	64.9 ± 0.4	3.1 ± 0.2	70.5 ± 0.2	4.1 ± 0.3	24.4 ± 0.2
0.2	46.1 ± 0.1	4.4 ± 0.3	63.6 ± 0.2	2.6 ± 0.2	70.8 ± 0.2	3.2 ± 0.2	24.7 ± 0.2
0.32	46.1 ± 0.1	3.6 ± 0.3	62.8 ± 0.1	2.0 ± 0.2	70.4 ± 0.1	2.2 ± 0.2	24.3 ± 0.1
0.64	45.8 ± 0.1	2.4 ± 0.1	62.7 ± 0.1	1.4 ± 0.1	69.9 ± 0.1	1.3 ± 0.3	24.1 ± 0.1

² Relative travel times extracted from the PWS with move-out correction using the estimated slowness values in Table 1 (stacks in Fig. 5). Amplitudes of Pds phases extracted from the LS and expressed as a percentage of the amplitude of the P phase in the vertical component.

4.2 Detection of Converted Phases from Individual Bins

To investigate lateral variations of the detected discontinuities a CCP stack is applied for three frequency bands with $f_2 = 0.08, 0.12$, and 0.2 Hz. For these corner frequencies all the converted phases are clearly visible in the global stacks of the study area. The number of detections for each phase and frequency band are summarized in Table 3, taking only into account good quality stacks for which the converted signals are detected with negative slowness (including three detections of $P410s$ at $f_2 = 0.08$ Hz with zero slowness value). The distribution of detections at each frequency band for each phase is shown in Fig. 6 a to c. The $P410s$ phase shows the most stable mean travel time value across the different frequency bands, differing by less than 0.2 s. The remaining phases show mean values across the different frequency bands which differ as much as 3.1 s for $P590s$, and 1 s for $P660s$. The histograms of $P590s$ suggest that the arrival time of this phase is frequency dependent, larger relative times are obtained for larger values of f_2 .

4.3 Average Depth of 410 and 660

The absolute average depth of 410 and 660 are obtained from the stack of the entire dataset. The estimated depth values for each frequency band are summarized in Table 4. The table shows the average depth of 410 and 660 computed from a modified *ak135* velocity model that includes the velocity anomalies extracted from MIT08 (Li et al., 2008) multiplied by a factor of 2. We also include the average depth of 410 and 660 when considering the anomalies without amplification. This is done to show that the average depths do not change substantially for velocity anomalies amplified by factors between 1 and 2. True average depth of 410 and 660 in the study area are probably somewhere in between the two extremes. In what follows, we consider the observations using the upper limit of the amplified depth corrections.

In general, we observe that the 410 is deeper than the reference value at 410 km. The discontinuity has a maximum average depth of (428 ± 1) km at the lowest frequency band (or (425 ± 1) km depending on the velocity model used to compute time corrections) and it is observed more shallow (423 ± 1) km to (420 ± 1) km for the rest

Figure 5. Stacks of the entire data set using the estimated slowness values in Table 1 for phases *P410s* (left panel) and *P660s* (right panel). Gray continuous traces correspond to the LS and black continuous traces, to the PWS. Each row corresponds to a different frequency band with lower frequency at 0.02 Hz and upper frequency, from top to bottom, at 0.08 Hz (1342 RFs), 0.12 Hz (862 RFs), 0.2 Hz (872 RFs), 0.32 Hz (872 RFs), 0.64 Hz (872 RFs). The discontinuous traces correspond to the 95% confidence interval. The amplitudes of the stacks are normalized to one using the maximum amplitude in the time interval 30-85 s.

Table 3. Number of detections.³

	Q	0.08 Hz	0.12 Hz	0.2 Hz
<i>P410s</i>	1	36	27	31
	2	14	16	21
	total	50	43	52
<i>P590s</i>	1	28	9	9
	2	5	8	0
	total	33	17	9
<i>P660s</i>	1	45	19	22
	2	14	20	8
	total	59	39	30

³ Phases detected in the CCP stacks in the relative time and slowness domain. Q stands for quality of detection (1 or 2).

Table 4. Estimated averaged depths in five different frequency bands with lower corner frequency of 0.02 Hz.

f_2 (Hz)	410 (km)	660 (km) ⁴	410 (km)	660 (km) ⁵	<i>TZT</i> (km)
0.08	425 $\pm$ 1	675 $\pm$ 2	428 $\pm$ 1	678 $\pm$ 2	250 $\pm$ 2
0.12	420 $\pm$ 1	665 $\pm$ 2	423 $\pm$ 1	668 $\pm$ 2	245 $\pm$ 2
0.2	420 $\pm$ 1	668 $\pm$ 2	423 $\pm$ 1	671 $\pm$ 2	248 $\pm$ 2
0.32	420 $\pm$ 1	664 $\pm$ 1	423 $\pm$ 1	667 $\pm$ 1	244 $\pm$ 1
0.64	417 $\pm$ 1	659 $\pm$ 1	420 $\pm$ 1	662 $\pm$ 1	242 $\pm$ 1

⁴ Values obtained after time corrections of +0.25 s for *P410s* and +0.35 s for *P660s*, computed from a modified *ak135* velocity model that includes velocity anomalies extracted from MIT08 (Li et al., 2008).

⁵ Values obtained after time corrections of +0.5 s for *P410s* and +0.7 s for *P660s*, considering the anomalies in MIT08 multiplied by a factor of 2.

of the frequency bands. The depth difference between the lowest and highest frequency band for the 410 is (8 $\pm$ 1)km. The 660 depth is more variable across different frequency bands and it becomes shallower with increasing frequency. It is deeper than the reference value at 660 km for frequencies lower than 0.32 Hz and is at the reference depth for 0.64 Hz. The depth difference between the lowest and highest frequency band for the 660 is (16 $\pm$ 2)km.

4.4 Topography of 410 and 660

We construct the 410 and 660 topographic maps by applying the nearest-neighbour interpolation algorithm from the GMT plotting tool (Wessel & Smith, 1991) to the discontinuity depth values of CCP bins (see Fig. 7). This smoothing algorithm assigns to each node an average depth value which is computed as a weighted mean of the nearest point from each sector inside the search disk with a radius of 1°.

Figure 6. Number of detections of *Pds* phases. a) Histograms for the lowest frequency band (0.02-0.08 Hz). Vertical solid lines correspond to mean values of relative travel time for the detected phases of: 45.5 ± 1.5 s for *P410s*, 64.3 ± 1.1 s for *P590s*, 71.3 ± 1.3 s for *P660s*. The vertical red dashed lines correspond to theoretical values in *ak135* for $\Delta_{REF}=50^\circ$ (45.2 for *P410s* and 70.4 for *P660s*). b) Histograms for the middle frequency band (0.02-0.12 Hz) with mean values of relative travel time of: 45.7 ± 1.3 s for *P410s*, 65.7 ± 1 s for *P590s*, 70.2 ± 1.5 s for *P660s*. c) Histograms for the highest frequency band (0.02-0.2 Hz) with mean values of relative travel time of: 45.7 ± 0.9 s for *P410s*, 67.4 ± 0.4 s for *P590s*, 70.9 ± 0.9 s for *P660s*.

