

P-wave Velocity Structures Under South America From Multi-Frequency Tomography

by

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Declaration

I declare that this thesis is entirely my own work, and except where otherwise stated, describes my own research.

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Abstract

This thesis presents P-wave tomographic images of mantle structures under South America and offers tectonic interpretations for the recovered features. To build this inversion, source time functions, a requirement for computing suitably realistic synthetic seismograms, were estimated from broadband waveforms of 3505 teleseismic earthquakes of magnitudes greater than 5.6, recorded between 1990 and 2016 by regional and global networks. Almost 5M traveltimes anomalies were calculated using cross-correlation of synthetic and observed seismograms in eight frequency bands, with centre periods ranging from 30.0 to 2.7 seconds. Sensitivity kernels were calculated using finite-frequency theory. In order to better illuminate weakly sampled mantle regions, integrated traveltimes picks from the International Seismological Centre's ISC-EHB catalogue were also included.

Two main features are inferred from the tomographic images under South America. First, extensive lower-mantle fast anomalies under north-western region of South America may indicate ancient intra-oceanic subduction settings, which might have occurred to the west of the western margin of the continent. Secondly, high velocity provinces beneath western margin of South America, which appears only in the upper mantle, is interpreted to imply a younger subduction history. This fast anomalies distribution is associated to the Andean segmentation which has been archived in the geological records.

Extended Abstract

Two main features of the Andean mountain belt are difficult to reconcile with the generally accepted history of a long-lived continental arc. The first is widespread Late Cretaceous deformation which marks the central segment of the Andes. The second is widespread extension which is extreme at both northern and southern ends of the Andes.

This dissertation is an attempt to investigate possible signatures of these features on the mantle structures. To understand subduction history over the past 180 million years, I consider the obtained tomographic images, especially the fast anomalies therein jointly with tectonic reconstructions, assuming the slabs sink vertically.

The following episodes are inferred from the geometry of the slab under vertical sinking rate of $\sim 10 \text{ mmyr}^{-1}$. Late Jurassic to Early Cretaceous history is marked with the presence of two intra-oceanic subduction zones in the proto-Pacific basin, which are associated with the subductions of Farallon Ocean, and Mezcalera (proto-Caribbean) Ocean. The two subduction systems co-existed until west-drifting South America overrode them. The second episode is associated with the opening of the Caribbean Oceanic Plateau $\sim 120 \text{ Ma}$, which is inferred from a sudden change in the geometry of the slabs between ~ 1200 and $\sim 800 \text{ km}$ depths. This slab rearrangement is also interpreted as the collision of the island arcs, produced by the late Cretaceous double subduction zone of Farallon-Mezcalera, with north-western margin of South America. This event is followed by a second phase of the collision, during which island arcs, formed by subduction of Farallon and Mezcalera along the Caribbean plateau, were jammed along the western Cordillera of the North Andes at $\sim 70 \text{ Ma}$. Collision of the continent with the intra-oceanic arcs is accompanied by the initiation of the east-dipping Farallon subduction along the western margin of South America, while Mezcalera ocean is closed in the northern latitudes. The subduction of the Farallon under Central Andes is characterised with a noticeable curvature below $\sim 800 \text{ km}$ depth, which is linked to the formation of the Bolivian orocline. These interpretations imply that the crustal growth of the northern segment of the Andes resulted from the collision of the intra-oceanic arcs, while the appearance of the subduction under Central Andes during late Cretaceous, might explain why the deformation of the Central Andes initiated then.

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List of Abbreviations

Ar		Argentina
Br		Brazil
Ca		Caribbean
CCC		Cross-Correlation Coefficient
Ch		Chile
Co		Colombia
COP		Caribbean Ocean Plateau
Ec		Ecuador
Fa		Farallon
GAC		Great Arc of Caribbean
GSN		Global Seismographic Network
Gp		Galápagos
IRIS		Incorporated Research Institute for Seismology
ISC		International Seismological Centre
JFR		Juan Fernandez Ridge
LIP		Large Igneous Province
MEZ		Mezcalera
P		Piñon
Pr		Prue
Rd		Ridge
SB		Serranía de Baudó
STF		Source Time Function
Ve		Venezuela

1

Introduction

This chapter is structured into four main sections. In section 1.1, the finite-frequency theory, the underlying method in this study, is introduced. To understand why a new tomography is still required in South America, section 1.2 provides a brief overview of the geological setting in the Andean mountain belt. I briefly outline the main questions in the investigation of the tectonic history of the Andes. This is followed by section 1.3, which explains previous contributions of tomographic studies to address some of these questions. In the final section, the motivation behind this study, and the contents of the following chapters are outlined.

1.1 Finite-frequency tomography

Seismic tomography is an inverse problem, that extracts information from the observed data to recover the interior structure of the Earth. More properly, if the Earth structure (m) is related to the data (d) by $g(m) = d$, seismic tomography solves this equation by finding m that fits the data (Rawlinson et al., 2010). Here g is a physical relationship connecting data and the unknown structure of the Earth. Traditionally, this relationship has been calculated based on the asymptotic high-frequency ($\omega \gg 1$, where ω is the frequency) solution of the wave equation, resulting in the geometrical ray representation of the wave. In the ray theory, it is assumed

that the length-scale of the variations of the medium to be much larger than 1) the wavelength of the waves, and 2) the width of the first Fresnel zone (Snieder and Lomax, 1996). This leads to the conservation of the propagated energy only along the infinitesimally narrow rays.

Despite pioneering successes, this assumption suffers from an important disadvantage, where it cannot sense high spatial wavenumber variations of the real Earth, due to its infinite frequency approximation (Liu et al., 2009). In other words, the technique is not able to resolve the small-scale structures of the media by the waves, often measured in hundreds of kilometres.

In practice, the wave propagated from a seismic source to the station is influenced by the medium outside the ray path. One way to take the scattering perturbations into account is to solve the equation of motion numerically without any approximation. However, this approach is not affordable. Another theory, is the Born approximation, in which first order perturbation is considered in solving the equation of motion. In other words, the total wave-field in a lateral heterogeneous medium is given as the summation of secondary wave-fields, produced by the scattering points (heterogeneities). Considering the first order scattering effects of off-ray heterogeneities has led to the development of fat rays based on the wave equation (e.g. Snieder and Lomax, 1996). The fat rays, also called Frechet kernels, reflects the finite-frequency sensitivity of the travel-time, and amplitude of the recorded signal to the perturbation of the media through which the wave has travelled. Due to the volumetric shape, these kernels sense larger regions of the Earth than the geometrical ray. Marquering et al. (1998) showed that the location of the ray itself was a region of zero sensitivity for the travel time of the wave. This contradicts the ray approximation, in that the only sensitive region stretches along the geometrical ray. Recent numerical studies (e.g. Maceira et al., 2015) provided evidences that finite-frequency can result in more accurate body wave inversions. Dahlen et al. (2000) invented a simple technique to estimate the so-called banana-doughnut kernels, in which the sensitivity of the wave is expressed as a double integration over the forward wave-field from the source and backward wave-field from the receiver

to any scattering points in the medium. Since then, their method of calculating finite-frequency has been applied to tomographic inversions to construct the mantle's structure (e.g. Montelli et al., 2004; Zhou et al., 2004; Ren and Shen, 2008).

A major effort of seismic tomography studies has been extracting as much information as possible from the data. While long period studies (e.g. Zaroli, 2016; Ritsema et al., 2011; Schaeffer and Lebedev, 2013) resolve large scale heterogeneities, short period tomographies constrain small scale features. Obtaining sensitivity kernels in multiple frequency bands, one can extract multi-scale heterogeneities from the Earth (e.g. Sigloch and Nolet, 2006; Hosseini, 2016). This idea led to the development of the multi-frequency tomography approach, which has been used in regional and global tomography studies (e.g. Sigloch et al., 2008; Tian et al., 2009; Hosseini, 2016).

1.1.1 Inversion methods

The inverse problem underlying both finite-frequency (multi-frequency) and ray theoretical tomographies is a linearized optimisation, which can be formulated as a linear system of equations as

$$Gm = d, \tag{1.1}$$

where d is traveltimes shift measured in multiple frequency bands, m is the P-wave velocity structure of the Earth, which is unknown, and G represent sensitivity of the measurements to the velocity model. To solve the inverse problem in this case, the misfit function is usually minimised using an iterative algorithm.

