Deep source regions for Patagonia backarc volcanism imaged by finite frequency body wave tomography

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Abstract

The subduction of Chile Ridge beneath South America beginning 12-16 Myr ago opened a gap in the subducting slab beneath southern Patagonia, which migrated northward and is located today at 46ºS. Geodynamic processes associated with the slab window are poorly understood. Here we apply finite-frequency P and S body wave tomography to seismic data from several temporary arrays as well as regional stations to image seismic heterogeneities down to 650 km depth. The results show strong low velocity anomalies extending to 400 km depth beneath recent back-arc volcanism between 46ºS and 48ºS, suggesting a link to thermal upwelling in the upper mantle. The southern edge of the Nazca slab extends aseismically down to at least 350 km and has steeper dip than previously suggested. We also image low upper mantle seismic velocities beneath the Patagonia icefields, suggesting low viscosity modulates the patterns of uplift and horizontal deformation observed by GNSS.

Plain summary

The presence of a tear in the downgoing plate brings hot material from the asthenosphere and modifies locally the mantle flow associated with the subduction zone. Patagonia is a good example where the subduction of the spreading ridge created an opening in the slab. Using arrival times of P and S waves, we develop a new seismic model of the mantle in this region. We show strong back-arc slow velocity anomalies extending to 400 km depth beneath recent back arc volcanism between 46ºS and 48ºS, suggesting this volcanism is linked to upwelling in the upper mantle. We show that the southern edge of the Nazca slab extends to at least 350 km and has steeper dip than previously suggested. Additionally, our new model suggests lateral variations in mantle viscosity that help to explain the present-day bedrock deformation across the Patagonia icefield.

1. Introduction

The subduction of a spreading ridge opens a window in the subducting slab, allowing the asthenosphere to upwell and interact with the shallow mantle and crust. Slab windows have a profound influence on geodynamics processes, mantle flow pattern, and thermal structure of subduction zones [Groome and Thorkelson, 2009; Sanhueza et al, 2023]. Magmatism in a slab window is distinctly different from that found in typical volcanic arcs, with adakitic melts associated with slab edges and a volcanic gap in the center of the window [Thorkelson and Breitsprecher, 2005]. Magmatism may shift to the backarc region, with tholeiitic lavas presumably due to mantle upwelling observed [Gorring et al, 1997; Thorkelson and Breitsprecher, 2005]. However, it is unclear whether this magmatism results from uppermost mantle processes or is instead due to deeper mantle upwelling.

The Patagonia slab window (**Figure 1a**) formed 14 Ma in Southern Patagonia with subduction of the Chilean spreading ridge beneath South America is a good example of a migrated slab window [Breitsprecher and Thorkelson, 2009]. Geochemical studies identified the variation of geochemical signature of magma along the subduction with a transition from adakitic to basaltic signature in the slab melt and presence of an asthenospheric window [Thorkelson and Breitsprecher, 2005]. The back arc region shows widespread recent volcanic activity with the development of massive plateau lavas [Ramos and Kay 1992; Gorring et al. 1997; Guivel et al., 2006, Guest et al, 2024]. From analogue modelling and the compilation of available ages of back arc magmatism and kinematic reconstruction of the edge of the Antarctic slab, Guillaume et al. (2010) suggest that recent volcanism could be due to lateral flow of sub-slab mantle.

The Patagonian slab window underlies the Northern and Southern Patagonian Icefields (NPI and SPI), so the thermal and rheological perturbations resulting from the window may be important for the solid earth response to the ice load [Klemann et al, 2007; Lange et al, 2014; Mark et al, 2022; Russo et al, 2022]. Currently the NPI and SPI are undergoing rapid ice mass loss [Richter et al. 2019], associated with a fast bedrock uplift [Dietrich et al. 2010; Richter et al. 2016]. The intensity of the solid-earth response to the glacial retreat can be explained by the regional scale rheology with low viscosity in the asthenosphere and a thin lithosphere [Lange et al. 2014; Mark et al. 2022; Russo et al. 2022].

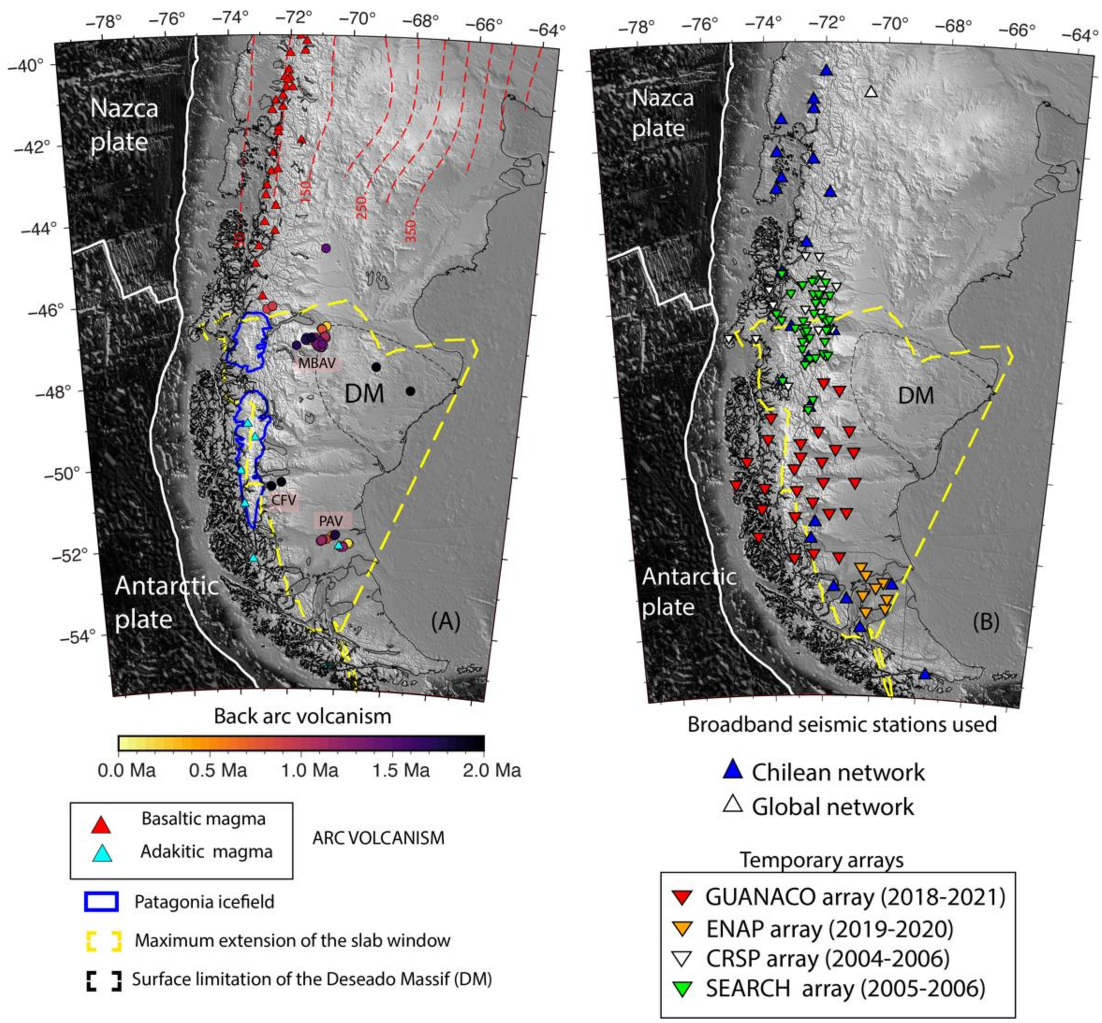

Figure 1. *(Left) Topographic map of Southern Patagonia with main tectonics features.* Red dashed lines represent iso-depth of the Nazca slab (50 km interval). *Colored circles represent young backarc volcanism (<2Myrs). MBAV: Meseta del Lago Buenos Aires Volcanic field, CFV: Cerro del Fraile volcanics, PAV: Pali Aike Volcanic field (Right) Broadband seismic stations used in this study.*

A recent regional scale Rayleigh wave imaging study [Mark et al. 2022] shows the presence of a slow velocity anomaly in the uppermost mantle which is interpreted as a thermal erosion of the lithospheric mantle over the youngest part of the slab window. From shear wave splitting analysis, Ben-Mansour et al. (2022) found a strong anisotropy with low uppermost mantle shear velocities and an absence of mantle lithosphere. In addition, other studies have imaged the structure immediately beneath the Chile Triple Junction (CTJ) [Russo et al, 2010; Miller et al, 2023] and provided large-scale regional tomography [Portner et al, 2020; Kondo et al, 2024].

Here we use the data-adaptive, multiscale tomographic approach of Hung et al. [2011] and relative sensitivity kernels to combine asynchronous datasets from several different temporary deployments [Maupin 2021] to map the 3-dimensional structure of the Patagonia slab window. The new P and SH seismic velocity models yield broad regional coverage, with better resolution than large-scale models and extend to greater depths than Rayleigh wave models, which are limited to the upper 150-200 km. We provide new constraints on the present-day state of the mantle and how it contributes to back arc magmatism. We also provide additional constraints on the manner in which earth structure contributes to the fast crustal uplift beneath the Patagonia icefields.

2. <u>Data and methods</u>

2.1. Broadband seismic data and processing

The dataset for this study (**Figure 1**) consists of seismograms from the GUANACO network of twenty-six broadband seismographs deployed from 2018-2021 [Wiens and Magnani, 2018], as well as previous deployments in the Chile triple junction region [Russo et al, 2010; Miller et al, 2023]. Additionally, we use permanent stations from the Chile Seismic Network [Barrientos and CSN team, 2018] and the global network. We used three component seismic waveforms of these stations from teleseismic events between 28º and 98º of epicentral distance and magnitudes greater than 5.4.

Numerous studies show that body wave travel time delays are frequency dependent [Hung et al. 2004; Hung et al.; 2011; Kolstrup and Maupin 2015]. To account for frequency-dependent delays, we measure direct *P* and *SH wave* travel times on waveforms band-pass-filtered around the secondary seismic noise peak (~0.2 Hz). We use two frequency bands for P-waves (0.03-0.125 Hz and 0.3-1.5 Hz) and one frequency band for S-waves (0.03-0.125 Hz). Relative travel times in different frequency bands are estimated from an automated processing procedure [Kolstrup and Maupin,2015] combining an iterative cross-correlation and stack algorithm (ICCS) [Lou et al., 2013] and the multichannel cross-correlation method [Van Decar and Crosson, 1990].

Data are corrected for crustal delay times using the crustal model from Mark et al. (2022) (**Figure S1**). To ensure a common reference time and include the effect of the topography, the crustal corrections are computed at each station by considering a model from 50 km below sea level and up to the free surface. In regions of thick sediments, reverberations and their interference introduce a significant frequency dependence of the traveltimes measured on band-passed waveforms [Yang and Shen, 2006; Ritsema et al., 2009; Kolstrup and Maupin, 2015]. The crustal correction of finite-frequency travel time residuals should therefore be made frequency dependent. This is done here following the procedure of Kolstrup and Maupin (2015).

2.2. Finite frequency body wave tomography

To invert the relative travel times to obtain relative seismic velocity anomalies, we use the 3-D Born-Fréchet kernels, which link the influence of velocity heterogeneities and finite frequency travel time shifts [Dahlen et al., 2000; Hung et al. 2000; Schmandt and Humphreys, 2010; Maupin and Kolstrup, 2015]. The kernels are demeaned to form relative kernels to correct for the unevenness in space and time of the data distribution [Maupin, 2021].