The 410 is consistently deeper everywhere for the three frequency bands. In general, for the same bin, depth values between frequency bands vary between 1 km and 20 km. We also observe some isolated bins with significantly large depth variation (of about 50-60 km) between frequency bands -e.g., the northernmost bins in the 410 maps near lat/long= $25^\circ\text{S}/68^\circ\text{W}$. We attribute these strong variations to poor data coverage in these regions potentially combined with small-scale heterogeneities that alter the arrival times of waves differently at each frequency band. The smoothest topography for the 410 is obtained for the higher frequency band ($f_2 = 0.2$ Hz) having the highest vertical resolution. For this frequency band, the 410 has a mean depth of 419 ± 2 km. At the lowest frequency bands mean depth values are similar: 417 ± 2 km for $f_2 = 0.08$ Hz and 419 ± 2 km for $f_2 = 0.12$ Hz. As expected, these mean values are in good agreement with the averages obtained by stacking the entire dataset.

For the 660 it is more difficult to identify a clear trend. The 660 maps in Fig. 7 shows less resemblance among the different frequency bands. This is mainly because there are less detections of *P660s* at higher frequencies. Bins with detections of *P660s* in at least two frequency bands show depth variations between 1 km and 25 km across the different frequency bands. There are some isolated bins with depth variations on the order of 30-40 km -e.g., lat/lon= $29^\circ\text{S}/71^\circ\text{W}$ and $27^\circ\text{S}/64.5^\circ\text{W}$. As observed for the global stacks for the study area, the mean depth value of 660 is variable across the different frequency bands: 676 ± 2 km for $f_2 = 0.08$ Hz, 665 ± 3 km for $f_2 = 0.12$ Hz and 670 ± 2 km for $f_2 = 0.2$ Hz.

4.5 Transition Zone Thickness

To minimize the errors in the 410 and 660 depth due to unaccounted upper-mantle and crustal structure, we compute the thickness of the TZ (TZT) by subtracting the 410

Figure 7. Absolute discontinuity depths for the 410 and 660 and transition zone thickness (TZZ) map. Results obtained from the analysis in three frequency bands: 0.02-0.08 Hz (top), 0.02-0.12 Hz (middle) and 0.02-0.2 Hz (bottom). Slab contours from (Hayes et al., 2012) are given for reference. The circles show the location of the CCP bins which are used in the interpolation -the position of each bin is obtained from averaging the latitude and longitude of the piercing points that fall within the bin.

Figure 8. Top: Relative PdS/P amplitude variation across several frequency bands (with upper corner frequencies at 0.08, 0.12, 0.2, 0.32 and 0.64 Hz) for signals detected in the global stack and identified as P-to-S conversions at depth of 410 km (black), 660 km (blue) and 590 km (red). Amplitude values are obtained from the LS of all the RFs that have passed the quality controls. Middle: Observed $P410s$ amplitude (solid line), amplitude without attenuation (dotted line) and amplitude corrected for topography effect (dashed line). Bottom: Observed $P660s$ amplitude (solid line), amplitude without attenuation (dotted line) and amplitude corrected for topography effect (dashed line). Error bars correspond to one standard error.

from the 660 depths ($h_{660} - h_{410}$). Differential travel times ($t_{P660s} - t_{P410s}$) are mainly sensitive to discontinuity topography and 3-D heterogeneities within the TZ because both phases, $P410s$ and $P660s$, are nearly identically affected by lateral variations in the Earth's structure above 410 km. The velocity anomalies within the TZ are expected and imaged to have smaller amplitudes than in the uppermost part of the mantle. Therefore, the analysis of the TZT yields a more reliable result than the individual 410 and 660 depths.

Table 4 summarizes the average values of TZT in all the frequency bands for the stack of the entire dataset, showing that the TZ becomes thinner for almost all frequencies higher than 0.08 Hz. This happens because the 660 becomes shallower for increasing frequency and the 410 is deeper for all frequencies. Therefore, the TZT is controlled by the depth variability of the 660 and it is not possible to determine a representative value of TZT for this area. The TZT map in Fig. 7 is dominated by TZT values smaller than the reference value of 250 km (red color dominates). The TZ is thinner beneath the flat slab by 241 ± 1 km and 240 ± 1 km for $f_2 = 0.08, 0.12$ Hz, respectively, and by 245 ± 1 km for $f_2 = 0.2$ Hz.

4.6 Frequency Dependent Visibility of TZ Discontinuities

We extract the relative amplitude values of each robustly detected signal ($P410s$, $P590s$, $P660s$) from the LS of the entire dataset for all the frequency bands. The amplitude values w.r.t to the main P phase are indicated in Table 2 and plotted against frequency in the top panel of Fig. 8. Phases $P590s$ and $P660s$ show an abrupt amplitude

drop of 50% and 57%, respectively, from 0.08 Hz to 0.12 Hz, while for the *P*410s the trend decreases very smoothly as the frequency increases.

Ideally, one would expect a constant amplitude of converted phases across any frequency band for sharp discontinuities. However, the presence of noise, misalignment of converted phases at higher frequencies (due to topography of the discontinuities or at the Earth's surface, the presence of scatterers, and the accuracy of 3D velocity corrections), attenuation and the assumption of constant slowness, would result in a smooth loss of amplitude at higher frequencies. We perform a synthetic test to quantify the effect that the topography of discontinuities have in the amplitudes of stacked RFs at different frequencies. We compute synthetic seismograms (for epicentral distances between 30 and 70 degrees) by changing the depth of the 410 and 660 with the depth values we obtain in the CCP stacks (in the frequency band 0.02-0.12 Hz). We correct the observed amplitudes for the suspected effects of observed topography to interpret the discontinuities. The idea is to obtain a set of synthetic seismograms with *P*410s and *P*660s phases arriving at slightly different times, as in the real dataset. We stack the synthetic RFs and compare the amplitudes of *P*410s and *P*660s against the corresponding amplitudes when we consider no topography on the depth of 410 and 660 in *prem*. We also estimate the frequency dependent effect that attenuation has on the amplitudes of *P*410s and *P*660s. Our synthetic seismograms are typically computed with a model with low Q_μ values in the upper mantle (Q structure of QL6 (Durek & Ekström, 1996)). Therefore, we compare the amplitudes of *P*410s and *P*660s in our synthetics against the corresponding amplitudes when we consider no attenuation (large Q_μ values). The amplitudes of *P*410s and *P*660s without attenuation and with topography effect removed are shown in Fig. 8 middle and bottom panel.

We consider that the smooth decrease of *P*410s amplitude with frequency indicates a relatively sharp velocity gradient, while the abrupt amplitude drop of *P*590s and *P*660s indicates broader velocity gradients. We estimate that the 410 converts *P*-waves with energy up to at least 0.32-0.64 Hz and that the 590 and 660, up to 0.12-0.2 Hz. Using Ineq. 4 and taking a *P* wave velocity -from *prem*- of about 9.1 km/s at the 410, 10 km/s at 590 km, and 10.8 km/s at the 660, suggests waves are being converted at discontinuities whose thicknesses are at most 7-14 km, 25-42 km and 27-45 km, respectively.