In contrast, full waveform tomographies, in which the wave-field equation is computed numerically, solve the gradient of the misfit function with respect to the unknown parameter (m) iteratively (e.g. Tromp et al., 2005; Fichtner et al., 2006). The gradient is used to update the reference model successively, a process which makes the propagation of waveform in a 3D reference model possible. Although the computation of full waveform in a 3D reference model can improve the quality of tomographic images, the quantitative evaluation of the resolution is a major challenge. This is caused by the expensive computations of conventional methods

for assessing the resolution, i.e. Covariance matrix and synthetic tests, in the case of full-waveform inversions (Rawlinson et al., 2014). For the calculation of Covariance matrix, the explicit computation of sensitivity kernel is required. Synthetic tests also need the same computation as the real inversion. With the large dataset measured in multiple frequency bands, the computation of either methods is far beyond reach.

1.2 Geological settings of the Andes

The Andes build up the largest orogenic system in the world, being developed along 8000 km of the western margin of South America, from Venezuela in the north to the northern Scotia Ridge in the south (Ramos, 2009). Stretching along the western margin of the continent, the Andes has been influenced by east Pacific's plates (Farallon/ Nazca) to the west and by the Caribbean plates to the north. Contribution of million-years of different geological processes including subduction, collision and accretion of the oceanic crusts created complicated segments of magmatism and sedimentary along the strike of the mountain belt. Three major divisions have been identified along the Andes since pioneering work of (Gansser, 1973), including the Southern, Central and Northern Andes. In contrast to the Central Andes, which is widely accepted to be shaped by prolonged subduction of oceanic crust (Ramos, 2009; Sempere, 2008), the evolution of the North Andes (like Southern) is not attributed to a pure subduction setting. The northern block is, in fact, characterized by the presence of oceanic terranes, which accreted to the Pacific margin, from the western flank of the Central Cordillera to the coast, since Jurassic (e.g. Coney, 1994; Spikings et al., 2015), contrasting with the Central Andes which lack Jurassic and younger terrane accretions (e.g. Sempere, 2008). Another paradoxical difference between the Andes' segments is in their elevation. The Central Andes comprises one of the largest world's plateaus, with the average elevation of 4000 m between 12° S and 42° S (Isacks, 1988; Ward et al., 2014), a significant feature that typifies the collisional boundaries. Whereas, the North Andes exhibit evidently lower volumes (and average elevation) than the Central Andes (Sempere, 2008).

These contrasted segmentation of Andes, i.e. 1) the high elevation of the Central

Andes, despite being a continental range, and 2) the presence of oceanic terranes in the North Andes reflect that each segment of Andean mountain belt has undergone dissimilar evolutions (Sempere, 2008), a feature which is not explained by traditional plate tectonics sufficiently.

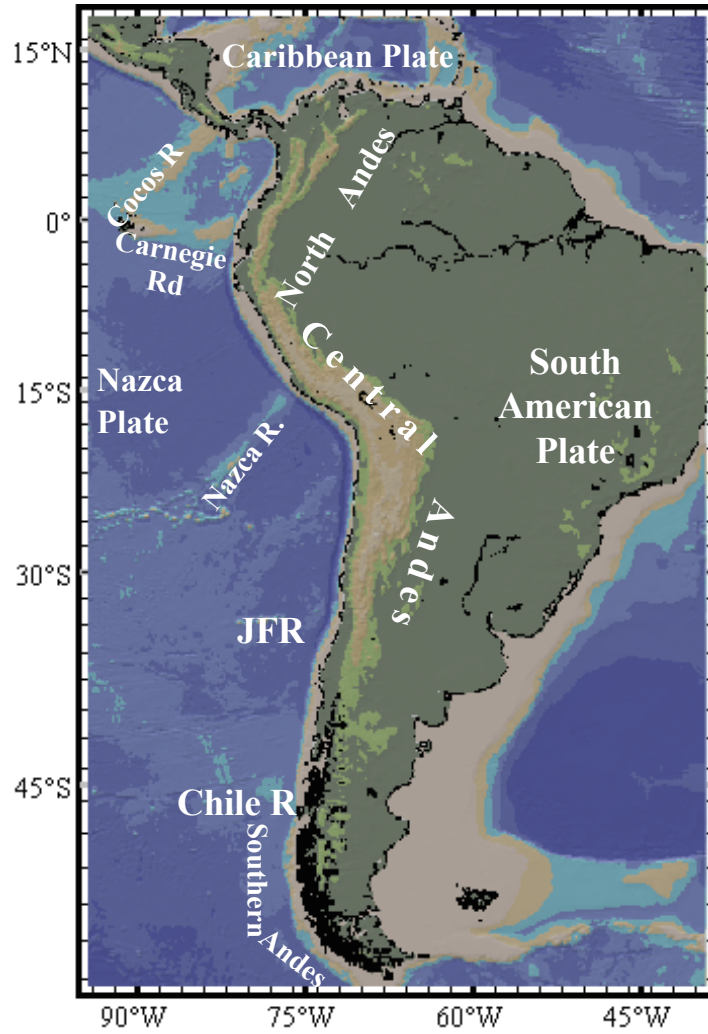


Figure 1.1: Geological map of South America. Rd: Ridge, JFR: Juan Fernandez Ridge. The northern termination of the Andean chain is formed by Andes of Ecuador, Colombia and Venezuela. The Central Andes stretches between Chile fracture zone at latitude 36° S and northern Peru.

1.3 Contribution of tomographic studies to understanding tectonic settings in the Andes

A subducting ocean basin leaves signatures on the surrounding mantle, which is sensed by seismic waves propagating through and recovered as faster velocity anomalies than a reference model by seismic tomography. On the surface, the subducted lithosphere builds up an orogen above the subduction zone. By integrating subsurface images and surface geological records, one may restore the origin of the slabs and the consumed ocean. Previous efforts to construct and interpret tomographic images in South America are summarised in tables 1.1 and 1.2.

Table 1.1: Technical aspects of previous global P-wave tomographies

P-wave tomography	Data type	Reference model	Forward modelling	Parametrization	Inversion technique
GAP-P4 (Obayashi et al., 2013)	P waves	GAP	Finite-frequency and ray theory	Adaptive blocks based on the kernel density	Linear inversion
HMSL-P06 (Houser et al., 2008)	Body waves, Surface waves	Ak135	Ray tracing	Blocks with 18 layers in the mantle	Linear, LSQR (Paige and Saunders, 1982)
Hosseini2016 (Hosseini, 2016)	P waves	IASP91	Ray tracing, finite-frequency in 8 bands	Adaptive tetrahedral mesh based on kernel density	Linear, LSQR
LLNL_G3Dv3 (Simmons et al., 2012)	P Pn	A customized model	Ray theory	Triangular tessellation grid	Multi-level linear
MITP08 (Li et al., 2008)	Body waves	AK135	Ray tracing, finite-frequency	Rectangular cells adaptive to sampling density	Linear, LSQR
MITP_USA_2016MAY (Burdick et al., 2016)	P waves	AK135	Ray tracing	Adapted grids	Linear LSQR
PRI-P05 (Montelli et al., 2006)	Body waves	IASP91	Finite-frequency	Tetrahedral mesh	Linear, LSQR
SPani-P (Tesoniero et al., 2015)	Surface and, body waves	PREM	Ray theory	Regular grid with 28 layers	Linear, LSQR
UU-P07 (Amaru, 2007)	Body waves	AK135	Ray theory	Cells with irregular size	Linear

Table 1.2: Summary of regional tomographies in South America

Model	Region	Data type
Grand (1994)	Western hemisphere	Shear waves
Vdovin et al. (1999)	South America	Surface waves
Heintz et al. (2005)	South America	Surface waves
Wagner et al. (2005)	Central Chile	Body waves (ray theory)
Miller et al. (2009)	South American-Caribbean border	Surface wave
Bezada et al. (2010)	Southern Caribbean	P waves (finite-frequency)
Porter et al. (2012)	Central Chile	Surface wave
Bianchi et al. (2013)	Argentina (Puna plateau)	Teleseismic P waves (ray theory)
Ward et al. (2013)	Central Andes	Ambient noise
Marot et al. (2013)	Central Chile and western Argentina	Body waves (ray theory)
Scire (2015)	Central Andes ($18^{\circ}S - 28^{\circ}S$)	P wave (ray theory)

Upper mantle structures under South America show common features in these studies, including a long sliver of slab along the western margin (e.g. Montelli et al., 2006; Amaru, 2007), which was associated to the subduction of Nazca plate (e.g. Heintz et al., 2005; Scire, 2015). The anomaly is intersected by two low velocity anomalies (e.g. Obayashi et al., 2013), which were associated with the subductions of the Nazca Ridge under Peru (Scire, 2015) and the Juan Fernández Ridge in Chile (Portner et al., 2017). Upper mantle structures underneath the Pacific ocean have been dominated by a wide volume of low velocity anomaly (e.g. Amaru, 2007; Burdick et al., 2016). The slow blanket, which parallel the western margin of the continent, was related to the presence of the several bathymetric highs in the eastern Pacific including Cocos, Carnegie, Nazca ridges by Heintz et al. (2005).