We use the data-adaptive, multiscale, finite-frequency tomography method of Hung et al (2011) to take advantage of the distribution of seismic stations and the azimuthal distribution of teleseismic events. The model is parametrized in terms of wavelet basis functions in 3-D, where the calculations assume an isotropic Earth. The model has 65x33 points in the horizontal plane with interspacing of 0.31°. The model extends from 50 to 650km depth with a 37.5 km grid spacing, but shallower and deeper models have also been tested (**Figure S2**). The inversion includes station terms, but tests with different weights for those show that they do not significantly affect the inverted model. As damping acts on the wavelet coefficients, there is no norm or gradient damping, but an ideal preservation of the resolved elements in the model [Hung et al. 2011, Kolstrup et al. 2015]. The resulting models do not provide absolute velocities but only velocity anomalies relative to an unknown average 1-D model. The optimal models for P and S are determined by selecting the models with the best compromise between model norm and variance on one side and data misfit reduction on the other side. The misfit reduction is high, well above 60% even for smooth models, and 88% and 86% for the chosen P and S models respectively (**Figure S3**).

3. Results

3.1. P and SH relative velocity models

The new finite-frequency P and SH models provide improved resolution and image the mantle down to 650 km (**Figure 2**). Overall, the new models show larger velocity contrasts compared to previous results [Portner et al. 2022; Kondo et al. 2024], but the tomographic inversion used here conserves amplitudes quite well compared to other methods. The major patterns of the P and SH models are similar, with the magnitude of the S wave anomalies larger than the P wave anomalies as expected. There are some differences between the P and SH models in the detailed structure of the anomalies. These differences could be due to the unequal effects of the compositional and thermal anomalies on P and S wave velocities. However, because these differences could alternatively result from differences in data volume, signal to noise ratio, and frequency content, we do not extensively describe and interpret these differences and reserve most of our interpretation to the larger patterns reflected in both P and SH models. (**Figure 2**).

The most prominent velocity anomaly is a large region of slow P and SH velocities anomalies (-4/-6%) between 46°S and 49°S (**Figures 2 and 3**), centered beneath the Meseta del Lago Buenos Aires volcanic region [Guivel et al., 2006, Guest et al, 2024]. Recent surface wave [Mark et al. 2022] and body wave [Russo et al. 2010, Portner et al., 2020; Rodriguez et al., 2021; Miller et al. 2023; Kondo et al. 2024] seismic models have shown the presence of a similar anomaly in this region but have not resolved its lateral and depth extent. Both P and S wave models show it extending down to at least 400 km, where it is underlain by faster velocities in the P-wave model. The anomaly is truncated in the north by faster velocities likely associated with the subducting Nazca slab. To the east, the lateral extent of the anomaly is limited at shallower depths (< 200 km) by the lithospheric mantle of the Deseado Massif (**Figures 2 and 3**). The E-W cross section C-C' suggests that this anomaly extends eastward beneath the Deseado Massif but the absence of seismic stations in Argentina limits the interpretation of anomalies in this part of the model.

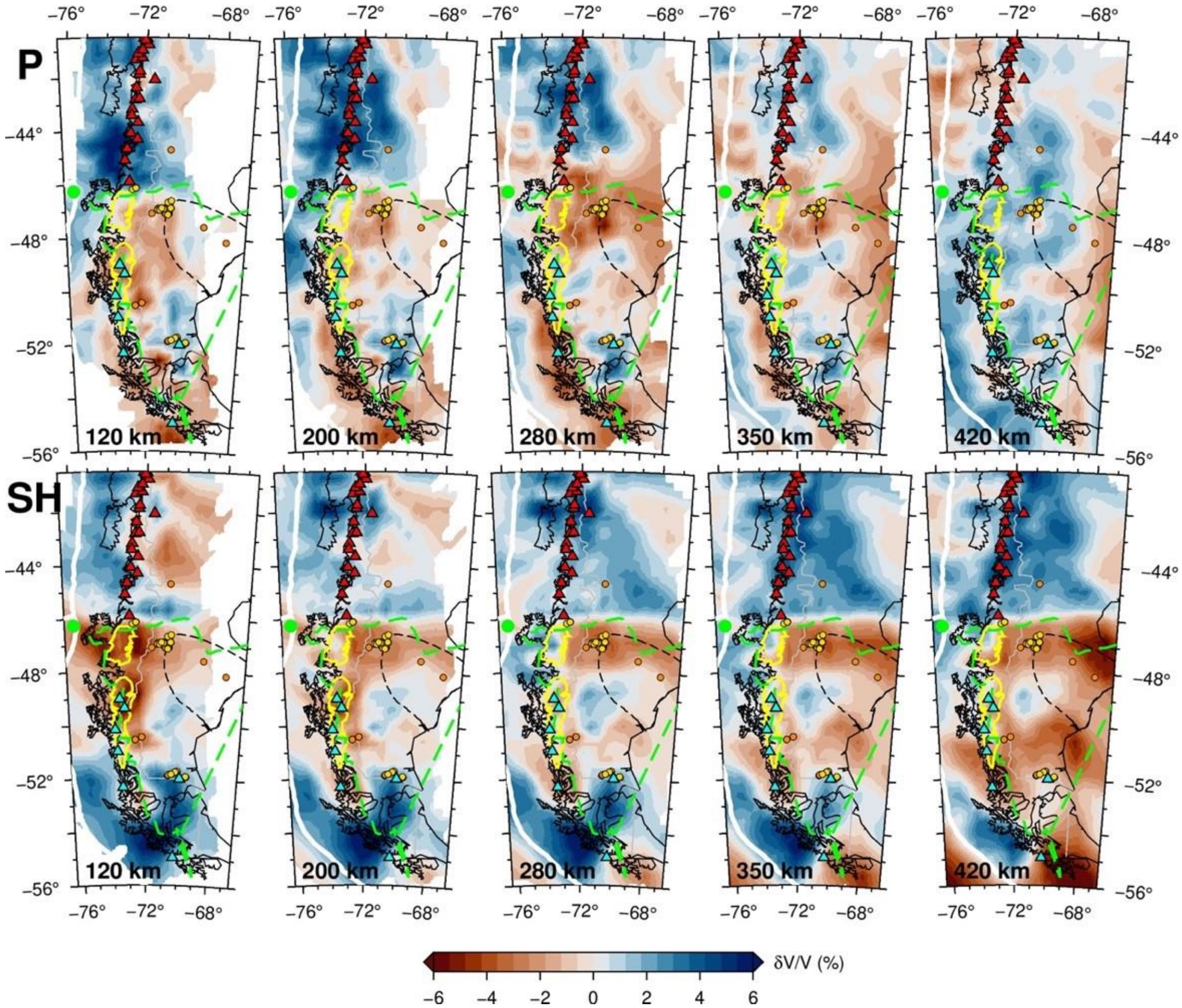

Figure 2. *Map views of $\delta V_P/V_P$ (top) and $\delta V_s/V_s$ (bottom) at 120km, 200km, 280km, 350 km and 420 km depth. The dashed black line represents the surface extension of the Deseado Massif, green dashed line represents the maximum extension of the slab window and yellow contours are the location of the Northern and Southern Patagonia Icefield.*

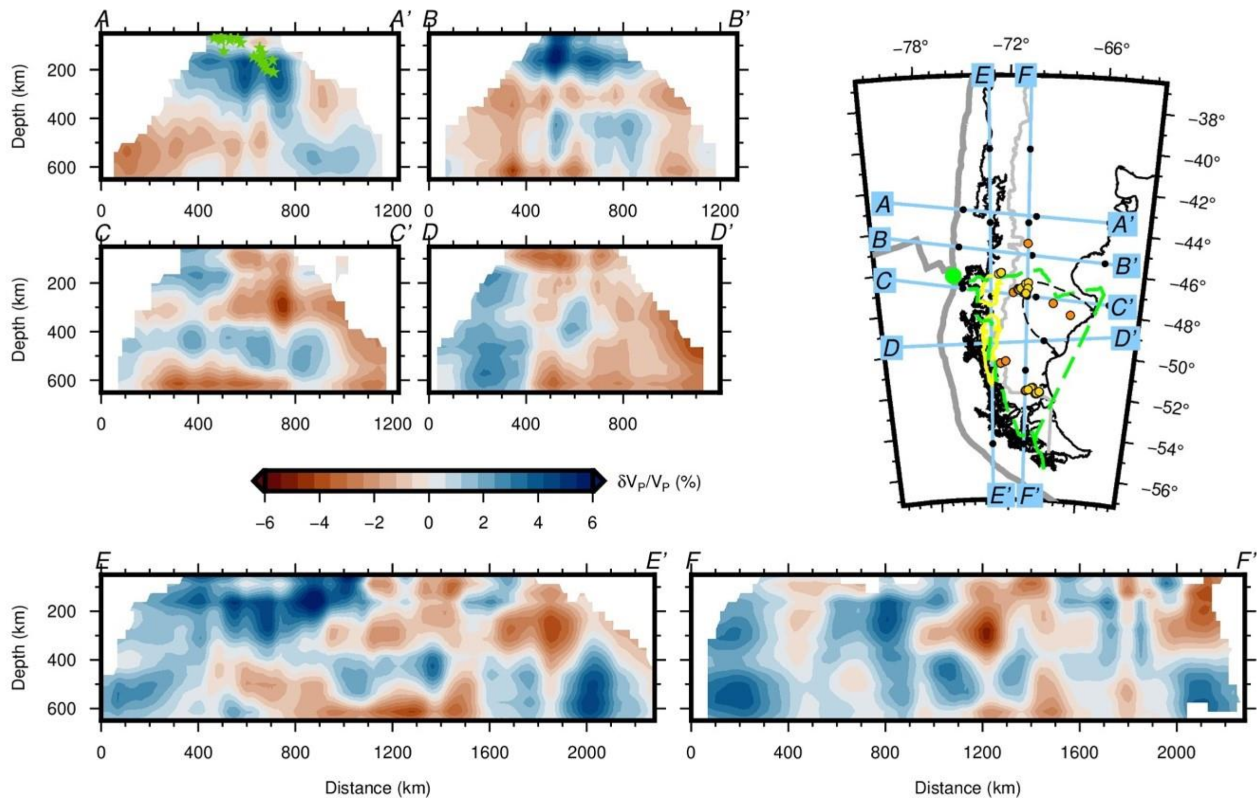

Figure 3. *Cross-sections across P model. Green stars represent earthquakes from the USGS catalogue. The tectonic features on the map follow the same conventions as in Figure 2.*

The northern part of the model region south to 46°S is dominated by fast seismic velocities likely associated with the subducting Nazca slab north of the triple junction. This is supported by the nice correlation with the Wadati-Benioff seismicity of the Nazca slab as seen in the E-W cross section A-A' in **Figure 3**. There is a clear contrast between a fast anomaly (3-4%) north of the CTJ and a slow velocity anomaly (3-4%) in the south. This contrast matches the present-day location of the slab edge from the paleo-reconstruction of the Patagonia slab window by Breisprecher and Thorkelson (2009). It matches also the boundary between a seismically active region to the North and a quiescent region to the South, where the absence of seismicity has been interpreted as the absence of a slab beneath south Patagonia. The P model shows clearly that the fast relative velocity anomalies to the north of the CTJ (green dot on **Figure 2**) extends eastward and down to 350 km. Due to the nature of S waves and their frequency content, the SH model does not constraint the shape of the fast anomaly as well.

3.2 Resolution tests

Checkerboard tests are used to evaluate the horizontal and vertical (**Figures S4-S6**) resolution in our datasets. For each synthetic model, we test different sizes of checkers and velocity perturbations (+/- 4%). We add Gaussian noise to the synthetic data prior to inversion with standard deviation of 0.04 s and 0.14 s for P and S waves, respectively, as estimated from the actual data. The inversion is performed using the same parameters and damping as used with real data. The checkerboard tests show good recovery down to 420 km for both P waves (**Figures S4 and S5**) and SH waves (**Figures S4 and S6**). Tests also show that the P-waves datasets can resolve anomalies that are about two times smaller than the SH waves datasets.

In addition, we test several scenarios concerning the geometry and extension of the Nazca slab (**Figures S7 and S8**), the lateral dimensions of the slow velocity perturbation beneath 46ºS and 48ºS (**Figures S9-S10**), and the presence of two distinct shallow and deeper slow velocity perturbations at the eastern edge of the study (**Figures S11**). Results show that a deeper and steep Nazca slab is better resolved down to 350km than a one segment slab for both P waves and SH waves (profile A-A' **Figures S8**). Tests on slow velocity perturbations between 46ºS and 48ºS (**Figures S10-S11**) show a good recovery in P and SH waves down to 420km and the ability to distinguish two distinct slow velocity anomalies in this region.