4.7 Comparison with Synthetic Data

4.7.1 Constraints on Velocity Contrast

The amplitude of a converted seismic phases is sensitive to velocity and density changes across a discontinuity. *Pds* phases are primarily sensitive to S-wave velocity jump (Ammon, 1991; Juliá, 2007). Since the velocity gradients act as low-pass filters to converted waves attenuating the high frequency energy, the amplitude of the conversions observed with low-frequency data are best used to infer the S-wave velocity contrast across the discontinuity. Therefore, assuming a range of values for the velocity contrast across the discontinuities 410, 590 and 660, we compare the theoretical amplitude of converted phases to the estimated amplitudes in order to obtain a rough assessment of the velocity jump across each discontinuity.

We use the elementary formulas 5.40 in Aki and Richards (2002) to compute the conversion coefficient (i.e., relative amplitude of *Pds* phase to *P* phase) for theoretical velocity contrasts for a range of slowness values of the incident P-wave. The slowness varies between 8.6 s/° and 6.5 s/° to account for the epicentral distance distribution of our data (mainly between 35° and 65°). We use the values of v_P , v_S and ρ in *prem* at 400 and 670 km, which are the depths of the 410 and 660 in this reference model. The v_S jumps across the 410 and 660 in *prem* are 3.4% and 6.7%, respectively. These values result in a conversion coefficient of 2-3% of the incident energy at the 410, and of 5-6% at the 660. Our estimated amplitudes for *P*410s and *P*660s at the lowest frequency

band -after removing the effects of discontinuity's topography and attenuation because the formulas 5.40 in Aki and Richards (2002) do not include these effects- are $(7.8 \pm 0.2)\%$ and $(11.6 \pm 0.3)\%$ of the vertical P-wave amplitude, respectively. Therefore, we compute the conversion coefficients for larger shear velocity jumps and find that our estimated amplitudes are consistent with a v_S jump of $(11.1 \pm 0.1)\%$ across the 410 and, of $(14.1 \pm 0.4)\%$ across the 660. The velocity contrasts at the 410 and the 660 are larger than the velocity jumps in the 1-D reference models (*prem*, *ak135*). The estimated amplitude of $P590s$ of $(7.2 \pm 0.2)\%$ of the incident P phase (after removing the effect of attenuation) is consistent with a decrease in v_S of $(8.8 \pm 0.2)\%$.

Fig. S6 shows the amplitude of $P410s$ and $P660s$ (without adjustments) in individual stacks as a function of RF number in each CCP bin. Individual bins with more than 20 RFs show amplitudes of $(8.5 \pm 2)\%$ for the 410 and $(9.7 \pm 2)\%$ for the 660, while those with 10-20 RFs show more anomalous values. The figure also shows that as the number of RF in individual bins increases, the amplitude of converted phases in individual stacks are closer to the amplitudes obtained in the global stacks after removing the effect of discontinuity topography. The amplitude in CCP bins can be affected by focussing effects due to the presence of heterogeneities. However, the fluctuations in our observed amplitudes is likely related to the small areas in the CCP stacks and the variable number of RFs inside each bin. We expect that our obtained amplitude values in the global stacks across the region are more robust to determine average velocity jumps.

4.7.2 Synthetic Velocity Gradients

We compute synthetic seismograms for *prem* using the reflectivity method (Fuchs & Müller, 1971) with discontinuity widths of 0, 20 and 30 km. We also compute synthetics for a modified *prem* model with shear wave velocity jumps of 10% at the 410 and 13% at the 660 with discontinuity widths of 0, 20 km for the 410 and of 0, 30 and 40 km for the 660 (Fig. 9). For the 660, we consider two different composite velocity gradients: (i) a 13% velocity jump formed by a linear v_S gradient of 6.5% and 10 km thick followed by a linear v_S gradient of 6.5% and 20 km thick (and 30 km thick); (ii) a 13% velocity jump formed by a linear v_S gradient of 11% and 10 km thick followed by a linear v_S gradient of 2% and 20 km thick (and 30 km thick). Although not shown in the figure, we also consider a linear v_S gradient of 13% and 30 km thick (and 40 km thick) which results in amplitude values very similar to the ones for composite velocity gradient (i). However, the composite velocity gradients are the only ones that show earlier arrival times for higher frequencies, which is what we observe for the real signal. Since P -to- S conversions are mainly sensitive to shear wave velocity jumps, P wave velocity and density in the synthetics are unmodified. Although, the analysis with real data shows a negative amplitude signal before $P660s$ probably related to a decrease in velocity with depth, for simplicity we don't include a negative velocity jump in this exploration. Fig. 9 shows both the results of the frequency dependence analysis of the real data (from the regional analysis in 4.1), and the results of the synthetic stacks for $P410s$ and $P660s$. **We also compute the mean value of the residuals as the relative amplitude in the model minus the observed relative amplitude ($\langle r \rangle$ in the legend of Fig. 9).**

Both the synthetic $P410s$ for the velocity jump of 3.4% (green lines in Fig. 9 a and b) and the one for a 10% velocity jump (red lines in Fig. 9 a and b) show a decrease in amplitude for increasing frequency, which is particularly noticeable for the broad transitions (dashed red and green lines in Fig. 9 b). In good agreement with the width estimation in 4.6, the observed $P410s$ -amplitude values are closer to the model with 10% v_S jump and 20 km discontinuity width, **which also shows the smallest mean residual.**

The observed amplitude of $P660s$ is better approximated with the composite velocity gradient (i), formed by two linear velocity gradients, the shallower of 6.5% and 10

Figure 9. Amplitudes of P-to-S converted phases as a function of frequency for synthetic cases compared to observed values (black solid) from the regional stack -with the topography effect removed (attenuation is included in the reflectivity method that we use). Shear wave velocities for the synthetics are plotted in a) for 410. The models represent a 3.4% velocity jump with a width of 0 km (green solid) and 20 km (green dashed), and a 10% velocity jump of 0 km (red solid) and 20 km (red dashed). Shear wave velocities for the 660 are plotted in c). The models represent a 6.7% velocity jump with a width of 0 km (green solid), 30 km (green dashed), a 13% velocity jump with a width of 0 km (red solid), a non-linear gradient composed of a velocity jump of 6.5% with a width of 10 km followed by a 6.5% velocity jump with a width of 20 km (red dashed), the same jumps for widths of 10 km and 30 km (red dotted), and a non-linear gradient composed of a 11% velocity jump with a width of 10 km followed by a 2.0% velocity jump with a width of 20 km (blue solid) and the same jumps for widths of 10 km and 30 km, respectively (blue dashed). The Pds phase amplitude dependence as a function of frequency is shown in b) for the 410 and in d) for the 660. **Error bars correspond to two times the standard error.** $\langle r \rangle$ value in the legend corresponds to the mean value of residuals (model - observations).

km thick and the deeper of 6.5% and 20 km (red-dashed lines in Fig. 9 d), **which shows the smallest mean residual**. There are probably other similar combinations of gradients that would also fit our observations. In this setting, the discontinuity depth at high frequencies would appear to be at the top of the discontinuity, and at lower frequencies, it may appear displaced downward on account of the time lag due to the longer interaction time with the gradient (e.g., Helffrich & Bina, 1994)

5 Discussion

5.1 Summary of Results

Beneath the study area, we obtain clear detections of converted phases at the 410 and 660 discontinuities for all frequency bands investigated. We also get robust observations of a precursory signal to *P660s* of negative amplitude, with absolute amplitude comparable to that of the *P660s* phase: we name it *P590s* in relation to the depth that a discontinuity responsible for this signal would have.