Another feature in the upper mantle images, is a high velocity anomaly under the north-western margin of South America, which was interpreted as Caribbean subduction beneath Eastern Cordillera of the North Andes (e.g. Taboada et al., 2000). High angle subduction of the Caribbean plate has also been inferred beneath NE Colombia and NW Venezuela by other researchers (e.g. Bezada et al., 2010; van Benthem et al., 2013). In the lower mantle (1500 km), a complex mass of slab can be observed under the northern regions (latitudes $> 10^\circ \text{ N}$) in global tomographies, although with different amplitudes and geometries (e.g. Houser et al., 2008; Montelli et al., 2006). Some studies also recovered slabs in the south-eastern area (e.g. Montelli et al., 2006; Amaru, 2007; Obayashi et al., 2013). A slab gap region under Central Andes is a dominant feature between 1300 km and 900 km in the majority of global models, while a NW striking high velocity anomaly is located toward the central parts (e.g. Amaru, 2007; Simmons et al., 2012). The deep structures under South America have mostly been interpreted as Farallon subduction beneath the continental margin (e.g. Ren et al., 2007; Li et al., 2008). The long-lived sinking of Farallon/Nazca might be consistent with the evolution of the Central Andes, which contains episodes of arc magmatism over the past 200 Ma (Ramos, 2009). However, perennial Farallon subduction beneath the continental

margin is unable to explain the accretion of sequences of oceanic terranes along the Northern Andes throughout Upper Jurassic to Eocene times.

The prolonged spreading of Farallon isochrones (since 180 Ma) seems to lead to a simple tectonic setting for the Andean range. A catalogue of lower-mantle slabs by van der Meer et al. (2018), which concurs with the prevailing eastward subduction of Farallon plate under the western margin of South America, is the latest example of incomplete reconstructions of the Andean contradictory evolution. The question remains why these studies were unable to propose a suitable model for the Andean tectonic history. One possibility can be the quality of the tomography models they used. Another can be the assumption they made for the subduction's dynamic. To find the possible answers of these questions, we set up a new tomography under South America with massive volume of data released in the recent years.

1.4 Objectives of this study

This research has principle goals including:

1. presenting a P-wave tomography under South America using a multi-frequency technique,
2. investigating mantle structures under South America,
3. describing the link between deep structures under the continent and abundant geological information in the Andean mountains, and
4. reconciling geological records in the Andean mountains and kinematic models.

In the following chapter, I will first explain the main steps taken to construct the tomographic model. It is followed by presenting tomographic results in chapter 3. Mantle structures under south America, and their connection with the evolution of the Andean mountain range are investigated in chapters 4 and 5. Finally, I come to the conclusion and summarise the main features of the current study in chapter 6.

2

Multi-frequency traveltime tomography

This chapter explains how the final tomographic model is obtained. In finite-frequency tomography, we invert $Gm = d$, where d vector is traveltime shift, m is the unknown vector of P-wave velocity model, and G is the matrix of the sensitivity kernels. To obtain and solve this equation, the following steps are required to be taken:

1. Selection of seismic source-receiver pairs and inversion of the Source Time Functions (STF) for the selected events,
2. Computation of finite-frequency travel times (d) in multiple-frequency bands,
3. Calculation of the sensitivity (G) of the traveltime measurements using finite-frequency approach,
4. Applying suitable source corrections,
5. Solving $Gm = d$ using a suitable regularisation.

In the following section, I present the selection of the observed seismograms and preparation of Source Time Functions. Section 2.2 covers the traveltime measurements (d) in multiple frequency bands. Subsequently, computation of sensitivity kernels (G) is explained in section 2.3. This is followed by section

2.4, in which the inversion of $Gm = d$ is investigated. In the final section, the assessment of the inversion solution is presented.

2.1 Data selection

Over the past few decades, the number of digital seismograph stations has increased significantly, and high-quality digital seismograms have become widely available. In this study, broadband seismograms from 3505 teleseismic earthquakes of $M_b \geq 5.6$, recorded between 1990 and 2016 by regional and global networks were used. Vertical components of broadband seismograms from Incorporated Research Institute for Seismology (IRIS) and other global networks were retrieved, using obspyDMT tools (Hosseini and Sigloch, 2017). The data set includes full track of USArray Transportable stations, which consist of 400 transportable broadband seismometers. Occupying locations for 2 years on a 70-km grid, the network covered the United States from the east in 2004 to the west in 2014. Figure 2.1a depicts geographical location of broadband sources used in this work.

2.1.1 Source Time Functions

An observed seismogram comprises the source signature and far-field propagation effects. If propagation effects are effectively removed, what remains is called a source time function (STF), which is a time series indicating the history of the seismic moment release during an earthquake. STFs can be used to calculate the radiated energy, as they contain the spectrum of the source process. Their characteristics can also be investigated to study the effects of the source on the earthquake rupture and ground motion prediction (Courboux et al., 2016) and to explore how the energy release behaves with respect to depth and tectonic settings (Houston, 2001). STF for a point source in an infinite non-attenuating medium can be directly represented by the body waves scaled by the radiation pattern, the distance, and the elastic properties (Vallée and Douet, 2016). In a more realistic case of spherical earth, estimation of STF becomes more complex. The STF configuration becomes more complicated in shallower depths, when surface-reflected waves (e.g. pP, sS)

interfere with the direct body wave. Regarding, the important role of STFs in analysing earthquake behaviour, Tanioka and Ruff (1997) published a catalogue of several STFs for earthquakes of the 1990s. However, the catalogue was not updated. Recently, Vallée and Douet (2016) presented a catalogue of 2800 STF for earthquakes occurred from 1992 to 2014, using SCARDEC approach by Vallée et al. (2011). The method offers the STFs, together with the first order earthquake source parameters, including seismic moment, depth and focal mechanism for moderate to large earthquakes occurred since 1992.

Due to the directivity of the rupture propagation, and earthquakes' finite slip areas, each seismic station provides a different estimate of the STF. Thus, the apparent source time function (ASTF), which is recovered from a single seismometer, may differ from the true STF (Plourde and Bostock, 2017). Sigloch and Nolet (2006) developed an approach to account for the near-source effects and azimuthal dependencies by clustering the geographically neighbouring seismograms. The algorithm proceeds in two major stages: 1) clustering waveforms into distinct groups, and 2) estimation of STF for separate clusters. This approach is especially effective in measuring the amplitude and traveltime anomalies for shallow earthquakes at teleseismic distances, which is complicated by interference of surface reflections, that may change polarity, depending on the epicentral distance or azimuth (Nolet, 2008).

Estimation of STF is an important step in waveform tomography. The Green's function is the solution of the elastodynamic equation for a point source and does not contain the near-fields effects of the source. To obtain a more realistic synthetic seismogram, which is required to measure the traveltime anomalies, the Green's function computed for each source-receiver pair needs to be convolved with the corresponding STF. Sigloch et al. (2008) obtained STFs of earthquakes, occurred between 1990 to 2008, removing the effect of interfering reflections. To complement the dataset, I employed the same technique as was employed by Sigloch et al. (2008) and estimated STFs of 2000 earthquakes with magnitude $M_b \geq 5.6$, which were recorded from 2008 to 2016. Teleseismic P waves recorded at the global stations of

Global Seismographic Network (GSN) were employed to determine the earthquake source parameters (focal mechanism, depth and STF). For each earthquake, the stations were downloaded using obspyDMT tools (Hosseini and Sigloch, 2017), by initially selecting all available BHZ data. Green's functions are calculated using Chapman's WKBJ code (Chapman, 1978) using PREM (Dziewonski and Anderson, 1981) as the reference model and published moment tensors.