A comparison of our Vsh model with the Rayleigh wave Vsv model from Mark et al. (2022) (**Figures S12**) reveals relatively good agreement regarding the spatial distribution of fast and slow velocity anomalies across the slab window.

4. Discussion

4.1. Slow velocity mantle anomalies associated with backarc volcanism

The tomographic models indicate an association between slow velocity mantle anomalies and backarc volcanism (**Figure 2**). Southern Patagonia exhibits extensive back arc volcanism, likely associated with the northward migration of a slab window over the last 14 Myr [Breitsprecher and Thorkelson, 2009; Guillaume et al. 2010]. The source of this magmatism and its distinctive geochemical signatures, as well as its relationship to the slab window are not well understood [Gorring et al, 2003; Guest et al. 2024].

A strong slow velocity mantle anomaly is located beneath the Meseta del Lago Buenos Aires volcanic region, a locus of much of the Pleistocene backarc volcanism (**Figure 1**). Continental backarc volcanism is often assumed to occur by decompression melting from lithospheric extension, with shallow sources in the upper 100 km [Vasey et al, 2021]. In contrast, the slow velocity seismic anomaly beneath the Mesta del Lago Buenos Aires extends to depths of at least 400 km, indicating a deeper geodynamic process. The slow velocity anomalies also indicate a connection eastward towards the Atlantic coast at depths greater than 250 km, although resolution of the eastward extent is limited by the absence of seismic stations in eastern Argentine Patagonia. An eastward mantle connection is contrary to the typical assumption that mantle flows through the slab window from west to east [Russo et al, 2010; Hu et al, 2017; Ben Mansour et al, 2022]. However, it is consistent with the geochemical affinity of Meseta del Lago Buenos Aires volcanics with South Atlantic mantle plume sources [Soager et al, 2021; Guest et al, 2024].

The slow velocity anomalies beneath the backarc volcanoes are truncated to the east at depths shallower than 200 km by the Deseado Massif, which is incompletely imaged in the new models. The Deseado Massif is part of the Malvinas/Falkland Islands lithospheric block, a Precambrian terrain with strong mantle isotopic affinities with the Namaqua-Natal belt in south Africa [Marshall, 1994; Mundl et al., 2015; Schilling et al. 2017]. This lithospheric block is imaged farther to the south (52º-54º) with higher velocities in the upper 200 km, consistent with a thicker continental lithosphere (Mark et al, 2022, and **Figure 2**). The thicker lithosphere of the Deseado Massif may partially inhibit deep seated volcanism and localize the Meseta del Lago Buenos Aires volcanics along its western boundary. Petrological and geochemical analyses of xenoliths from the lithospheric mantle support the idea of mantle heterogeneity inherited from South Atlantic hotspots (Discovery-Shona-Bouvet) [Soager et al. 2021; Mallick et al. 2023; Jalowitzki et al. 2024] prior to the opening of the Atlantic Ocean. Contamination of melts by the lithospheric mantle with Atlantic geochemical affinity provides an alternative explanation for the Atlantic mantle plume signature in the Mesata del Lago Buenos Aires volcanics. Interestingly, the Pali Aike volcanic field (PAV, **Figure 1**), with extensive Pleistocene and Holocene activity, is located on the interior of the Malvinas/Falkland lithospheric block [Shilling et al, 2017]. The occurrence of these volcanics in a region of thicker lithosphere, as imaged by high seismic velocities in the upper 200 km (**Figure 2**), demonstrates that backarc volcanics can erupt even on the interior of older lithospheric blocks, possibly aided by fractures or gaps in the lithosphere that are too small to image in this study.
Between 50ºS and 52ºS, recent magmatism (< 2Myrs) has been recorded in the region of Cerro del Fraile. This volcanism is located on the top of a slow velocity anomaly in the shallow mantle (~ 120-160 km depth). Our models are consistent with geochemical analysis of mantle xenoliths

showing metasomatism of adakitic basalt and interpreted as the magmatic signature of the subduction of the current oceanic crust of the Antarctica Plate [Killian and Stern 2002; Stern 2004].

4.2. The geometry of the aseismic southernmost Nazca slab

The present-day geometry of the southern edge of the Nazca slab is poorly constrained. Intermediate-depth earthquake hypocenters extend to depths of about 150 km. The depth extent of seismicity is limited by warm slab temperatures due to the slab's young lithospheric age, which is 7 Ma at the 45ºS trench axis and increases northward [Tebbens et al, 1997]. Breitsprecher and Thorkelson (2009) estimated the position of the slab edge using plate reconstructions and assuming a poorly constrained shallow dip angle. The updated Nazca slab model by Portner et al. (2020) suggests an aseismic extension of the Nazca slab down to ~300 km in this region.

By combining data from all previous temporary seismic deployments with the new Chilean seismic network, we have a larger picture of the lateral and depth extent of the fast anomaly associated with the Nazca slab. The new P-wave tomographic model shows that the slab dip increases substantially at depths of 150-200 km, becoming steeply dipping at about 300 km depth (**Figure 3**). The lower resolution S-wave model shows high velocity slab material north of the triple junction but does not define the geometry as well. Due to the steep dip, the slab does not extend east of 69ºW within the upper 500 km, in contrast to the reconstruction in Breitsprecher and Thorkelson [2009], which shows the slab underlying the Atlantic coast at about 300 km depth. As a result of the steeper dip, as well as the northward motion of the triple junction over the past 12 Myrs, considerable Nazca slab material may lie in the transition zone beneath the Deseado Massif. Particularly at larger depths of around 300km, the high-velocity region associated with the slab moves eastwards when approaching the slab edge. This is reminiscent of the slab deformation by mantle toroidal flow around slab edges in the numerical experiments of Kiraly et al. (2017).

4.3. Rheological heterogeneity beneath the Patagonia icefields

The NPI and SPI are located on top of the slab window between 46°S and 51°S. Their ice mass has been responding very sensitively to changing climate and atmospheric circulation. Changes of the ice load drive a delayed bedrock deformation due to glacial isostatic adjustment (GIA). According to GIA models, a homogeneous viscoelastic half space responds to an ice-mass loss with an elliptical geographic pattern through a) uplift with a concentrical pattern about the barycenter of the lost ice mass, and b) a symmetrical horizontal extension radially away from that barycenter. The effect of a reduced mantle viscosity of that half space is that the new isostatic equilibrium be restored in a shorter time, implying larger deformation rates [Weerdesteijn et al, 2022]. A localized low-viscosity anomaly, eccentrical with respect to the ice-load symmetry, enhances the horizontal deformation rates in the direction from the load center towards that anomaly [Kaufmann et al. 2005; Klemann et al. 2007].

GNSS observations in the SPI area reveal intense uplift, with rates reaching 4 cm/year in the northern part of the SPI, consistent with a GIA model assuming a homogeneous half space [Dietrich et al. 2010; Lange et al. 2014; Richter et al. 2016]. However, the observed uplift rates are smaller than the modelled ones in the southern part of SPI, south of 50°S, and larger than predicted by the simple GIA model in the northern part of SPI. Furthermore, the observed horizontal deformation is not symmetric about the N-S axis of the icefields, with a stronger eastward extension to the east of the load center compared to the westward extension to the west [Richter et al. 2016].

Seismic models offer an opportunity to estimate lateral changes in mantle viscosity, since both velocity and viscosity are largely controlled by temperature [Ivins et al, 2023]. However, recent detailed Rayleigh wave seismic structure models for southern Patagonia are limited to depths of about 150 km [Mark et al, 2022]. The new models presented here can extend our understanding of rheological variations in this region to deeper depths and can thus provide explanations for the discrepancies between the observed deformation patterns and those predicted by GIA models based on homogenous rheology. The new seismic models show a north-south dichotomy, with the NPI and the northern part of the SPI underlain by low velocities over most of the upper 300 km, and the southern SPI underlain by faster velocities (**Figure 4**). This is in agreement with the Rayleigh wave results of Mark et al (2022) but shows that this dichotomy extends deeper than previously imaged. This dichotomy explains the increase from South to North of the observed uplift compared to the homogeneous model, as a consequence of the decrease in mantle viscosity in that direction, provided that the GIA model assumes a correct ice-load distribution. The new seismic models also show very strong low-velocity anomalies in the backarc near both icefields at 120 km depth, and for the NPI at deeper depths. These velocities anomalies can explain the asymmetric horizontal deformation pattern as a modulation of the GIA-driven radial extension by a localized, low viscosity just east of the ice-load axis.

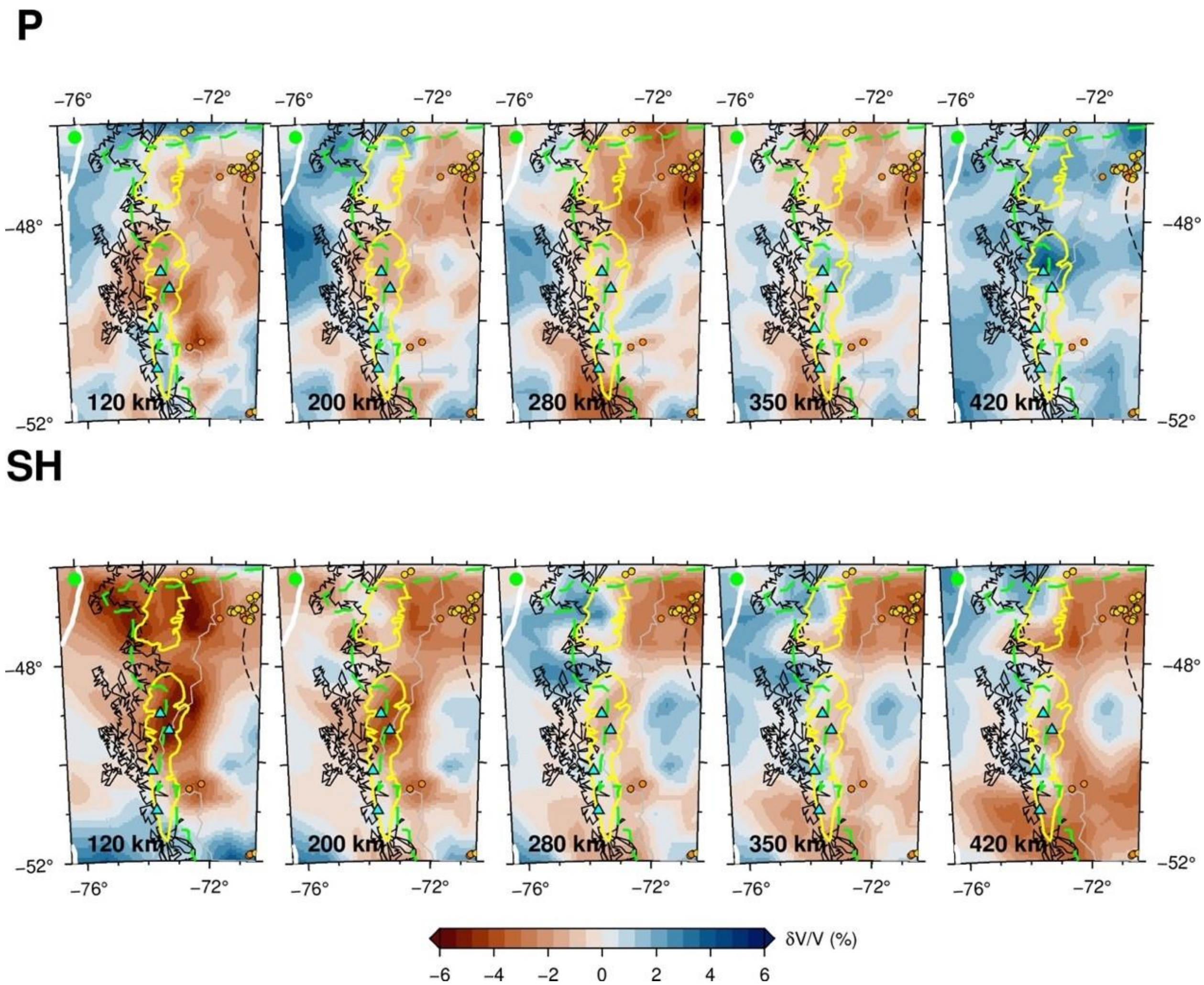

Figure 4. *Enlarged view of the tomographic model in the vicinity of the Patagonia icefields at depths of 120km, 200km, 280km, 350 km and 420km. The tectonic features on the map follow the same conventions as in Figure 2.*

5. Conclusions

We present new regional scale finite-frequency P and SH waves tomography models of southern Patagonia, using a wavelet-based multiscale parametrization that increases the recovery of amplitude and location of seismic velocity anomalies. The new models show a fast velocity anomaly north of the CTJ associated with the aseismic southernmost Nazca slab and a slow velocity anomaly in southern Patagonia related to the present-day location of the slab window. We find that slow velocity anomalies beneath back-arc volcanoes in the northern part of the slab window extend to depths of ~ 400 km, suggesting that the volcanism does not result from shallow mantle processes but is instead linked to upwelling thermal anomalies from deeper regions of the upper mantle. We also show that the southern edge of the Nazca slab extends downward to at least 350km and has a steeper slope than in previous models. Finally, the new tomographic models illuminate the structural heterogeneity beneath the Patagonia icefields. Our models show a north-south dichotomy, with the NPI and the northern part of the SPI underlain by low velocities and the southern SPI underlain by faster velocities down to depths of about 300 km. Very strong low-velocity anomalies in the backarc near both icefields support the idea of that more complex rheological models may be able to explain geodetic observations in Patagonia.