The 410 is on average deeper than its reference depth –at 410 km– and the results of the frequency analysis –smooth decrease of *P410s*’s amplitude with frequency– suggest that the 410 is a sharp discontinuity. From the analysis of synthetic RFs, the amplitude of *P410s* is consistent with a 10% velocity jump over less than 20 km.

The 660 depth varies with frequency; it is deeper by 18 ± 2 km for lower frequencies and close to reference –at 660 km– for higher frequencies. The results of the frequency analysis –abrupt amplitude drop of *P660s* between corner frequencies 0.08 Hz and 0.12 Hz– suggest that the 660 beneath the study area is a broad discontinuity. From the analysis of synthetic RFs, the amplitude analysis of *P660s* is consistent with a 13% composite velocity gradient, formed by two velocity jumps of similar magnitudes and different widths; the shallower v_S jump is sharper than the deeper one.

The variable depth of the 660 makes it difficult to determine a representative value of TZT in this area. However, taking only into account the depth values estimated from the frequency bands normally used in RFs studies of the TZ ($f_2 = 0.12, 0.2, 0.3$ Hz), the average TZT is slightly and consistently thinner than the reference value in *ak135* beneath this flat-slab region.

5.2 Thermal and compositional constraints from mineral physics

Fig. 10 shows theoretical values of relative depth (dh), thickness (depth interval of velocity gradient) and v_S jumps for the 410 and 660 discontinuities extracted from the seismological properties of a mantle comprised of an equilibrium assemblage (EA) of basalt and harzburgite (Xu et al., 2008; Stixrude & Lithgow-Bertelloni, 2011) as function of temperature. The same figure for a mechanical mixture (MM) model is shown in Fig. S7. The values of the three parameters (dh, thickness and v_S jumps) were extracted from the velocity profiles (for each composition and temperature) by visually picking the steepest gradients from the velocity profiles. The temperature axis in Fig. 10 (and Fig. S7) corresponds to surface potential temperature (temperature of a given portion of the mantle if it were adiabatically decompressed to the surface of the Earth).

If we consider the estimated depth of the TZ discontinuities for the velocity model that includes MIT08 anomalies without amplification (gray lines in Fig. 10 and Fig. S7), the relative depth of the 410 corresponds to a potential temperature of about 1725 ± 10 K for the EA pyrolitic model in Fig. 10 a (1685 ± 10 K for the MM model in Fig. S7 a). The relative depth of the 660 corresponds to a potential temperature of 1585 ± 10 K for the EA pyrolitic model in Fig. 10 d (1565 ± 25 K for the MM model in Fig. S7 d). The relative depths could be interpreted as the top of the TZ being affected by a hot anomaly and the bottom by a cold one, with a relative temper-

Figure 10. Results from the regional analysis (dashed lines and gray-solid lines) against mineral physics models (EA). Top: a) relative depth (bottom of discontinuity) for the velocity model that includes MIT08 anomalies (gray lines) and for the one that includes the anomalies multiplied by a factor of two (dashed-lines), b) thickness, and c) velocity jumps of the 410 discontinuity for different potential temperatures and basalt fractions, 0% (red, pure Harzburgite), 18% (black, Pyrolitic), and 40%. Bottom: Same parameters in d), e) and f) for the 660. Black dotted-vertical lines are observed values. For potential temperatures lower than 1400 K, the 660's behaviour in the three parameters considered is less clear due to the presence of olivine- and garnet-related phase transitions. **The thickness of the gray and black-dashed vertical lines in a and d corresponds to the standard error of the depth values.**

ature difference over 140 ± 14 K in the EA model (120 ± 27 K in MM). If we consider the estimated depth of the TZ discontinuities for the velocity model that includes MIT08 anomalies multiplied by a factor of two (black dashed lines in Fig. 10 and Fig. S7), the temperature difference inside the TZ is even larger. In global tomographic models, the TZ is indeed dominated by high velocity anomalies interpreted as a stagnating slab (e.g Fukao & Obayashi, 2013), consistently with the observation of the depressed 660. At the top of the TZ, however, there is no evidence of a low velocity (i.e., high temperature) anomaly to explain the depressed 410. In the EA model, the interpreted temperature anomaly can be slightly reduced when considering compositional differences. A deep 410 jump as observed here could be explained by a purely harzburgitic model with a potential temperature of 1685 ± 10 K (Fig. 10 a). The strongest observation, however, is that an olivine phase transition in pure harzburgite predicts the relatively large shear wave velocity jump of 10% observed here (both in EA Fig. 10 c and in MM Fig. S7 c).

Both the velocity jumps observed at the 410 and 660 (Fig. 10 c and f, respectively) are larger than the velocity jumps in either *prem* or *ak135* (Dziewonski & Anderson, 1981; Kennett, 1991). Regional seismic studies also find a large variability of the velocity jumps at TZ depth. Shearer (2000) compiled data for transition zone velocity models from different regional studies, mostly from continental regions, showing generally larger v_P and

v_S jumps at 410 km than those in *prem* and *ak135* models, and slightly lower values at 660 km. Cammarano et al. (2005) tested the compatibility of a constant pyrolite composition for the mantle (including the effects of phase changes) directly against the seismic data commonly used in the 1-D global seismic reference models. The resulting models fit the seismic data (P- and S-wave travel times and fundamental spheroidal and toroidal modes) as satisfactorily as *prem* or *ak135*. However, compared to *prem* or *ak135*, the resulting seismic profiles in Cammarano et al. (2005) have larger jumps near 410 km, lower jumps around 660 km, and stronger gradients directly below, consistent with more garnetitic mineralogy in the TZ.

Our estimate of the 410's velocity jump indicates a pure harzburgite composition (vertical dashed line in Fig. 10 c and Fig. S7 c). Large velocity jumps at the 410 discontinuity are also a signature of a dry composition (Buchen et al., 2018). For the 660 discontinuity, our estimation for a single large velocity jump does not fit with any of the compositional models considered. As mentioned above, the values in Fig. 10 were obtained considering the steepest v_S gradient in the velocity profiles for an EA model (composed of harzburgite and different proportions of basalt), while the velocity jump estimated from the data might include a more complex and broader gradient (probably composed of a sharp gradient followed by a broader one and both of similar magnitude, as explored in 4.7.2).

A complex 660 can result by the presence of the olivine and garnet phase transitions predicted around this depth. We only consider MM and EA models with basalt proportions showing similar olivine-related and garnet-related v_S gradients for potential temperature of 1600 K. We obtain a shear wave velocity jump of 9.5%-10% in a depth interval of ~ 53 km at the bottom of the TZ for an EA with 18%-25% of basalt. For the MM model, we obtain 9.1%-9.5% in a depth interval of ~ 75 km for 30%-40% of basalt. None of these models explain the 13% velocity jump that we estimate with our dataset. If the velocity jumps we estimate from the stacking of the whole dataset are overestimated, a better parameter to infer the composition of the mantle is to compare the ratio of the estimated velocity jumps ($r_E = \Delta v_{410} / \Delta v_{660} = 0.79 \pm 0.02$) against different mantle compositions. Again, we only consider MM and EA models with basalt proportions showing similar olivine-related and garnet-related $v_{S(660)}$ gradients for potential temperature of 1600 K. We find that the estimated ratio is consistent with the EA model with 18%-20% of basalt proportion ($r_{EA18\%} = 0.75$, $r_{EA20\%} = 0.76$) and the MM model with 30% of basalt ($r_{MM30\%} = 0.75$). **Other basalt proportions result in smaller or larger values of r_{EA} and r_{MM} : $r_{EA10\%} = 1.5$, $r_{EA30\%} = 0.57$, $r_{MM20\%} = 0.92$, $r_{MM40\%} = 0.51$**

From the amplitude analysis, we conclude that in the study area the 410 and 660 can be explained with a mantle with 18-30% of basalt at TZ depths.