In this method, for each earthquake, observed seismograms are clustered into groups with highest similarities, for which the STF inversion is then implemented at different depths in a linearized inversion. For each cluster, first, all source permutations are convolved with the Green's functions. The resulting synthetic seismograms, are then cross-correlated against observed waveforms over a 25-sec long window around P phase. The STF with the highest cross-correlation coefficient (CCC) > 0.8 , is proposed as the best solution. This leads to a more confident depth estimation than earthquake catalogues, a parameter which is important in waveform tomography. For some events, STF can be empty due to either no high-quality data within the teleseismic distance and azimuth range, or the best fitting STF has a CCC below 0.8. Preferred STFs were then qualified, based on their simplicity and non-negativity. Simplicity means unimodal STFs were preferred over bimodal or noisy solutions, provided that the RMS misfit between predicted waveforms and data is not larger for unimodal. In total 3505 STFs, including those from Sigloch et al. (2008), were created. Figure 2.1 indicates the global distribution of earthquakes with high quality STFs. This dataset has the advantage of being consistent, since all the events with magnitudes $M_b \geq 5.6$, that have occurred since 1990 until 2016, were analysed with the same method.

2.2 Traveltime measurements

Traditionally, short period recordings have been chosen for approximation of the seismic waves arrival time. In this case, each traveltime is measured by hand-picking the first deflection. The ray theory assumes such a picked time corresponds

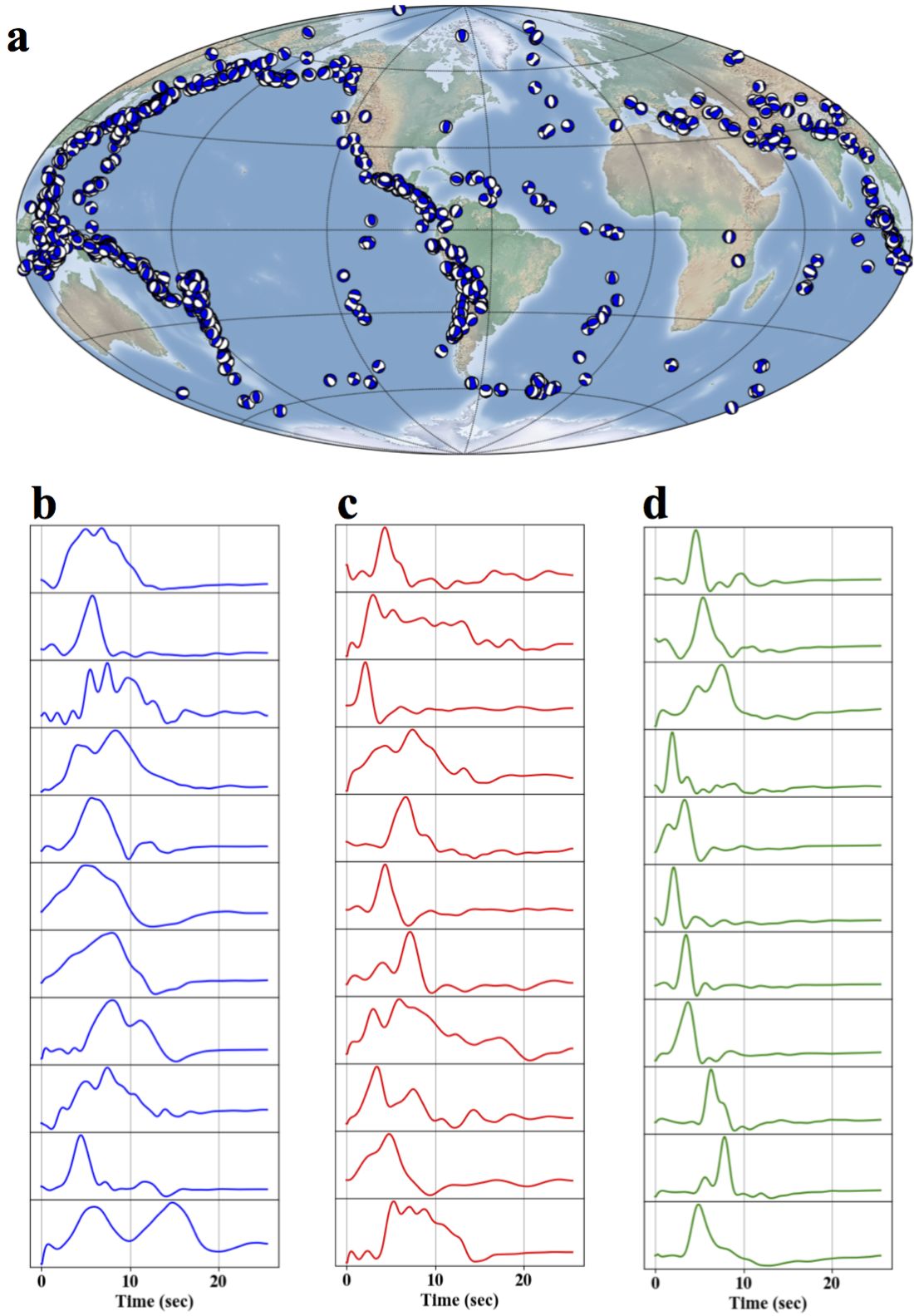


Figure 2.1: a) Geographical location of sources with qualified STFs. The estimated beachball solution are displayed on the location of each earthquake. Examples of broadband STFs for earthquakes in South America at b) 0 to 25 km , c) 25 to 50 km, and d) 100 to 250 km depths.

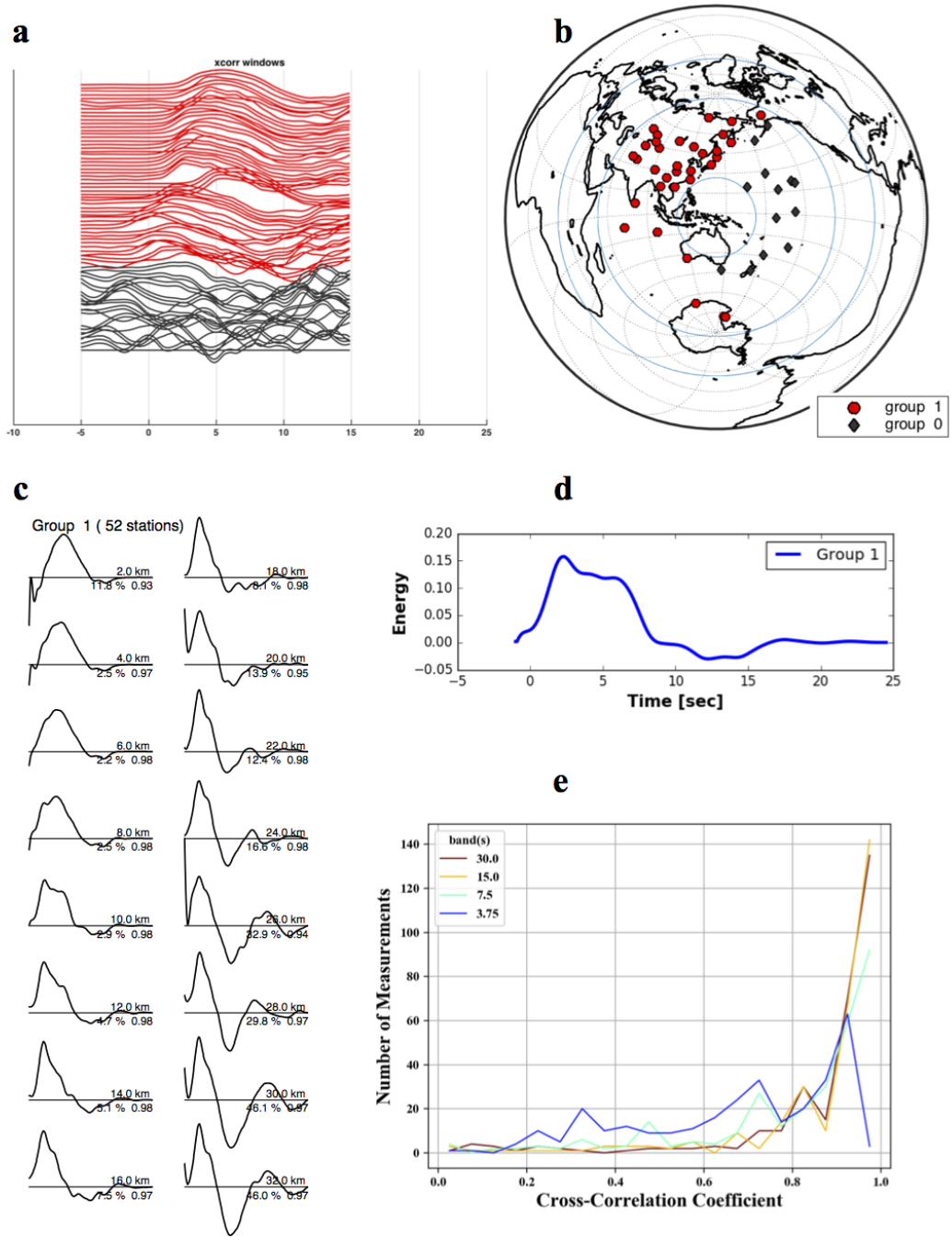


Figure 2.2: An example of a STF inversion for 2013-04-16 earthquake in north coast of New Guinea. a) clustering of the waveforms from the broadband stations. The red coloured waveforms are chosen for the inversion. The black colour waveforms are not considered in the inversion. b) global distribution of the broadband stations. c) estimated STF for the selected waveforms at different depths. d) Preferred STF at 10 km depth. e) Cross-correlation coefficients in four frequency bands. Synthetic seismograms for the figure are measured by the convolution of the selected STF in d with Green's functions.