Open Research

Data used in this study are from the temporary seismic networks GUANACO, SEARCH and CRSP temporary seismic networks respectively under the network code 1P, YJ and XJ. Permanents stations from the Chile network and Global network are under the network code C/C1 and GT. Data can be obtained from the EarthScope data center:

1P : https://www.fdsn.org/networks/detail/1P_2018/,

XJ: https://www.fdsn.org/networks/detail/XJ_2004/,

YJ: https://www.fdsn.org/networks/detail/YJ_2004/,

C: https://www.fdsn.org/networks/detail/C/,

C1: https://www.fdsn.org/networks/detail/C1/

GT: https://www.fdsn.org/networks/detail/GT/).

Tomography models showed and discussed in this study can be found in Ben-Mansour and Maupin (2025).

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Figures were produced using GMT software [Wessel et al., 2013] using perceptually uniform colour maps generated by Fabio Crameri [Crameri, 2018].

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Supplementary Materials: Deep source regions for Patagonia backarc volcanism imaged by finite frequency body wave tomography

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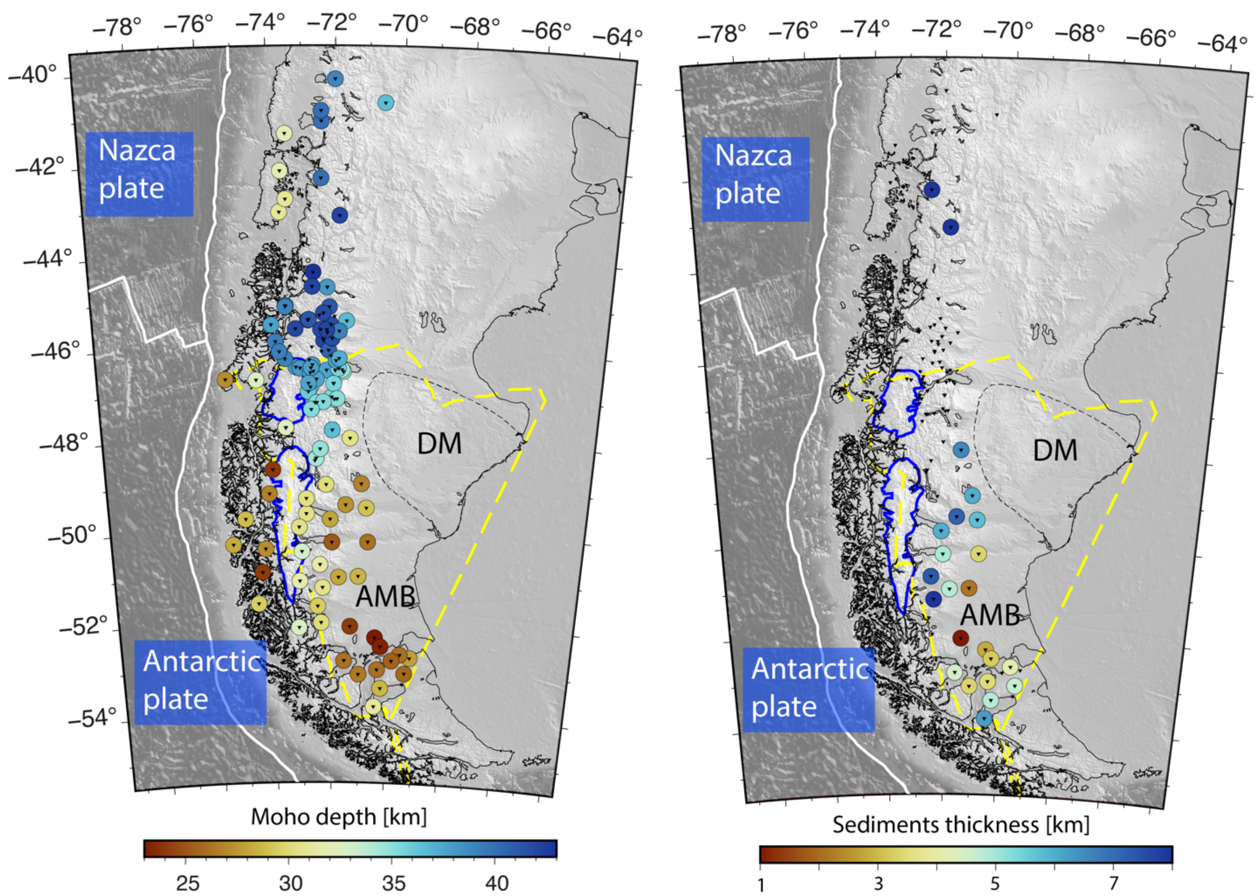


Figure S1a. Moho depth (left) and sediments thickness (right) from Mark et al. (2022) used for the crustal correction. DM: Deseado Massif, AMB: Austral Magallanes Basin. Yellow dashed lines represent the maximum extension of the slab window and blue contour the Patagonia icefield. Black triangles are the locations of seismic stations used in this study.

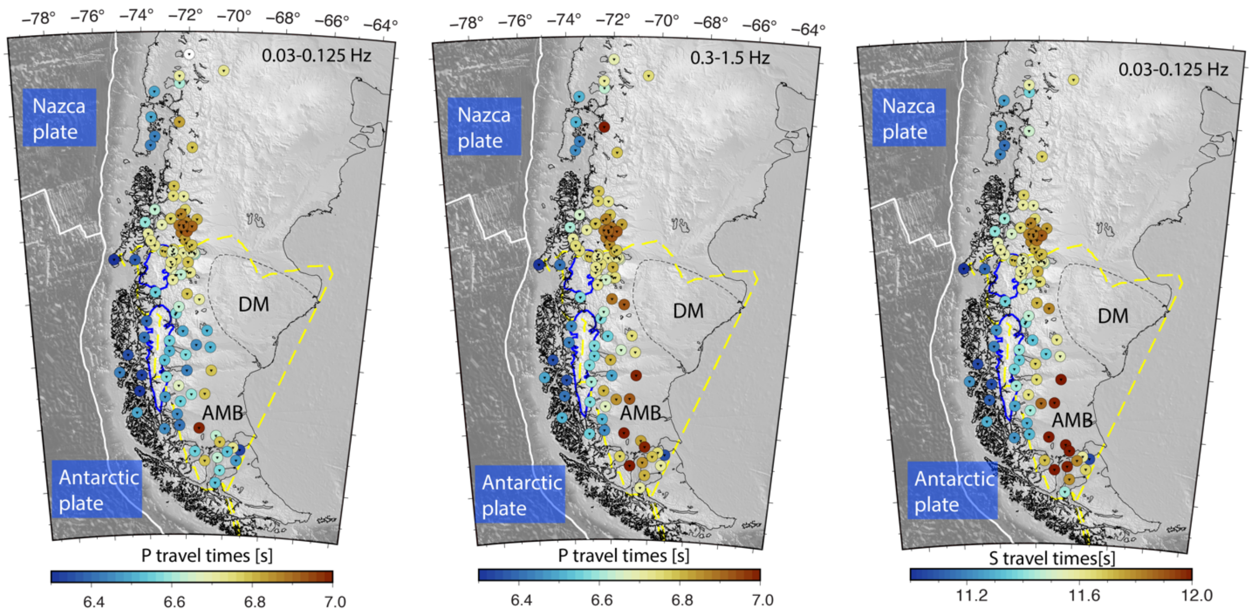


Figure S1b. Travel times from 50km below sea-level to the surface, including topography, for, from left to right, low-frequency P waves, high-frequency P waves and SH waves. Note the frequency dependence of the P-wave travel times in regions with thick sediments where low frequencies display reduced travel times.

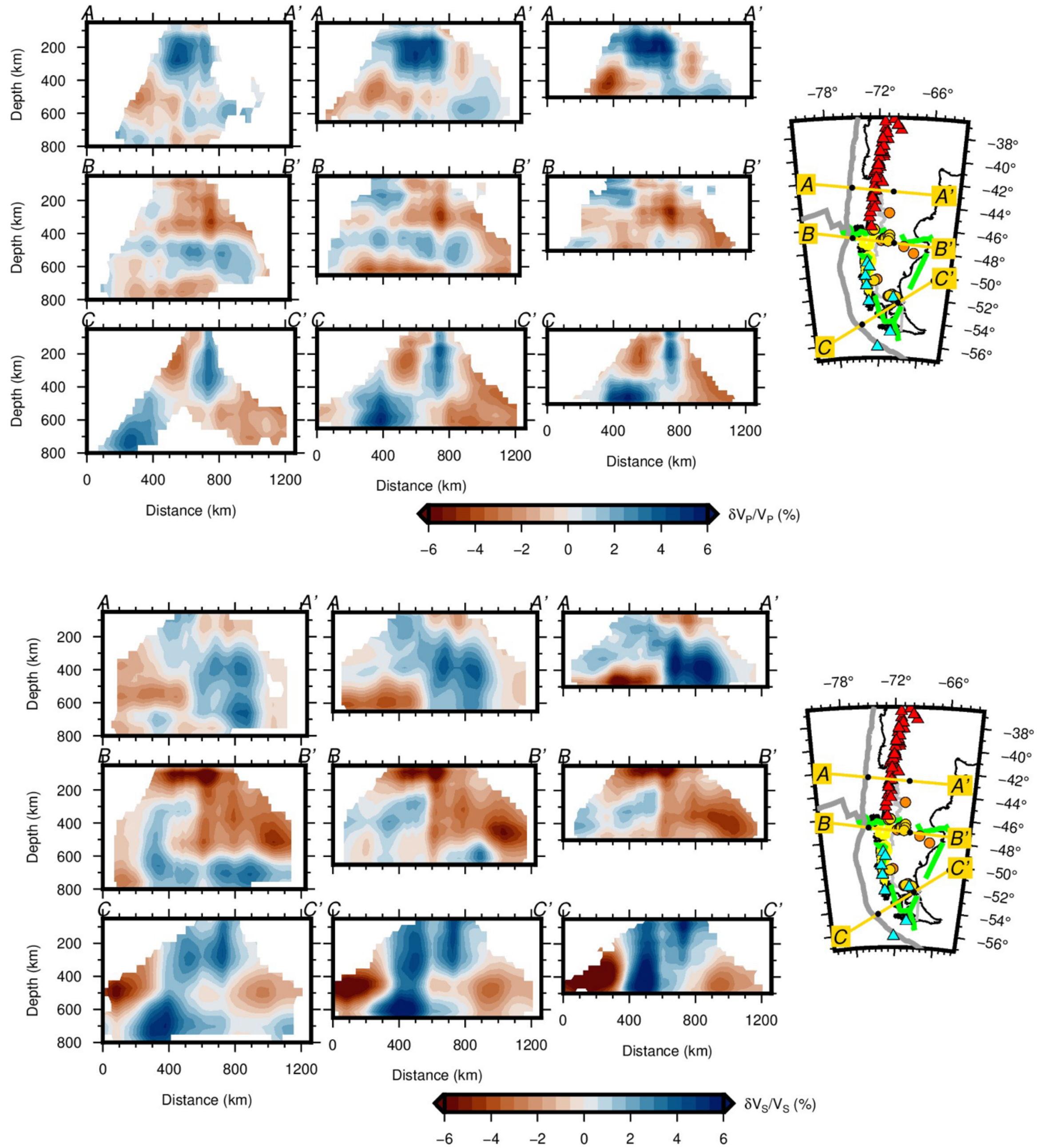


Figure S2. Test of the effect of the depth extent of the model. Vertical cross sections in P and SH models extending from left to right down to 800, 650 and 500km depth. *The tectonic features on the map follow the same conventions as in Figure 2 in the manuscript.*