5.3 Is the *P590s* a Sidelobe?

We observe a precursory signal to *P660s*, with the slowness of a converted phase and a predicted velocity gradient which corresponds to a decrease in shear-wave velocity of $(8.8 \pm 0.2)\%$, at a depth of about 590 km. Synthetic receiver functions based on commonly used 1-D reference models also show a negative amplitude signal before *P660s* that has been attributed to the coherent stacking of sidelobes (e.g. fig. 6 and 7 in Andrews and Deuss, 2008).

This negative signal in our synthetic RFs is shown in Fig. S8 (see Text S4). The signal that we identify as *negative lobe* (marked with an arrow in Fig. S8) shows some of the characteristics that might be attributed to a sidelobe. A summary of the synthetic stacks in Table S1 shows frequency dependent behavior to the synthetic negative sidelobe to the *P660s* phase: with increasing frequency, the negative lobe moves to larger relative time values and the *P660s* moves to smaller relative time values, decreasing the

overall differential travel time. A similar variability is observed in the real data across the lower frequencies $f_2 = 0.08, 0.12$ Hz (Table 2). For larger frequencies, however, the signal that we identify as *P590s* is clearly separated from *P660s*, and thus, it is unlikely to be a sidelobe (see Fig. 5).

The synthetics consistently show sidelobes for the main *P* phase and the *P410s*, in the lower frequency band, with absolute amplitudes of 30% of their related phase, and of 60% for the one related to the *P660s*. For the real data, the precursory signal is only present before the *P660s* and therefore less likely an artefact of the processing. Additionally the absolute amplitude of *P590s* is 85% of the amplitude of *P660s* (in the lower frequency band). We believe that the negative amplitude precursory signal contains information about the velocity structure of the transition zone near the 660 and that its interpretation does not necessarily requires a strong low velocity layer atop the 660. It is possible that the conversion at the 660 is causing a phase shift or some asymmetry in the waveform, and this is causing the negative lobe at lower frequencies. Therefore, we cannot exclude the possibility that, at lower frequencies, the *P590s* signal that we detect in the real dataset could be merged with a negative lobe that does not require a low velocity atop the 660. As mentioned above, at frequencies higher than 0.2 Hz, the negative signal that we identify as *P590s* is clearly not a sidelobe.

As a consequence of this analysis, we conclude that the amplitude of the *P590s* at lower frequencies is probably overestimated, implying that the related velocity drop is likely smaller than 8.8% (the value estimated in 4.7.1).

5.4 Nature of the Negative Precursory Signal to *P660s*

The negative precursory signal to *P660s* is also detected in other tectonic settings, with amplitudes corresponding to shear-wave velocity contrasts of -2% to -9% (Eagar et al., 2010; Shen & Blum, 2003; Shen et al., 2008; Tauzin et al., 2013; Bonatto et al., 2015; Tauzin et al., 2017). It has been suggested that this negative signal is the seismic signature of a global feature resulting from the accumulation of oceanic crust (MORB) at the base of the TZ (Shen & Blum, 2003; Shen et al., 2008).

At a depth of 600-700 km, the density of MORB is smaller than that of harzburgite (or a pyrolytic mantle) (Ringwood, 1982; Nakagawa et al., 2010), namely, a density crossover occurs –caused by different depths of the post-spinel and post-garnet transitions. Under the assumption of a small excess of basalt in the lower mantle, global-scale geodynamic simulations show that the density crossover in the transition zone is responsible for the trapping of basalt in the lower TZ (Ballmer et al., 2015; Nakagawa et al., 2010). The trapped basalt does not peel directly off slabs as they sink through the transition zone, but rather it is basalt that has separated from the slab at the CMB, and later been entrained in mantle flow (Nakagawa et al., 2010). The basalt layer in the TZ predicted by simulations (e.g. fig. 4 Ballmer et al. (2015)) might explain slab stagnation between depths of 650 km and 1000 km. Since the presence of basalt reduces seismic velocities at TZ depth (Xu et al., 2008), we would expect a decrease of velocity at the top of this basalt-enhanced layer and an increase at its base. The velocity contrast between pyrolite and basalt at a depth of 590 km is -3.3% (and between a MM mantle with 30-40% basalt and pure basalt is $-2.8/-2.4\%$), which is a fraction of the velocity drop of 8.8% we estimate at 590 km. As discussed in 5.3, the estimated amplitude of *P590s* at $f_2 = 0.8$ Hz is probably overestimated, and thus, the velocity jump as well. If the *P590s* is the evidence of a change in bulk chemistry, then we would expect it to be a sharp velocity gradient. Therefore, the amplitude of *P590s* at higher frequencies would still be representative of the velocity jump. Using the amplitude of *P590s* at $f_2 = 0.2$ Hz, $P590s/P = (-3.2 \pm 0.2)\%$, we estimate a shear-wave velocity jump of $\Delta v_s = (-3.9 \pm 0.3)\%$.

As already mentioned, a possible scenario that would explain signal *P590s* is basalt entrained in mantle flow and accumulated over time at the lower TZ. It might seem that

Figure 11. Schematic summary of results and interpretation. Observed discontinuity jumps and widths are shown compared to PREM discontinuity jumps. The column on the right represents the interpreted composition of the transition zone.

there is a lack of coherence between the absence of low velocity anomalies at TZ depths in the global tomographic images and the presence of a low velocity heterogeneity (MORB like composition) that would explain the negative amplitude signal identified as P590s. However, at TZ depth, the resolution of tomography is likely too low to detect a thin compositional heterogeneity, although potential evidence for a signal of compositional layering around the 660 has been interpreted in a long-wavelength tomographic model beneath the Samoan hotspot (Maguire et al., 2017).

Alternatively, we know from tomography that there is likely a horizontal slab stagnating at transition zone depths probably further East (e.g., Fukao & Obayashi, 2013). If the P590s phase represents the subducted oceanic crust of the stagnant slab could be explored through more detailed modelling including a thermal model of the slab. However, this analysis is beyond the scope of this paper.

6 Conclusions

Our analysis shows that beneath the Chile-Argentina flat-subduction the transition zone is characterized by (See Fig. 11 for schematic summary):

1) a relatively sharp 410 discontinuity, deeper than reference by (10 ± 1) km and (18 ± 2) km in the higher and lower frequency bands, respectively (or by (7 ± 1) km and (15 ± 2) km, depending on the velocity model used to compute time corrections), and consistent with a 10% shear-wave velocity jump of less than 20 km radial thickness;

2) a broad 660 discontinuity, deeper by up to (18 ± 2) (or (15 ± 1) km, depending on the velocity model used to compute time corrections) for lower frequencies and close to reference depth for higher frequencies, and consistent with a 13% shear-wave velocity gradient, likely composite of a velocity jump and broad gradient across 30-40 km, of similar magnitudes;

3) the presence of a negative amplitude precursory signal to P660s (named P590s), with absolute amplitude comparable to that of the P660s phase.