to the arrival of the highest-frequency observable phases (Dahlen et al., 2000). Traveltime of each observation is then compared to that of model predictions, leading to a time anomaly. Modern digital instrumentation has significantly led to the development of automatic traveltime measurement methods, by cross-correlating observed and synthetic seismograms (e.g. Montelli et al., 2004). For a well-isolated phase, arriving in a time interval of $[t_1, t_2]$, the cross-correlation of synthetic and observed seismograms (s) can be written as

$$\Gamma(\tau) = \int_{t_1}^{t_2} s(t - \tau) s^{obs}(t) dt \quad (2.1)$$

in which, the traveltime residual (dT) is found when the cross-correlation function is maximised (Dahlen et al., 2000). The frequency content of the wave can be preserved, if we measure the dT of the phase, here P, in different frequency bands. In this study, I used a cross-correlating algorithm, developed by Hosseini and Sigloch (2017), to measure traveltime anomalies in 8 frequency bands. The flowchart in Figure 2.3a illustrates the main steps to estimate traveltime anomalies. The Green's functions were computed with the Instaseis (van Driel et al., 2015), a database of synthetic seismograms in global and regional scales using AxiSEM (Nissen-meyer et al., 2007), which is an axisymmetric spectral element method to generate wavefield. IASP91 velocity reference model (Kennett and Engdahl, 1991), completed with the Q model from PREM (Dziewonski and Anderson, 1981), and attenuation were used to determine the predicted seismograms, which were then convolved with the source time function for each source-receiver pair to produce synthetic seismograms. The predicted and observed waveforms are then sent to the filters and cross-correlated. We use Gabor filters, which are Gaussian in the log-frequency domain. Having constant fractional bandwidth, the eight chosen filters cover a broad band of frequency content, ranging from 2.7 to 30 seconds (Figure 2.3b). The upper limit of the frequency is chosen to be 2.7 s, since beyond which the number of the successful fits decrease. One important advantage in using the Gabor filters is that they are effective in separating the low-frequency bands, which are very narrow and close to each other in the linear-frequency domain (Tian et al., 2009).

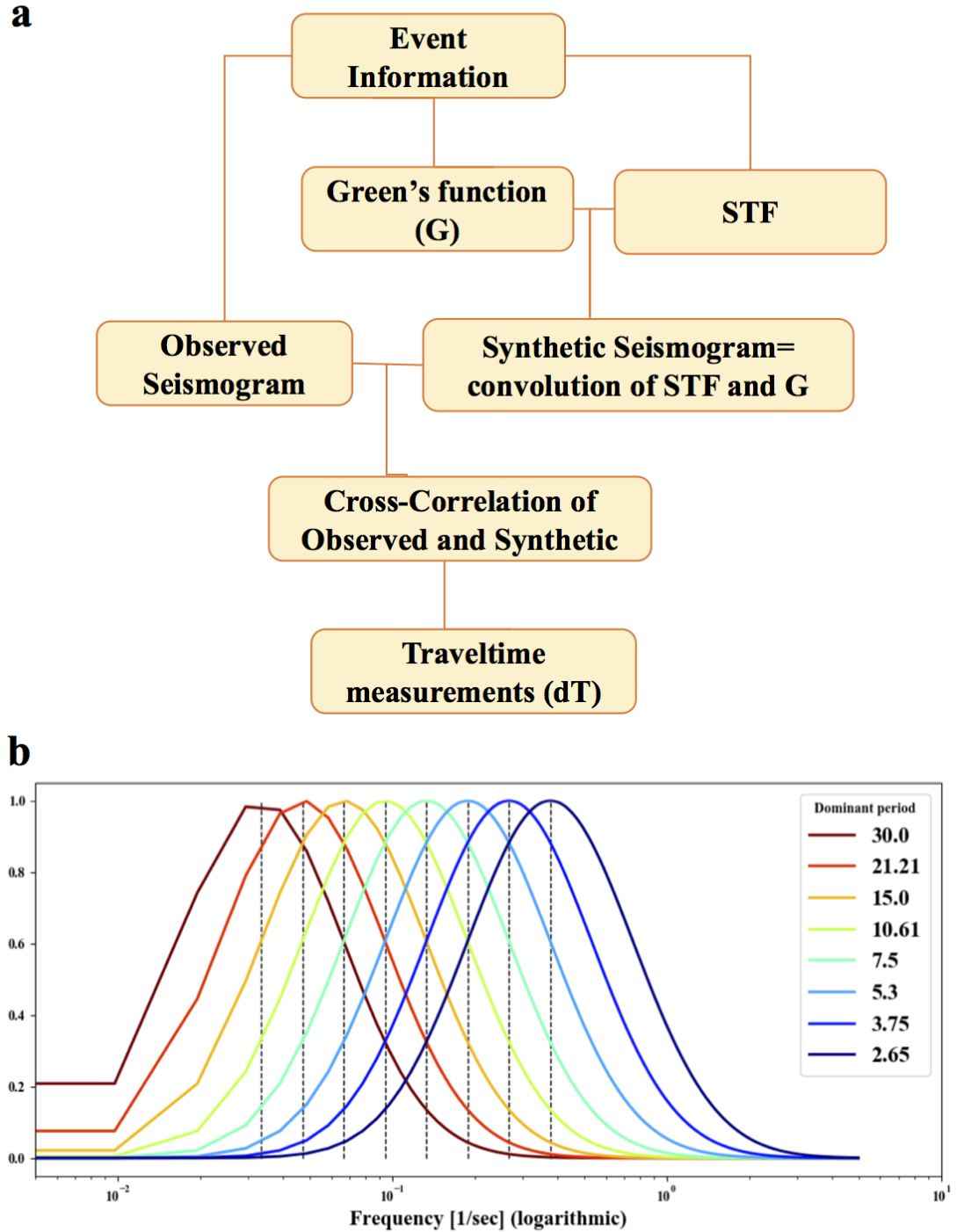


Figure 2.3: a) Flowchart representing the main steps in the extraction of the multi-frequency traveltime. Note that the cross-correlation is implemented in 8 frequency bands, leading to 8 measurements for each source-station pair. b) Impulse response of eight Gabor filters in logarithmic frequency domain, with dominant periods between 30.0 and 2.7 sec.

Cross-correlation window in the algorithm is an empirical choice, defined as a fixed multiple of duration of each filter’s impulse response (IR). I have experimented and found that $1.0 \times \text{IR}$ produces robust results. The traveltime shift which maximises the cross-correlation coefficient is considered as the optimized measurement (Dahlen et al., 2000). The sign of the anomaly (delay or advance) tells us whether the wave has travelled through faster or slower media than the reference earth model that has been used to calculate the Green’s functions. A positive traveltime anomaly represents an advance shift, while a negative anomaly corresponds to a delay. The procedure of the multi-frequency measurements is indicated in Figure 2.4.

Since the measurements are used in the tomographic inversion, it is important to determine the observational errors for the inversion. The standard deviations of the traveltime measurements have been determined heuristically in the algorithm. The higher the cross-correlation coefficients are, the smaller value of standard deviations are chosen. Figure 2.5 shows the histograms of traveltime anomalies for all source-receiver pairs in eight bands. The number of successful measurements progressively decreases by increasing the centre frequency, due to the lower signal to noise ratio in higher frequency bands. We imposed a lower limit of 0.8 on the cross-correlation coefficient. Regarding all frequency bands, the dataset contains $\sim 5\text{M}$ traveltime shifts. However, to reduce the data redundancy and reducing the computational costs, only measurements from four frequency bands, i.e. $\sim 2.5\text{M}$ traveltime anomalies, have been corrected for the Earth’s ellipticity, station topography, and crustal structure.