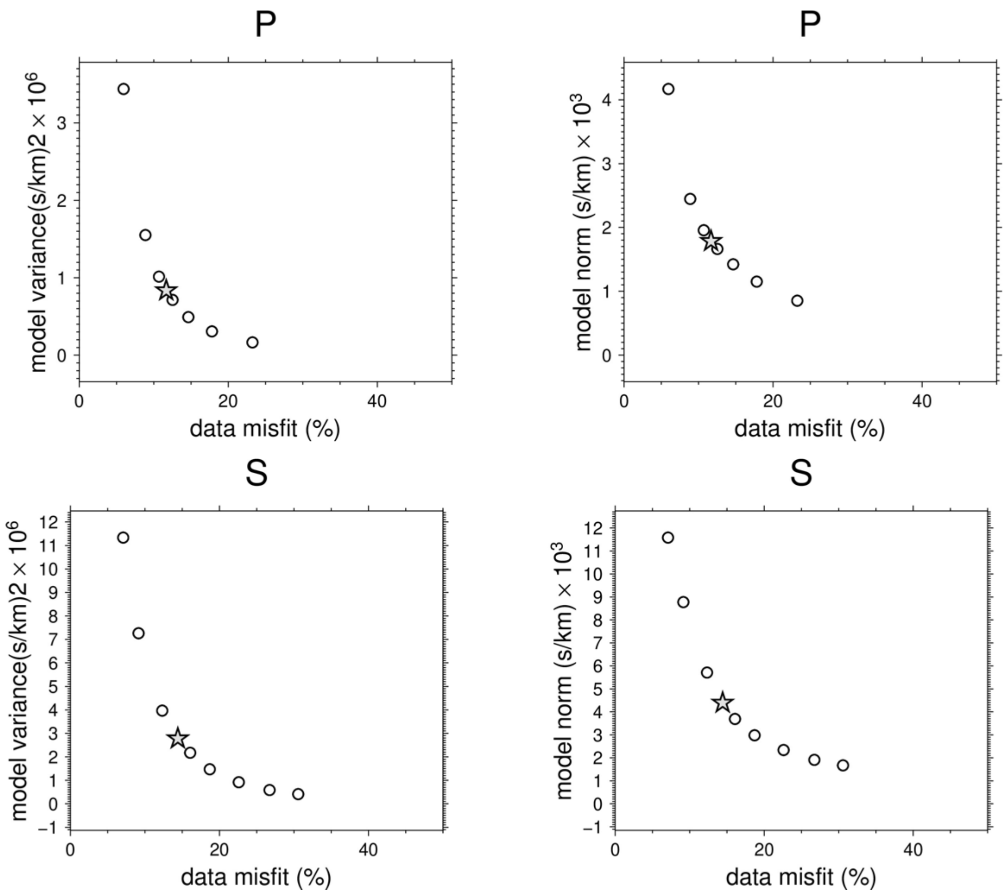


Figure S3. Trade-off between model variance and data misfit (left) and Trade-off between model norm and data misfit. The star represents the preferred solutions shown in the manuscript for P and SH.

Checkerboard tests

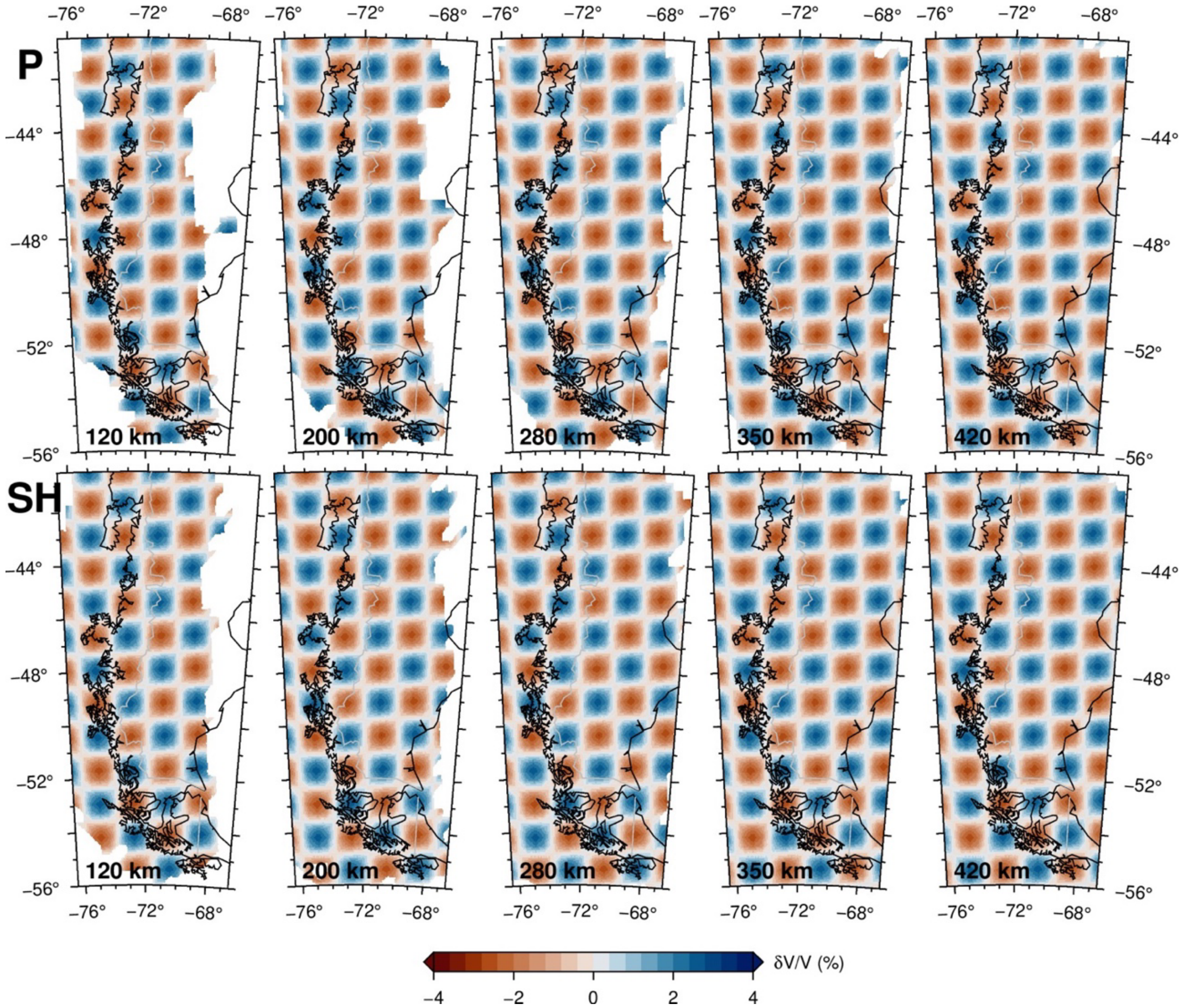


Figure S4a. Synthetic input model for a checkerboard test with cells of 1.25° in horizontal dimensions, 150km in vertical dimension, δlnV_P = ±4% (top), and δlnV_S = ±4 % (bottom). Maps at 120, 200, 280, 350 and 420 km depth.

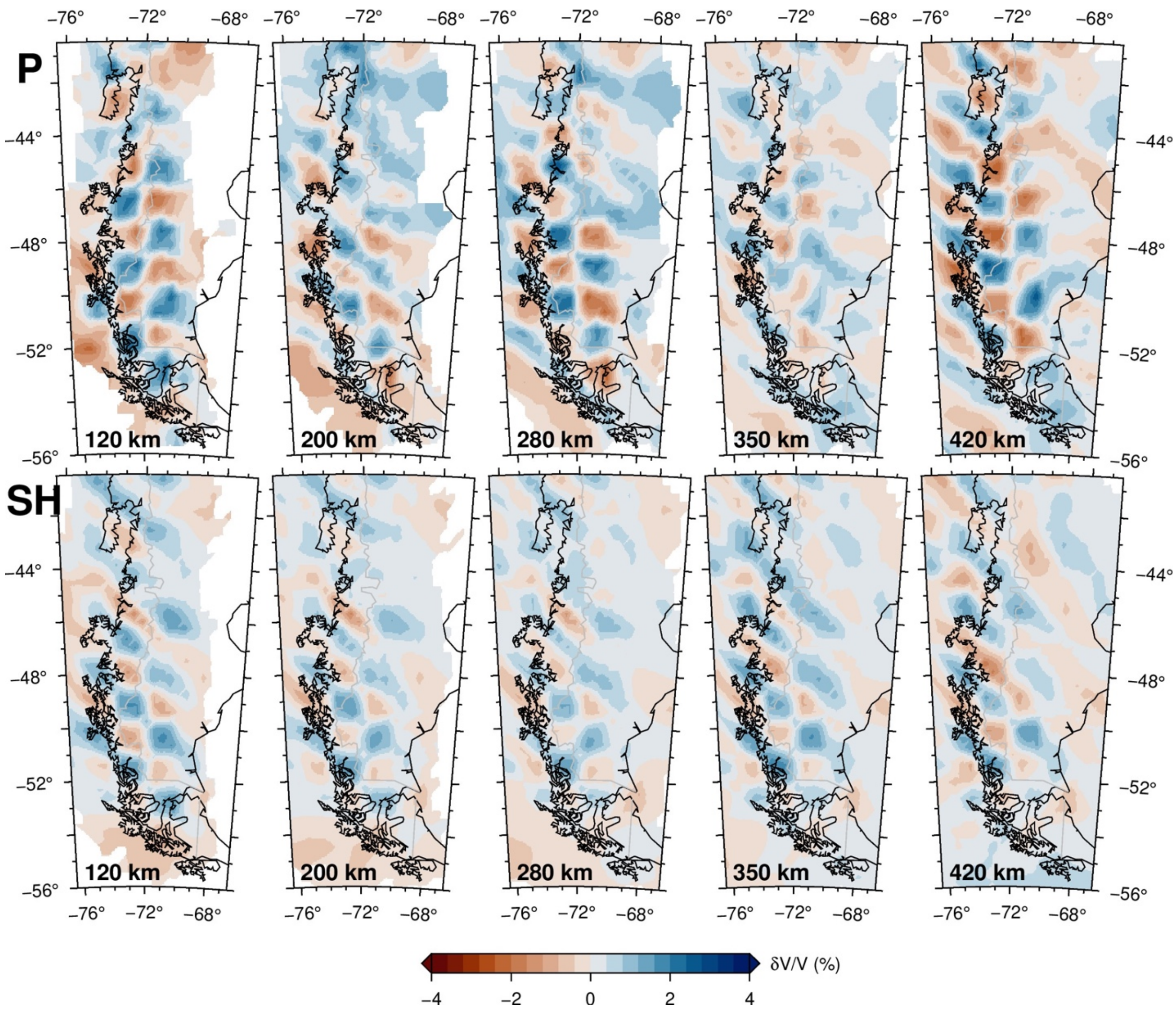
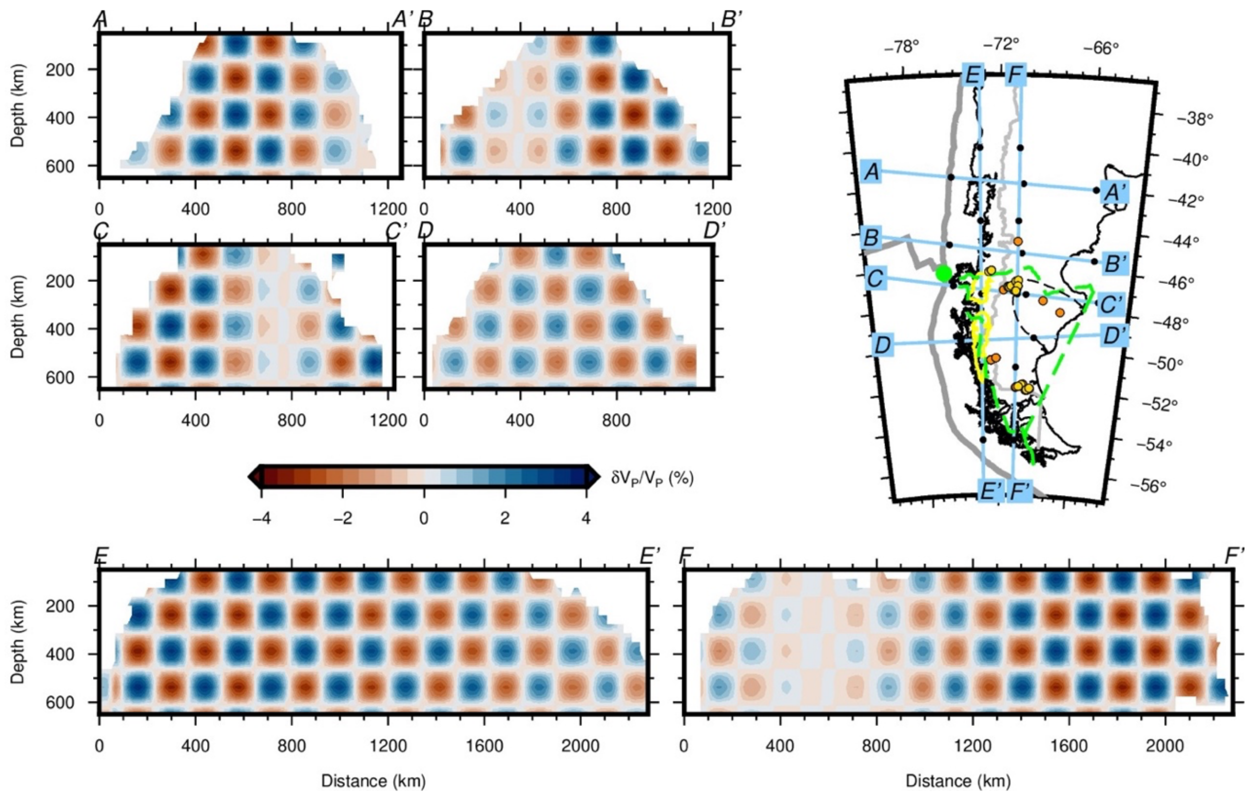