Even though depth-variability of the 660 hampers the determination of a representative value for the TZT in this area, taking average depths of the 410 and 660 discon-

tinuities from intermediate frequency bands of (423 ± 1) km and (668 ± 2) km, respectively, the TZZ results (245 ± 2) km thick, which is slightly thinner than the 250 km reference value in *ak135*. The seismic velocity structure of the upper mantle in the region suggests that the hypothesis of a purely thermal origin for the depressed 410 is unlikely. A better understanding on this could be provided by a higher resolution velocity model of the study area.

If we trust the absolute amplitudes of *P*410s and *P*660s, we could interpret a purely harzburgitic composition atop the TZ in the study area. The absolute amplitude of the observed *P*660s cannot be entirely explained with the compositional models explored. However, when interpreting the amplitude ratio between the 660 and 410, a near pyroclitic composition fits best. This interpretation is shown in Fig. 11. The combination of the velocity gradients due to the phase transitions in the olivine and garnet components of the pyroclitic mantle would explain the broad 660 and the large velocity jump estimated. The negative *P*590s signal in this region could be signal from the velocity reduction due to basalt accumulation at the base of the transition zone.

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References

- Aki, K., & Richards, P. (2002). *Quantitative seismology*. University Science Books.
- Allègre, C. J., & Turcotte, D. L. (1986). Implications of a two-component marble-cake mantle. *Nature*, *323*, 123-127.
- Ammon, C. J. (1991). The isolation of receiver effects from teleseismic p waveforms. *Bull. Seism. Soc. Am.*, *81*(6), 2504-2510.
- Anderson, D. L., & Bass, J. D. (1986). Transition region of the earth's upper mantle. *Nature*, *320*, 321-328. doi: 10.1038/320321a0
- Ballmer, M. D., Schmerr, N. C., Nakagawa, T., & Ritsema, J. (2015). Compositional mantle layering revealed by slab stagnation at 1000-km depth. *Science Advances*, *1*(11). doi: 10.1126/sciadv.1500815
- Bastow, I. D. (2012). Relative arrival-time upper-mantle tomography and the elusive background mean. *Geophysical Journal International*, *190*(2), 1271-1278. doi: 10.1111/j.1365-246X.2012.05559.x
- Beck, S., Wallace, T., & Zandt, G. (2000). Slab geometry in the southern andes.

- International Federation of Digital Seismograph Networks. Dataset/Seismic Network. doi: 10.7914/SN/YC_2000
- Beck, S., & Zandt, G. (2007). Lithospheric structure and deformation of the flat slab region of argentina. *International Federation of Digital Seismograph Networks..* doi: 10.7914/SN/ZL_2007
- Benz, H. M., & Vidale, J. E. (1993). Sharpness of upper-mantle discontinuities determined from high-frequency reflections. *Nature*, *365*(6442), 147-150. doi: 10.1038/365147a0
- Bina, C. R., & Helffrich, G. (1994). Phase transition clapeyron slopes and transition zone seismic discontinuity topography. *Journal of Geophysical Research*, *99*(B8), 15853-15860. doi: doi:10.1029/94JB00462
- Bonatto, L., Schimmel, M., Gallart, J., & Morales, J. (2015). The upper-mantle transition zone beneath the ibero-maghrebien region as seen by teleseismic pds phases. *Tectonophysics*, *663*, 212 - 224. (Special issue on Iberia geodynamics: An integrative approach from the Topo-Iberia framework) doi: 10.1016/j.tecto.2015.02.002
- Bostock, M. G. (1999). Seismic waves converted from velocity gradient anomalies in the earth's upper mantle. *Geophys. J. Int.*, *138*(3), 747-756. doi: 10.1046/j.1365-246X.1999.00902.x
- Brandenburg, J. P., & van Keken, P. E. (2007). Methods for thermochemical convection in earth's mantle with force-balanced plates. *Geochemistry, Geophysics, Geosystems*, *8*(11). Retrieved from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1029/2007GC001692> doi: 10.1029/2007GC001692
- Buchen, J., Marquardt, H., Speziale, S., Kawazoe, T., Ballaran, T. B., & Kurnosov, A. (2018). High-pressure single-crystal elasticity of wadsleyite and the seismic signature of water in the shallow transition zone. *Earth and Planetary Science Letters*, *498*, 77 - 87. doi: 10.1016/j.epsl.2018.06.027
- Cammarano, F., Deuss, A., Goes, S., & Giardini, D. (2005). One-dimensional physical reference models for the upper mantle and transition zone: Combining seismic and mineral physics constraints. *J. Geophys. Res.*, *110*(B1), B01306. doi: 10.1029/2004JB003272
- Clayton, R. W., & Wiggins, R. A. (1976). Source shape estimation and deconvolution of teleseismic body waves. *J. R. Astr. Soc.*, *47*(1), 151-177. doi: 10.1111/j.1365-246X.1976.tb01267.x
- Collier, J., & Helffrich, G. (1997). Topography of the "410" and "660" km seismic discontinuities in the izu-bonin subduction zone. *Geophys. Res. Lett.*, *24*(12), 1535-1538. doi: 10.1029/97GL01383
- Collier, J. D., & Helffrich, G. R. (2001). The thermal influence of the subducting slab beneath south america from 410 and 660 km discontinuity observations. *Geophys. J. Int.*, *147*(2), 319-329. doi: 10.1046/j.1365-246X.2001.00532.x
- Crotwell, H. P., Owens, T. J., & Ritsema, J. (1999). The taup toolkit: Flexible seismic travel-time and ray-path utilities. *Seismological Research Letters*, *70*, 154-160. doi: 10.1785/gssrl.70.2.154
- Durek, J. J., & Ekström, G. (1996, 02). A radial model of anelasticity consistent with long-period surface-wave attenuation. *Bulletin of the Seismological Society of America*, *86*(1A), 144-158.
- Dziewonski, A. M., & Anderson, D. L. (1981). Preliminary reference earth model. *Physics of the Earth and Planetary Interiors*, *25*(4), 297 - 356. doi: 10.1016/0031-9201(81)90046-7
- Eagar, K. C., Foucha, M. J., & James, D. E. (2010). Receiver function imaging of upper mantle complexity beneath the pacific northwest, united states. *Earth Planet. Sci. Lett.*, *297*(1-2), 140-152. doi: 10.1016/j.epsl.2010.06.015
- Farra, V., & Vinnik, L. P. (2000). Upper mantle stratification by p and s receiver functions. *Geophys. J. Int.*, *141*(3), 699-712. doi: 10.1046/j.1365-246x.2000.00118.x