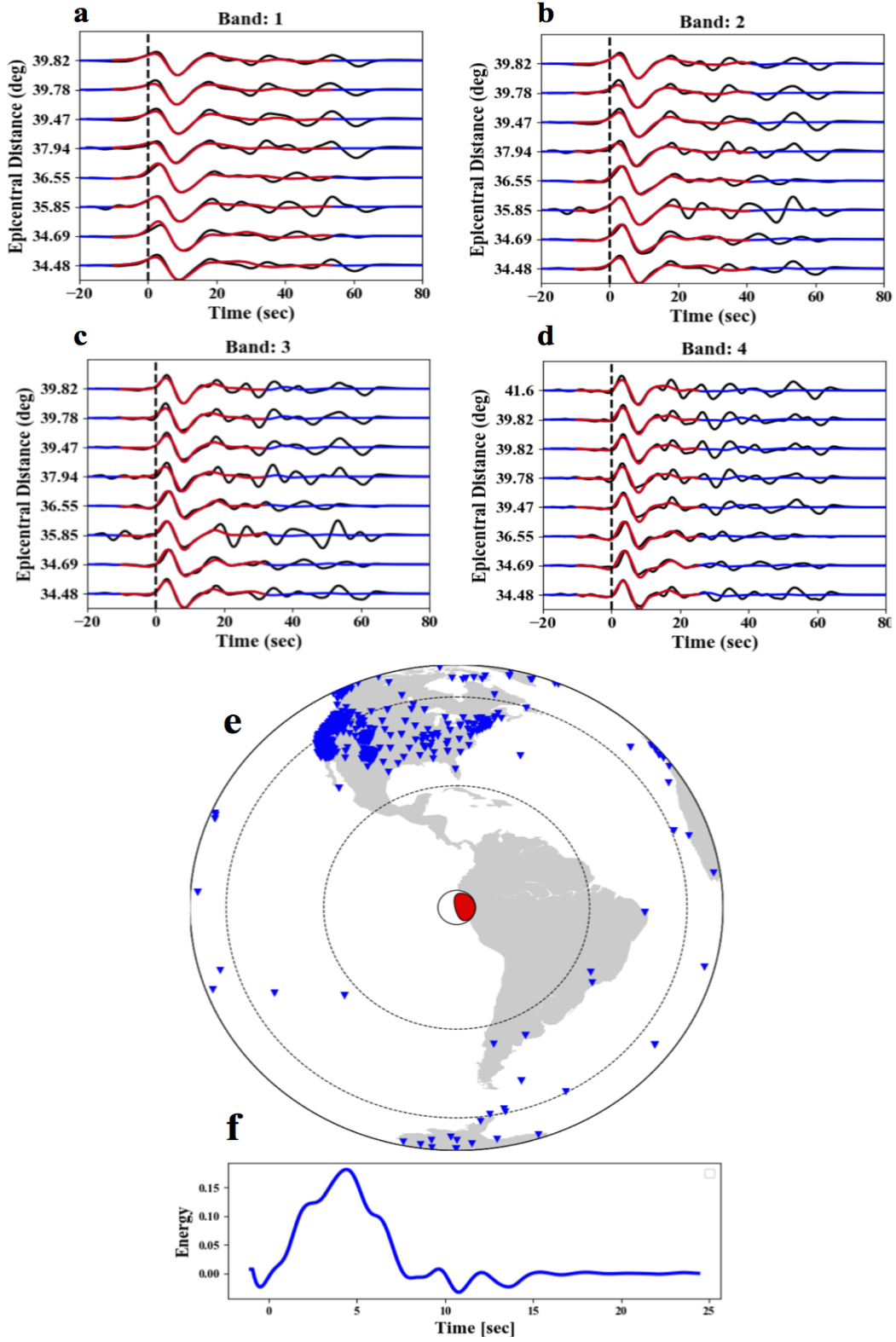


Figure 2.4: Procedure of multi-frequency measurements for an event occurred in northern Peru at 2009-02-09, 14:09:02.98. Observed (black) and shifted synthetic (red) waveforms are filtered in 4 frequency bands with centre period a) 30.0, b) 21.21, c) 15.0, d) 10.61 sec. The red beachball and blue triangle in (e) represent locations of the source and stations. Estimated STF is shown in f.

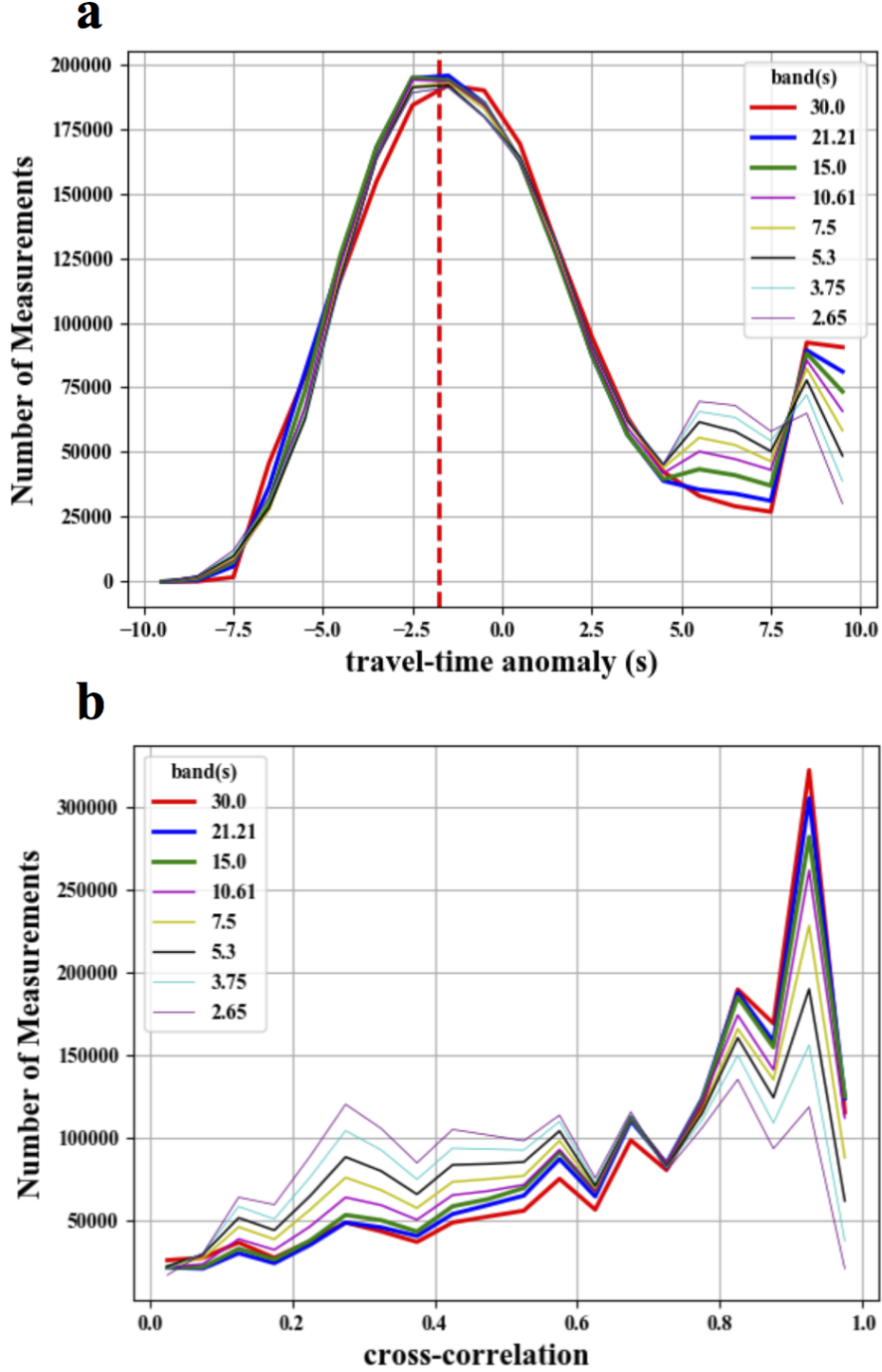


Figure 2.5: Histograms of a) traveltime shifts, and b) cross-correlation coefficients in 8 frequency bands. The cross-correlation values for each band peak well above 0.9. The number of successful measurements ($cc > 0.8$) decreases by increasing the centre frequency.