Figure S4b. Recovery for checkerboard tests presented in Figure S4a



Figures S5a. Vertical cross-sections in the checkerboard synthetic model for P waves presented in Figure S4a. Map view of study area with the location of different profiles. *The tectonic features on the map follow the same conventions as in Figure 2 in the manuscript.*

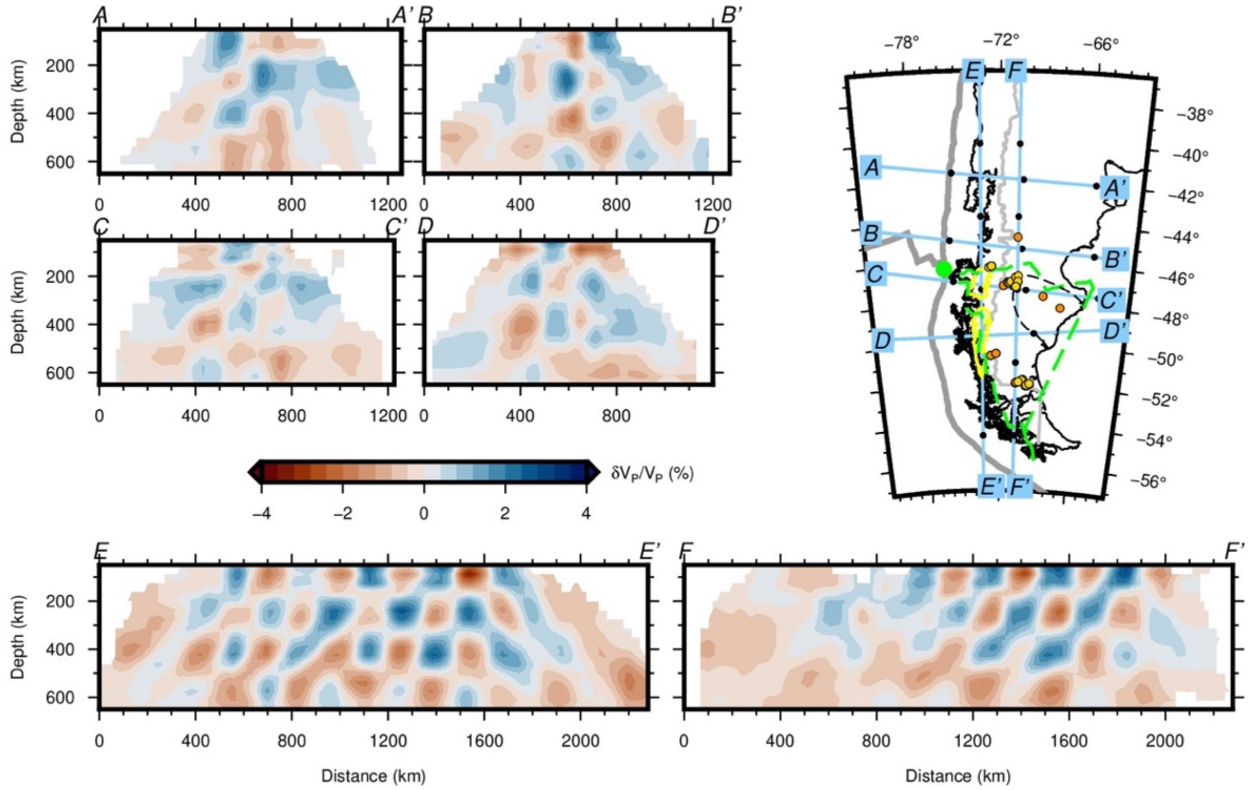
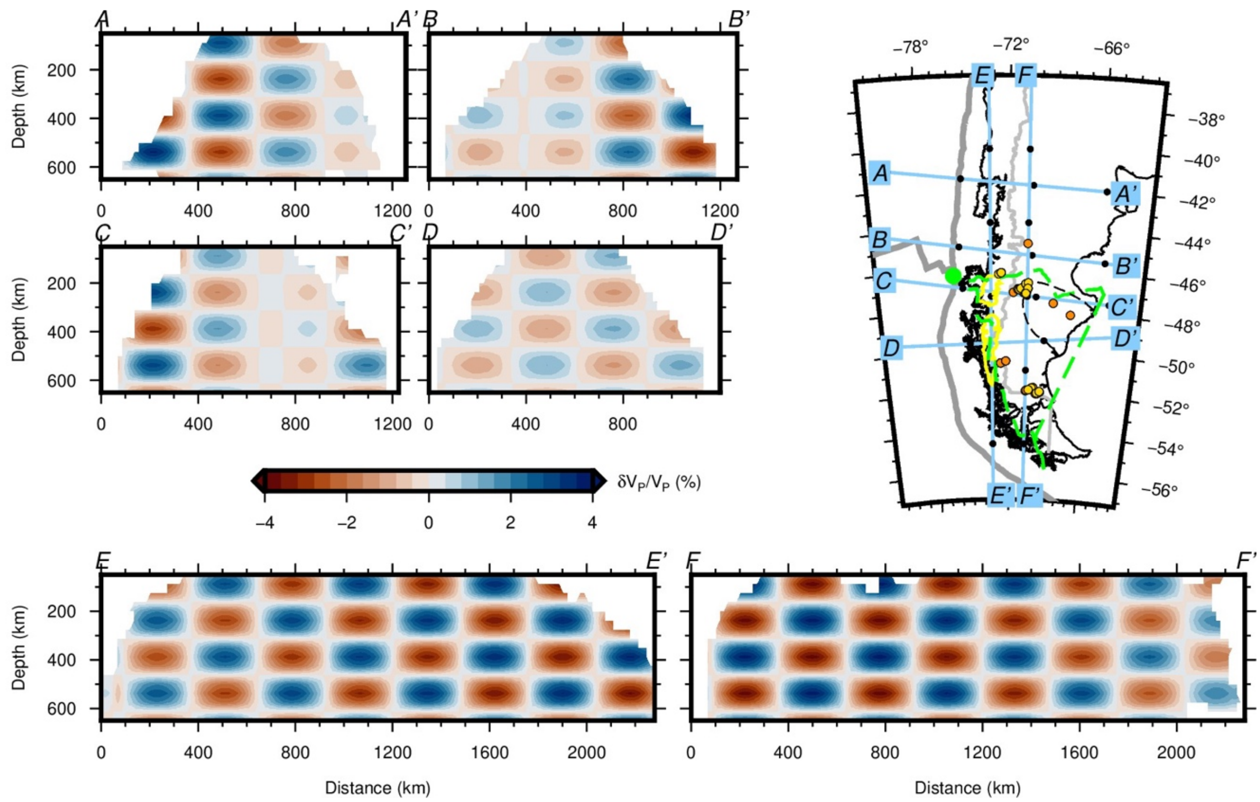


Figure S5b. Recovery for checkerboard test presented in Figure S5a



Figures S6a. The same as Figure 5a for horizontal dimensions of 2.5°.