- Fromm, R., Zandt, G., & Beck, S. (2004). Crustal thickness beneath the andes and sierras pampeanas at 30°s inferred from pn apparent phase velocities. *Geophys. Res. Lett.*, *31*(L06625). doi: 10.1029/2003GL019231
- Frost, D. (2003). The structure and sharpness of $(mg, fe)_2sio_4$ phase transformations in the transition zone. *Earth Planet. Sci. Lett.*, *216*, 313-328. doi: 10.1016/S0012-821X(03)00533-8
- Fuchs, K., & Müller, G. (1971). Computation of synthetic seismograms with the reflectivity method and comparison with observations. *Geophys. J. Int.*, *23*(4), 417-433. doi: 10.1111/j.1365-246X.1971.tb01834.x
- Fukao, Y., & Obayashi, M. (2013). Subducted slabs stagnant above, penetrating through, and trapped below the 660 km discontinuity. *Journal of Geophysical Research: Solid Earth*, *118*(11), 5920-5938. (2013JB010466) doi: 10.1002/2013JB010466
- Gans, C. R., Beck, S. L., Zandt, G., Gilbert, H., Alvarado, P., Anderson, M., & Linkimer, L. (2011). Continental and oceanic crustal structure of the pampean flat slab region, western argentina, using receiver function analysis: new high-resolution results. *Geophys. J. Int.*, *186*, 45-58. doi: 10.1111/j.1365-246X.2011.05023.x
- Gilbert, H. (2008). Eastern sierras pampeanas, lithospheric structure above the variably dipping nazca slab. *International Federation of Digital Seismograph Networks*. doi: 10.7914/SN/XH_2008
- Gilbert, H., Beck, S., & Zandt, G. (2006). Lithospheric and upper mantle structure of central chile and argentina. *Geophys. J. Int.*, *165*, 383-398. doi: 10.1111/j.1365-246X.2006.02867.x
- Goldstein, A., P. & Snoke. (2005, 4). Sac availability for the iris community. *Incorporated Research Institutions for Seismology Newsletter*, *7*(1).
- Gutscher, M. (2002). Andean subduction styles and their effect on thermal structure and interplate coupling. *Journal of South American Earth Sciences*, *15*(1), 3 - 10. doi: 10.1016/S0895-9811(02)00002-0
- Gutscher, M., Spakman, W., Bijwaard, H., & Engdahl, E. (2000). Geodynamics of flat subduction: seismicity and tomographic constraints from the andean margin. *Tectonics*, *19*(5), 814-833. doi: 10.1029/1999TC001152
- Hayes, G., Wald, D., & Johnson, R. (2012). Slab 1.0: A three-dimensional model of global subduction zone geometries. *J. Geophys. Res.*, *117*(B01302). doi: 10.1029/2011JB008524
- Heit, B., Yuan, X., Bianchi, M., Sodoudi, F., & Kind, R. (2008). Crustal thickness estimation beneath the southern central andes at 30°s and 36°s from s wave receiver function analysis. *Geophys. J. Int.*, *174*, 249-254. doi: 10.1111/j.1365-246X.2008.03780.x
- Heit, B., Yuan, X., Kind, R., & Asch, G. (2007). Lithospheric dynamics in the southernmost andean plateau (pudel). *Deutsches GeoForschungsZentrum GFZ. Other/Seismic Network*. doi: 10.14470/7O092361
- Helfrich, G. (2000). Topography of the transition zone seismic discontinuities. *Reviews of Geophysics*, *38*(1), 141-158. doi: 10.1029/1999RG000060
- Helfrich, G., & Bina, C. R. (1994). Frequency dependence the visibility depths mantle of and of seismic discontinuities. *Geophys. Res. Lett.*, *21*(24), 2613-2616. doi: 10.1029/94GL02341
- Ito, E., & Takahashi, E. (1989). Postspinel transformations in the system mg_2sio_4 - fe_2sio_4 and some geophysical implications. *Journal of Geophysical Research: Solid Earth*, *94*(B8), 10637-10646. doi: 10.1029/JB094iB08p10637
- Jenkins, J., Deuss, A., & Cottaar, S. (2017). Converted phases from sharp 1000 km depth mid-mantle heterogeneity beneath western europe. *Earth and Planetary Science Letters*, *459*(Supplement C), 196 - 207. doi: 10.1016/j.epsl.2016.11.031
- Juliá, J. (2007). Constraining velocity and density contrasts across the

- crust—mantle boundary with receiver function amplitudes. *Geophys. J. Int.*, 171(1), 286-301. doi: 10.1111/j.1365-2966.2007.03502.x
- Katsura, T., & Ito, E. (1989). The system $mg_2sio_4 - fe_2sio_4$ at high pressure and temperature: precise determination of stabilities of olivine, modified spinel and spinel. *J. Geophys. Res.*, 94(B11), 15663-15670. doi: 10.1029/JB094iB11p15663
- Katsura, T., Yamada, H., Nishikawa, O., Song, M., Kubo, A., Shinmei, T., . . . Ito, E. (2004). Olivine-wadsleyite transition in the system $(mg, fe)_2sio_4$. *J. Geophys. Res.*, 109, B02209. doi: 10.1029/2003JB002438
- Kennett, E. R., B. L. N. & Engdahl. (1991). Traveltimes for global earthquake location and phase identification. *Geophys. J. Int.*, 105(2), 429-465. doi: 10.1111/j.1365-246X.1991.tb06724.x
- Kustowski, B., Ekström, G., & Dziewoński, A. M. (2008). Anisotropic shear-wave velocity structure of the earth's mantle: A global model. *Journal of Geophysical Research: Solid Earth*, 113(B6). (B06306) doi: 10.1029/2007JB005169
- Landes, M., Ritter, J. R. R., O'Reilly, B. M., Readman, P. W., & Do, V. C. (2006). A n-s receiver function profile across the variscides and caledonides in sw ireland. *Geophys. J. Int.*, 166(2), 814-824. doi: 10.1111/j.1365-246X.2006.03052.x
- Langston, C. A. (1979). Structure under mount rainier, washington, inferred from teleseismic body waves. *J. Geophys. Res.*, 84(B9), 4749-4762. doi: 10.1029/JB084iB09p04749
- Lawrence, J. F., & Shearer, P. M. (2006). A global study of transition zone thickness using receiver functions. *J. Geophys. Res.*, 111, B06307. doi: 10.1029/2005JB003973
- Li, C., van der Hilst, R. D., Engdahl, E. R., & Burdick, S. (2008). A new global model for p wave speed variations in earth's mantle. *Geochemistry, Geophysics, Geosystems*, 9(5). (Q05018) doi: 10.1029/2007GC001806
- Litasov, K. D., & Ohtani, E. (2005). Phase relations in hydrous morb at 18-28 gpa: Implications for heterogeneity of the lower mantle. *Phys. Earth Planet. Inter.*, 150, 239-263. doi: 10.1016/j.pepi.2004.10.010
- Lynner, C., Anderson, M. L., Portner, D. E., Beck, S. L., & Gilbert, H. (2017). Mantle flow through a tear in the nazca slab inferred from shear wave splitting. *Geophysical Research Letters*, 44(13), 6735-6742. doi: 10.1002/2017GL074312
- Maguire, R., Ritsema, J., & Goes, S. (2017). Signals of 660 km topography and harzburgite enrichment in seismic images of whole-mantle upwellings. *Geophysical Research Letters*, 44(8), 3600-3607. Retrieved from <https://agupubs.onlinelibrary.wiley.com/doi/abs/10.1002/2017GL073120> doi: 10.1002/2017GL073120
- Nakagawa, T., Tackley, P. J., Deschamps, F., & Connolly, J. A. (2010). The influence of morb and harzburgite composition on thermo-chemical mantle convection in a 3-d spherical shell with self-consistently calculated mineral physics. *Earth and Planetary Science Letters*, 296(3), 403 - 412. doi: 10.1016/j.epsl.2010.05.026
- Obayashi, M., Yoshimitsu, J., Nolet, G., Fukao, Y., Shiobara, H., Sugioka, H., . . . Gao, Y. (2013). Finite frequency whole mantle p wave tomography: Improvement of subducted slab images. *Geophysical Research Letters*, 40(21), 5652-5657. (2013GL057401) doi: 10.1002/2013GL057401
- Paulssen, H. (1988). Evidence for a sharp 670-km discontinuity as inferred from p-to-s converted waves. *J. Geophys. Res.*, 93(B9), 10489-10500. doi: 10.1029/JB093iB09p10489
- Phinney, R. (1964). Structure of the earth's crust from spectral behavior of long-period body waves. *J. Geophys. Res.*, 69(14), 2997-3017. doi: 10.1029/JZ069i014p02997