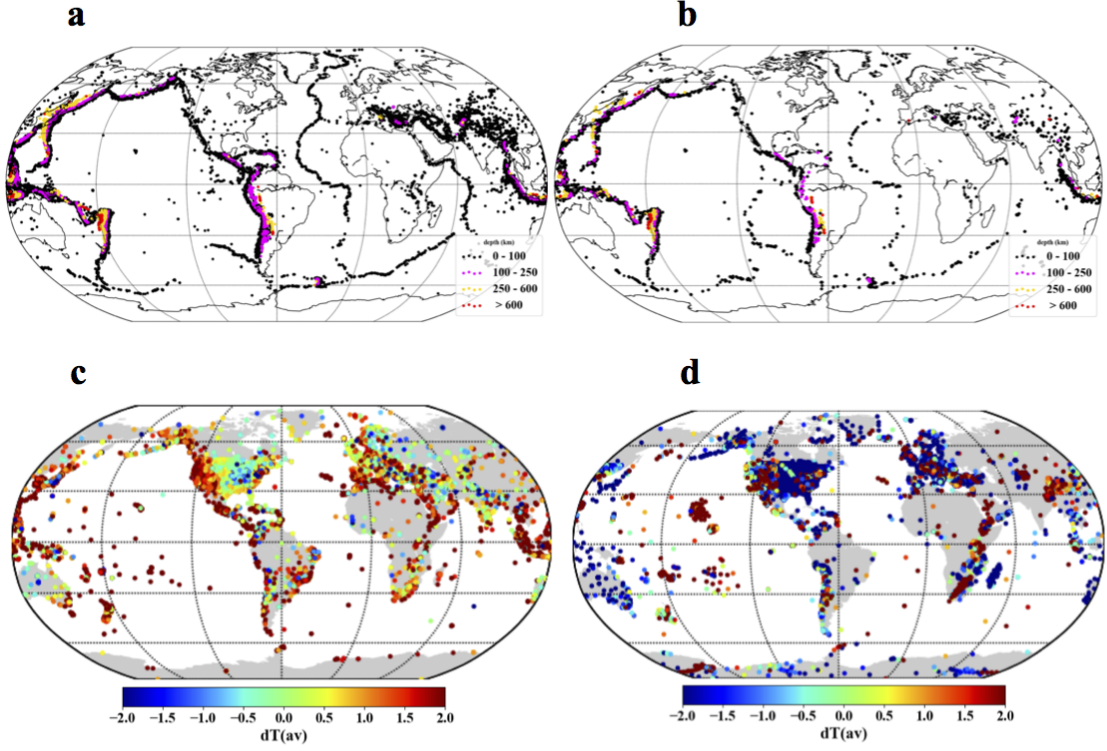


Figure 2.6: Global distribution of 48381 and 3505 earthquakes selected from a) ISC-EHB and b) broadband catalogues respectively. Panels c and d indicate global distribution of ISC and broadband stations used in the inversion. Average traveltime measurements are plotted at the location of each station. Red colour reflects the delay time shift, while blue represents the advance shift.

The P wave finite-frequency is a powerful tool to obtain a high resolution image of the mantle structures. However, the uneven distribution of the earthquakes and stations impose strong limitations to the data coverage, hence image resolution, in some regions. In South America, tectonic setting has narrowed the seismicity mainly along the western and north-western margin of the continent. Broadband stations are also mostly confined to these borders of the plate. To alleviate this shortage, P wave arrival picks provided by EHB catalogue of International Seismological Centre (ISC) from 1967 to 2014 are included. The ISC-EHB bulletin (Weston et al., 2018) contains millions of measurements for different seismic phases. I only choose P phases for teleseismic sources with magnitude ≥ 5.6 and arrival time reading precision < 0.1 s. I considered these short-period traveltimes as the highest frequency band measurements, and included them in the data vector.

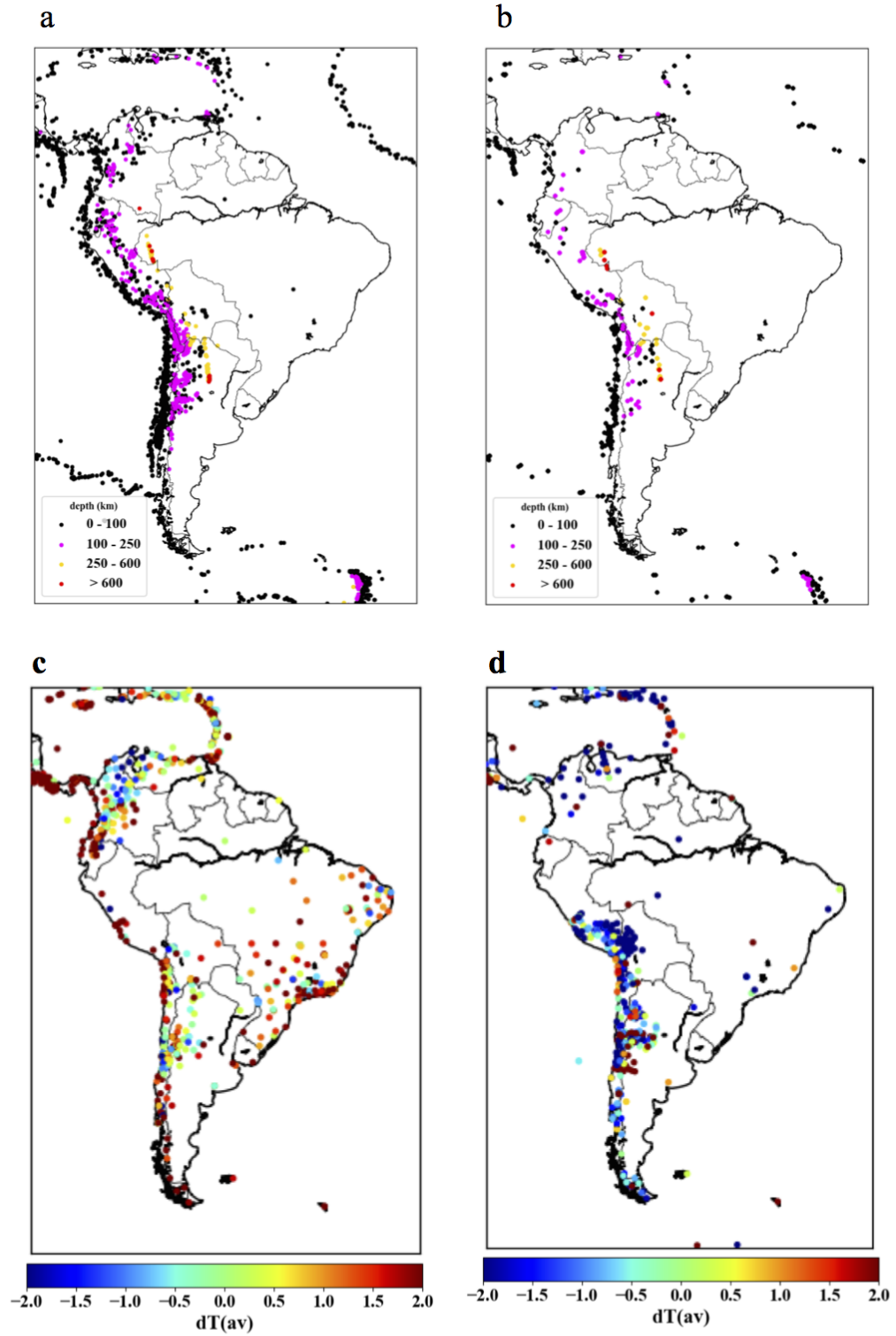


Figure 2.7: Distribution of earthquakes from a) ISC-EHB, and b) broadband datasets in South America. Panels c and d indicate distribution of ISC and broadband stations used in the inversion. Average traveltime measurements are plotted at the location of each station. Red colour reflects the delay time shift, while blue represents the advance shift.

Figures 2.6 and 2.7 illustrate global and regional (in South America only) distribution of the ISC stations, comparing them with those of the broadband stations used for multi-frequency measurements.

2.2.1 Measurements Corrections

The reference models, which are used to calculate the traveltime measurements, are spherical models with a radial distribution of velocity. As a result, traveltime measurements are influenced by only the epicentral distance between source and receiver. However, in practice, the Earth is not as simple as the reference models. The locations of both the source and receivers affect the measurements as well. Therefore, crustal properties are needed to be included in the computations. One of the common corrections is the ellipticity, which is caused by the rotation of the Earth. Dziewonski and Freeman (1976) presented a precise method, which considers the deflection along the ray, caused by the ellipsoidal shape of the Earth, and presented a tabulation of ellipticity corrections. Kennett and Gudmundsson (1996) also presented tabulated corrections for different phases, which is available as `ellip.f` in the public domain.