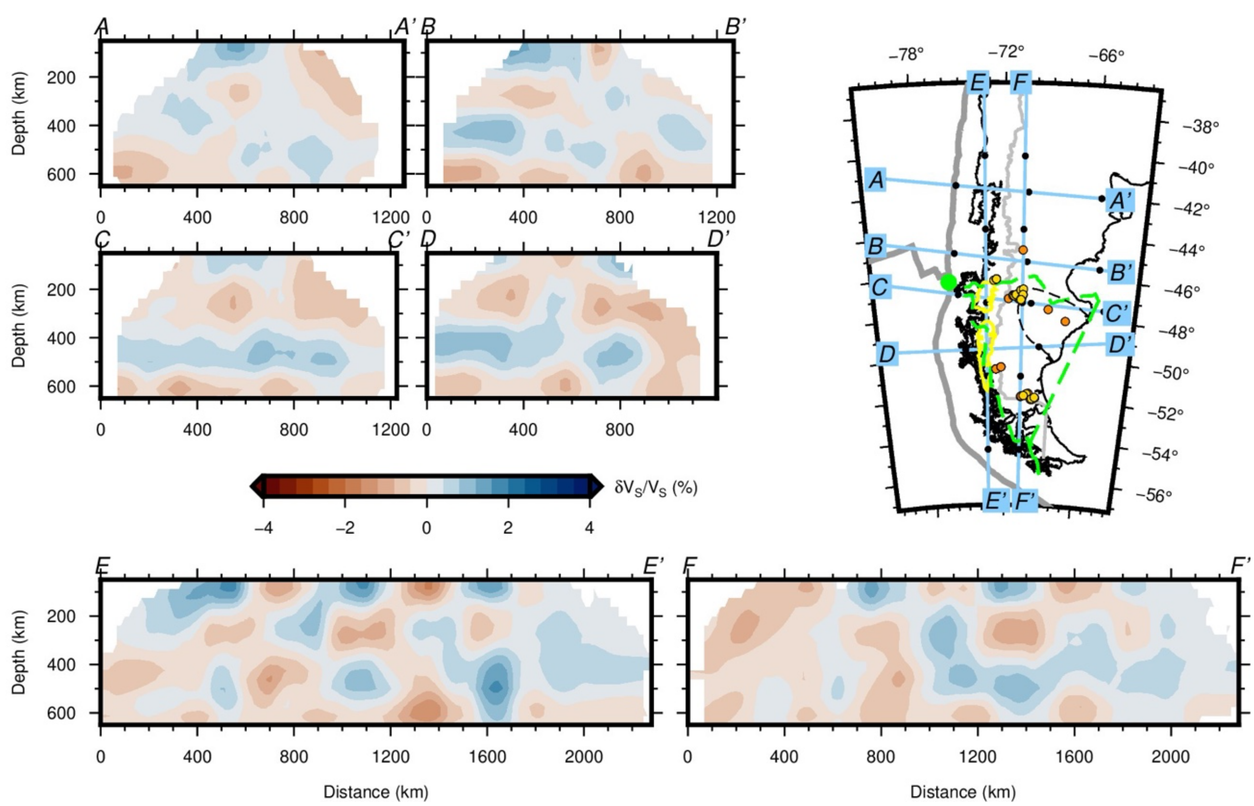


Figure S6b. Recovery for checkerboard test presented in Figure S6a

Structural tests

Fast velocity perturbation/Nazca slab

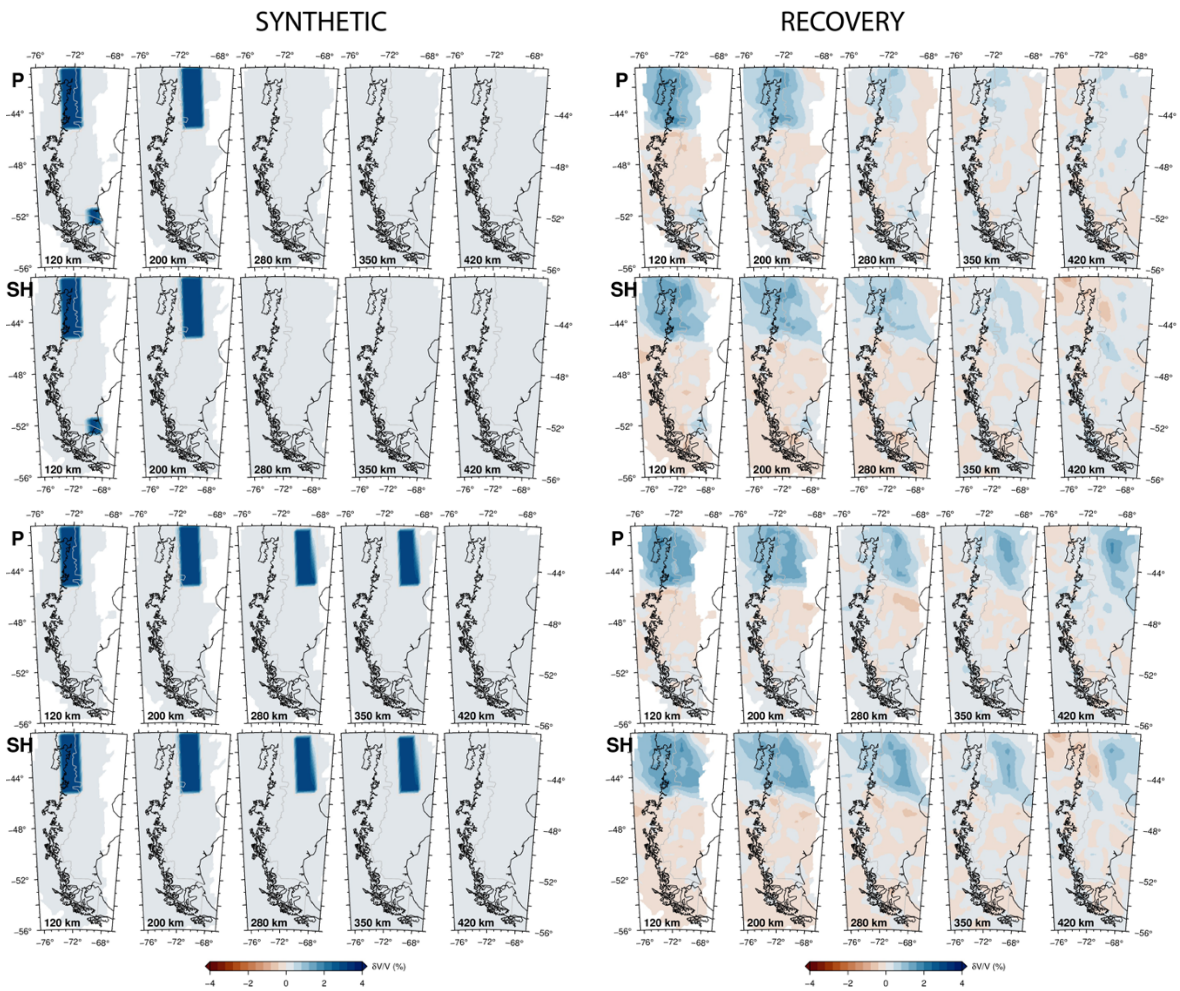


Figure S7. Synthetic and recovery results for P and SH waves associated with a fast velocity perturbation (+3%) between 40°S and 46°S simulating a 100km thick slab penetrating down to 200 km and 350km respectively.

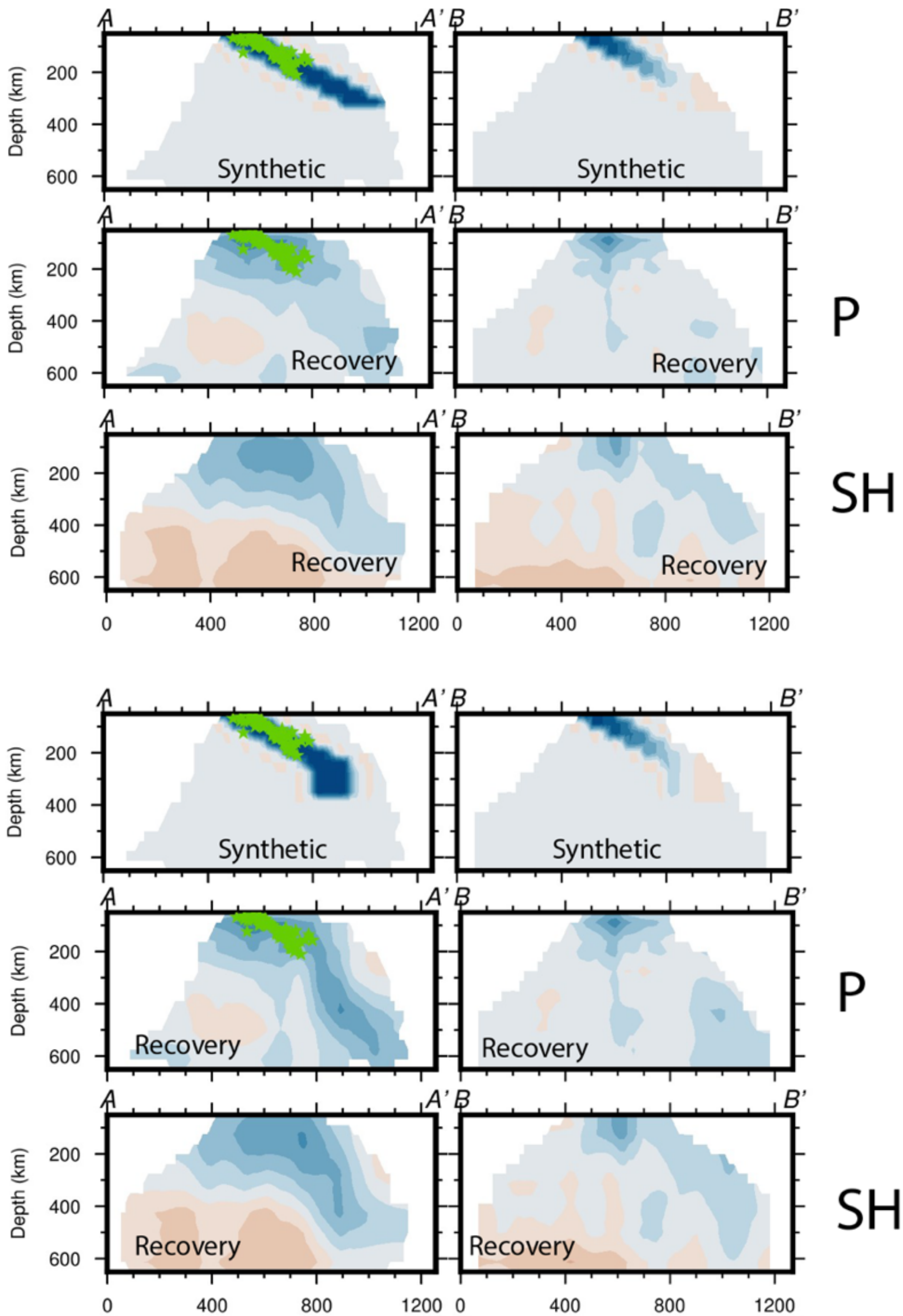


Figure S8. Synthetic and recovery results for P and SH waves in cross section associated with a 100km thick fast velocity perturbation (+3%) representing two possible options for the depth extension of the Nazca slab. Green stars represent earthquake from USGS catalogue.

Slow velocity perturbation/Slab window

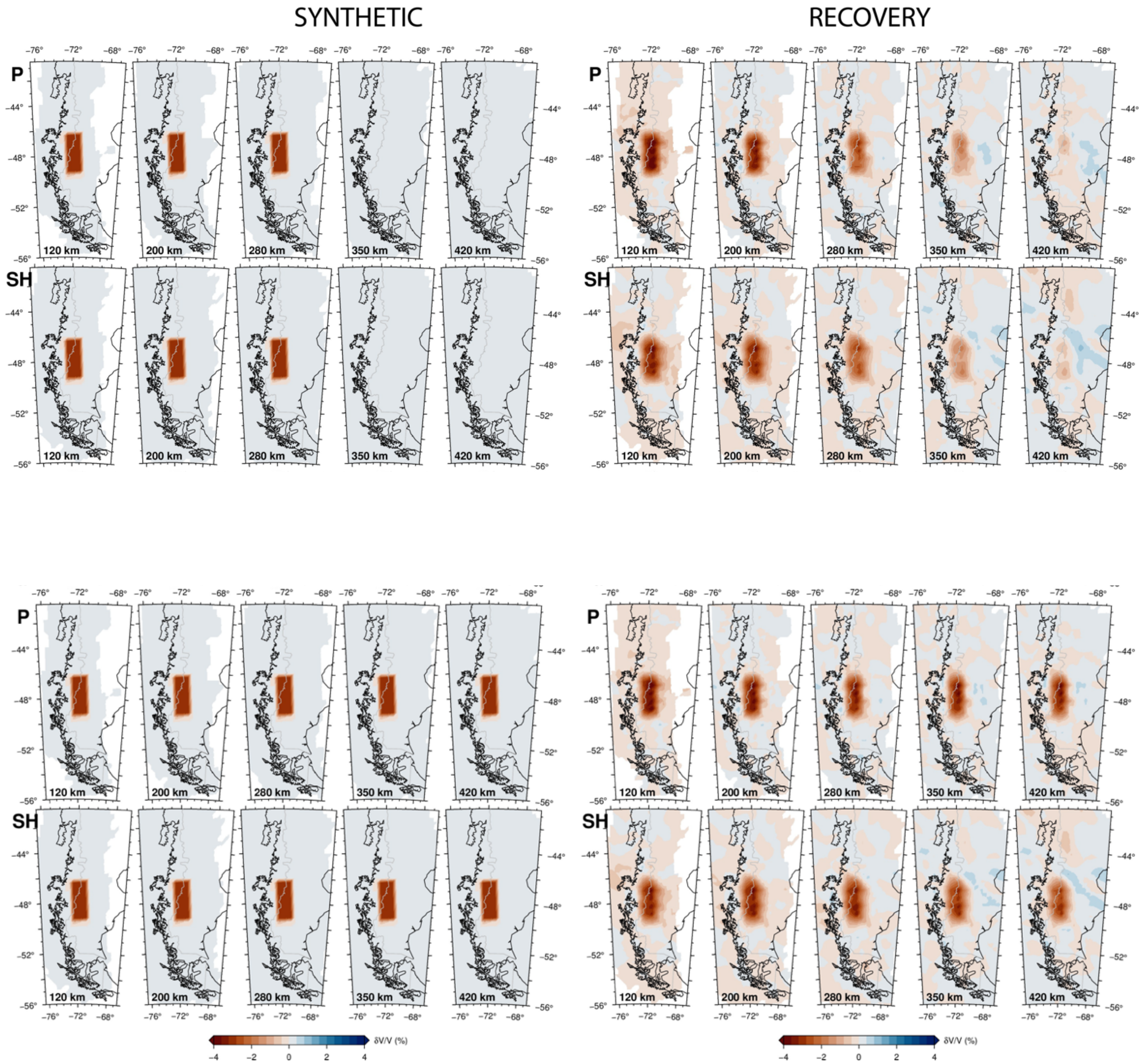


Figure S9. Synthetic and recovery results for P and SH waves associated with a low velocity perturbation (-3%) between 46°S and 49°S down to 280km and 420km depth respectively.