- Portner, D. E., Beck, S., Zandt, G., & Scire, A. (2017). The nature of slab slow velocity anomalies beneath south america. *Geophysical Research Letters*, 44(10), 4747-4755. doi: 10.1002/2017GL073106
- Richards, P. (1972). Seismic waves reflected from velocity gradient anomalies within the earth's upper mantle. *J. Geophys. Res.*, 38, 517-527. doi: 10.1046/j.1365-246x.1999.00902.x
- Ringwood, A. (1970). Phase transformations and the constitution of the mantle. *Physics of the Earth and Planetary Interiors*, 3, 109 - 155. doi: 10.1016/0031-9201(70)90047-6
- Ringwood, A. (1991). Phase transformations and their bearing on the constitution and dynamics of the mantle. *Geochimica et Cosmochimica Acta*, 55(8), 2083 - 2110. Retrieved from <http://www.sciencedirect.com/science/article/pii/001670379190090R> doi: 10.1016/0016-7037(91)90090-R
- Ringwood, A. E. (1966). The Chemical Composition and Origin of the Earth. In *Advances in earth science* (Vol. 65).
- Ringwood, A. E. (1982). Phase transformations and differentiation in subducted lithosphere: Implications for mantle dynamics, basalt petrogenesis, and crustal evolution. *The Journal of Geology*, 90(6), 611-643. Retrieved from 10.1086/628721 doi: 10.1086/628721
- Ritsema, J., Deuss, A., van Heijst, H. J., & Woodhouse, J. H. (2011). S40rts: a degree-40 shear-velocity model for the mantle from new rayleigh wave dispersion, teleseismic traveltime and normal-mode splitting function measurements. *Geophysical Journal International*, 184(3), 1223-1236. doi: 10.1111/j.1365-246X.2010.04884.x
- Ritsema, J., & Van Heijst, H. J. (2002). Constraints on the correlation of p- and s-wave velocity heterogeneity in the mantle from p, pp, ppp and pkpab travel times. *Geophys. J. Int.*, 149(2), 482-489. doi: 10.1046/j.1365-246X.2002.01631.x
- Schimmel, M., & Paulssen, H. (1997). Noise reduction and detection of weak, coherent signals through phase-weighted stacks. *Geophys. J. Int.*, 130(2), 497-505. doi: 10.1111/j.1365-246X.1997.tb05664.x
- Schmandt, B. (2012). Mantle transition zone shear velocity gradients beneath us-array. *Earth Planet. Sci. Lett.*, 355-356, 119-130. doi: 10.1016/j.epsl.2012.08.031
- Shearer, P. M. (2000). Upper mantle seismic discontinuities. In S.-I. Karato, A. Forte, R. Liebermann, G. Masters, & L. Stixrude (Eds.), *Earth's deep interior: Mineral physics and tomography from the atomic to the global scale, geophysical monographs* (Vol. 117, p. 115-131). Washington, D. C.: AGU.
- Shen, X., & Blum, J. (2003). Seismic evidence for accumulated oceanic crust above the 660-km discontinuity beneath southern africa. *Geophys. Res. Lett.*, 30(18), 1925. doi: 10.1029/2003GL017991
- Shen, X., Zhou, H., & Kawakatsu, H. (2008). Mapping the upper mantle discontinuities beneath china with teleseismic receiver functions. *Earth Planets Space*, 60, 713-719. doi: 10.1186/BF03352819
- Stixrude, L., & Lithgow-Bertelloni, C. (2005). Thermodynamics of mantle minerals — I. Physical properties. *Geophysical Journal International*, 162(2), 610-632. doi: 10.1111/j.1365-246X.2005.02642.x
- Stixrude, L., & Lithgow-Bertelloni, C. (2011). Thermodynamics of mantle minerals — ii. phase equilibria. *Geophysical Journal International*, 184(3), 1180-1213. doi: 10.1111/j.1365-246X.2010.04890.x
- Tauzin, B., Kim, S., & Kennett, B. (2017). Pervasive seismic low-velocity zones within stagnant plates in the mantle transition zone: Thermal or compositional origin? *Earth and Planetary Science Letters*, 477, 1 - 13. doi: 10.1016/j.epsl.2017.08.006
- Tauzin, B., van der Hilst, R. D., Wittlinger, G., & Ricard, Y. (2013). Multiple tran-

- sition zone seismic discontinuities and low velocity layers below western united states. *J. Geophys. Res. Solid Earth*, 118, 2307-2322. doi: 10.1002/jgrb.50182
- van der Meijde, M., Marone, F., Giardini, D., & van der Lee, S. (2003). Seismic evidence for water deep in earth's upper mantle. *Science*, 300(5625), 1556-1558. doi: 10.1126/science.1083636
- Vidale, J. E., & Benz, H. M. (1992). Upper-mantle seismic discontinuities and the thermal structure of subduction zones. *Nature*, 356(6371), 678-683. doi: 10.1038/356678a0
- Vidale, J. E., Ding, X.-Y., & Grand, S. P. (1995). The 410-km-depth discontinuity: A sharpness estimate from near-critical reflections. *Geophys. Res. Lett.*, 22(19), 2557-2560. doi: 10.1029/95GL02663
- Vinnik, L. (1977). Detection of waves converted from p to sv in the mantle. *J. Phys. Earth Planet. Inter.*, 15(1), 39-45. doi: 10.1016/0031-9201(77)90008-5
- Vinnik, L., Oreshin, S., Kosarev, G., Kiselev, S., & Makeyeva, L. (2009). Mantle anomalies beneath southern africa: evidence from seismic s and p receiver functions. *Geophys. J. Int.*, 179, 279-298. doi: 10.1111/j.1365-246X.2009.04261.x
- Wessel, P., & Smith, W. H. F. (1991). Free software helps map and display data. *EOS Trans., AGU*, 72, 441.
- Xu, W., Lithgow-Bertelloni, C., Stixrude, L., & Ritsema, J. (2008). The effect of bulk composition and temperature on mantle seismic structure. *Earth and Planetary Science Letters*, 275(1), 70 - 79. doi: 10.1016/j.epsl.2008.08.012
- Yañez, G., Ranero, C., Von Huene, R., & Diaz, J. (2001). Magnetic anomaly interpretation across the southern central andes (32° -34° s), the role of the juan fernández ridge in the late tertiary evolution of the margin. *Journal of Geophysical Research: Solid Earth*, 106(B4), 6325-6345. doi: 10.1029/2000JB900337
- Yuan, X., Kind, R., Li, X., & Wang, R. (2006). The s receiver functions: synthetics and data example. *Geophys. J. Int.*, 165(2), 555. doi: 10.1111/j.1365-246X.2006.02885.x