Two distinct low velocity perturbations/Slab window and one shallow fast velocity perturbation

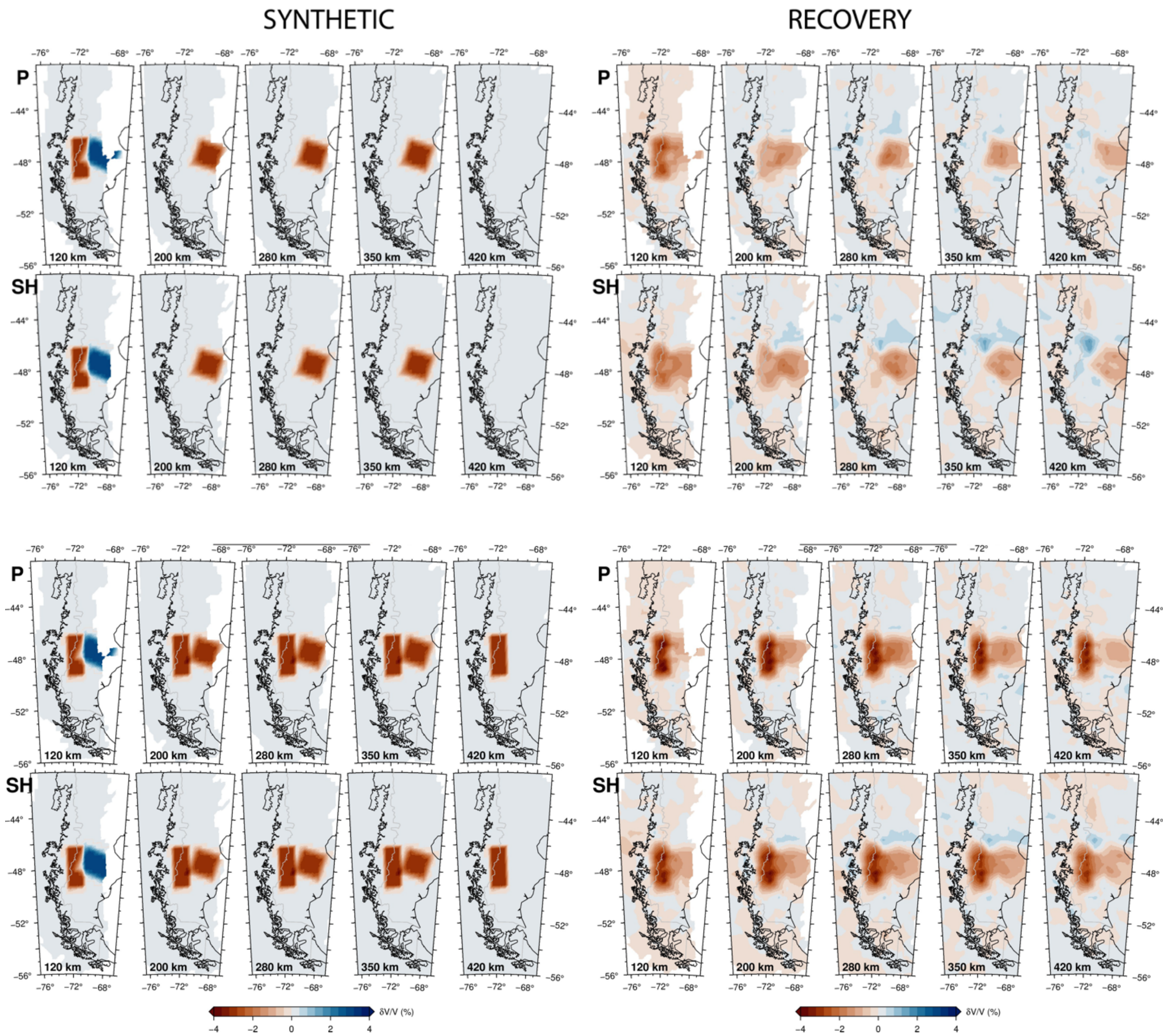


Figure S10. The same as Figure S9 with addition to the east of the previous anomaly of a deep low velocity perturbation (-3%) and a shallow fast one (+3%).

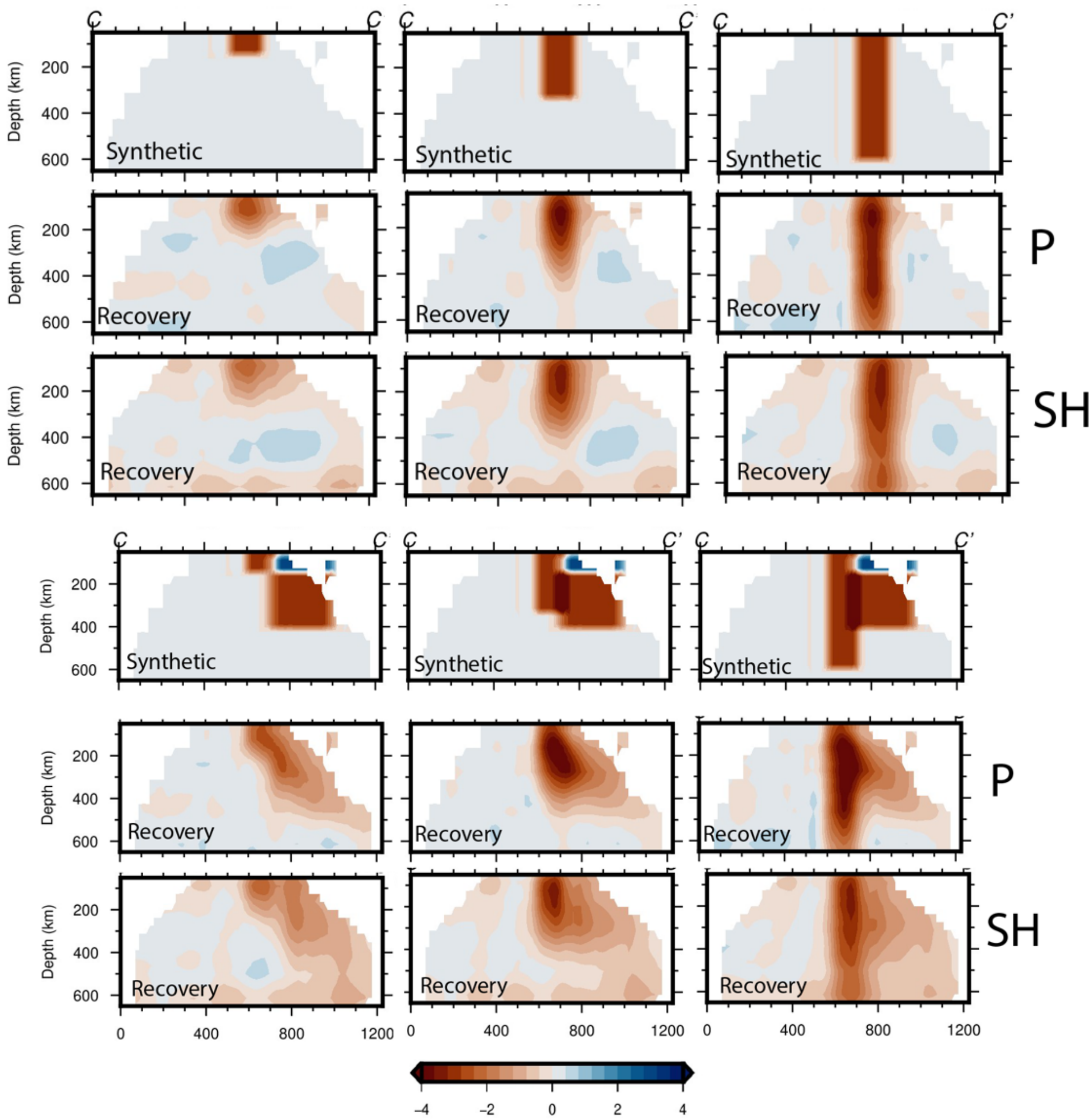


Figure S11. Vertical cross sections for the resolution tests presented in Figure S10.

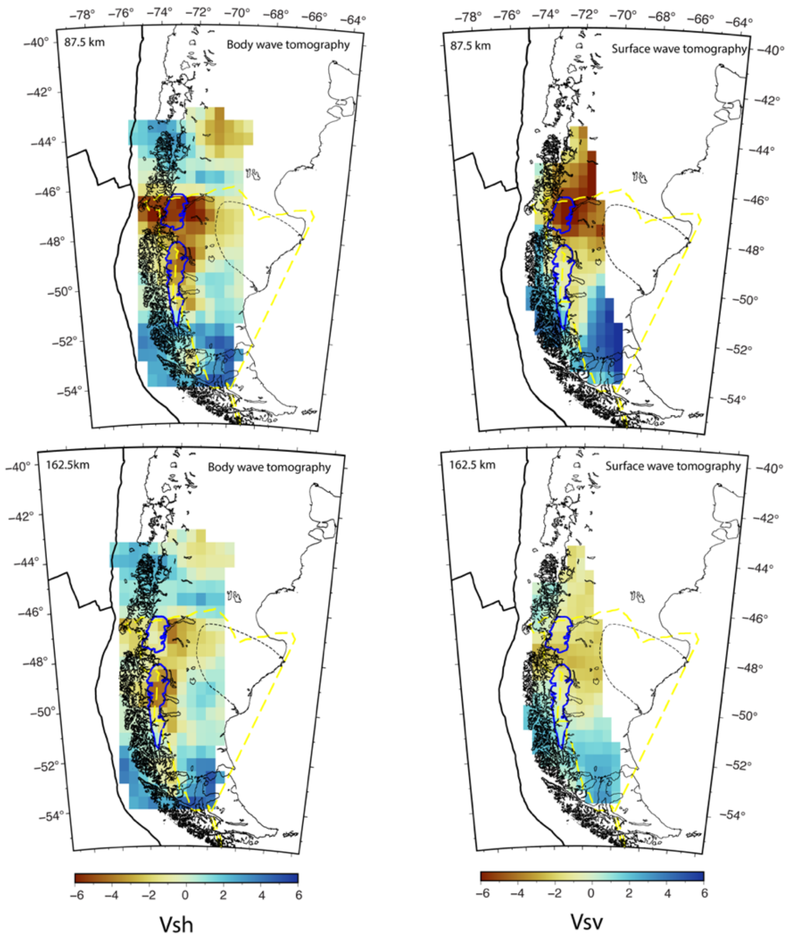


Figure SM12. Comparison of shear wave structure at 87.5km (top) and 162.5km(bottom) derived from body wave tomography (this study) and surface wave tomography (Mark et al. 2022). *The tectonic features on the map follow the same conventions as in Figure 2 in the manuscript.*