I used a program developed by Tian et al. (2007), which computes ellipticity, crustal and topographic corrections and attenuation. The ellipticity correction is the sum of anomalies caused by lengthening or shortening of the ray at its ends, displacements of discontinuities along the ray, and an integral over velocity perturbations result from the elliptical perturbation. In addition to the ellipticity, the algorithm corrects the traveltime anomalies for the crustal structure using crustal model CRUST2.0 by Laske (available from <https://igppweb.ucsd.edu/~gabi/crust2.html>).

Although, in global tomography, it is not generally possible to image the heterogeneities in the crustal structure, one can still correct traveltimes for the anomalies induced by the crustal thickness and velocity. Normally, this is done by applying a known crustal model, here CRUST2.0, rather than resolving the crustal structure in the tomographic inversion. In `raydata.f` (Tian et al., 2007), crustal

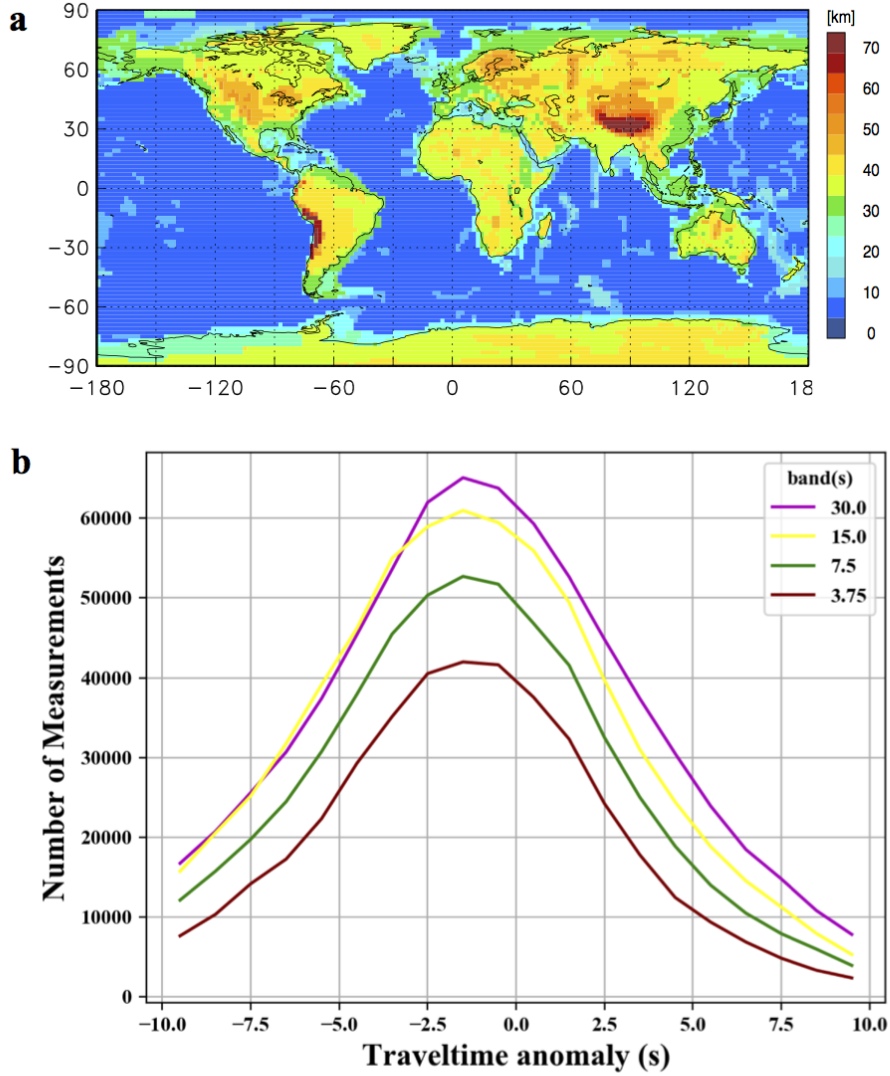


Figure 2.8: a) 2D projection of CRUST2.0 crustal thickness (from <https://igppweb.ucsd.edu/~gabi/crust2.html>). b) Histograms of multifrequency traveltime anomalies at four frequency bands that were used for the final inversion, after correction for Earth's ellipticity, crustal structure, and station topography.

corrections are implemented by first subtracting the traveltime in the reference model's crust, and then adding the traveltime of the known 3D crustal model.

Topography corrections are embedded in CRUST2.0. However, averaging over its $2^\circ \times 2^\circ$ grid creates a small residual topography for the stations, which is corrected as

$$\delta T = (h - h_{3D}) \frac{\cos i}{C_{3D}} \quad (2.2)$$

where i and C_{3D} are the incident angle and the velocity in the 3D crustal model, h

and h_{3D} stand for elevation of the receiver and 3D crust surface (Tian et al., 2007). Overall traveltime measurements were corrected for three types of residual caused by ellipticity (T_{ell}), crustal effects (T_{cru}) and topography (T_{top}). Figures 2.6c-d and 2.7c-d indicate global projection of the average traveltime measurements at each station after being corrected. Histogram of corrected measurements is displayed in fig. 2.8b.

2.3 Computation of sensitivity kernels

In this section, computation of sensitivity kernels for the ray and finite-frequency traveltimes is explained. In ray theory, traveltime of a ray is estimated by a one-dimensional integral along the ray travelled through an unperturbed reference model as follows

$$T = \int_L \frac{dl}{c(r)} \quad (2.3)$$

where L refers to the length of the ray path, $c(r)$ is the velocity of the seismic wave through which the waves travel and dl is the differential arc length along the ray. Assuming the traveltime of the ray, calculated through a reference model, is given by

$$T_0 = \int_{L_0} \frac{dl}{c_0(r)} \quad (2.4)$$

the traveltime anomaly, i.e. the difference between observed (T) and predicted (T_0) rays is calculated as

$$\delta T = T - T_0 = \int_{L_0} \left(\frac{1}{c} - \frac{1}{c_0} \right) dl \approx - \int_{L_0} \frac{\delta c}{c^2} dl. \quad (2.5)$$

In finite-frequency, the 1D integral in the equation eq. (2.3) is replaced by a volume integral over the mantle ($\oplus$) as:

$$\delta T = \int_{\oplus} \kappa(x) \frac{\delta c}{c} d^3x \quad (2.6)$$

in which $K(x)$ represents the finite-frequency sensitivity of the traveltime anomaly (δT), to the wave speed perturbation (δc).

For a more general case, the total perturbed wavefield can be divided into discrete scattering points, integrating over the enclosing volume, provided that the wave

is a single phase such as P or S (Nolet, 2008). Born approximation considers all single scattered waves, which are propagated from perturbations between the source and the receiver. This means that for every single scattered wave, we can add a perturbation term $\delta s(t)$ in the computation of the seismogram as follows

$$s(t) = s(t) + \delta s(t). \quad (2.7)$$

Taking Born approximation into account in conjunction with paraxial ray tracing, Dahlen et al. (2000) developed a technique, in which calculation of the $K(x)$ was reduced to a double sum over all scattered body waves. Their paraxial Fréchet kernel for body wave traveltime can be written as

$$K(x) = \frac{-1}{2\pi c} \sqrt{\det |M_1 + M_2|} \times \frac{\int_0^\infty \omega^3 |\dot{m}(\omega)|^2 \sin \Phi d\omega}{\int_0^\infty \omega^2 |\dot{m}(\omega)|^2 d\omega} \quad (2.8)$$

where M_1 and M_2 are forward and backward traveltime Hessian along the central ray, and ω shows angular frequency. The quantity $|\dot{m}(\omega)|^2$ represents the power spectrum of the synthetic seismogram, indicating the frequency dependence of the kernel (Dahlen et al., 2000; Montelli et al., 2004). Although the integration limit is up to infinity, in practice it is restricted to the bandwidth of $|\dot{m}(\omega)|^2$. Φ stands for possible phase shift produced by scatter point, defined as

$$\Phi = \frac{1}{2} \omega q^T \cdot (M_1 + M_2) \cdot q - \text{sig}[(M_1 + M_2) - 2] \frac{\pi}{4} \quad (2.9)$$

where q is the location vector of the scattering point in the ray coordinates. The symbol “sig” indicates number of positive minus the number of negative eigenvalues of $M_1 + M_2$ (Montelli et al., 2004).

The most significant characteristic of the finite-frequency kernels is that the sensitivity kernel along the central ray is zero. It results from $\Phi \rightarrow \infty$, implying that traveltime required for the ray to visit the scattering point located on the central ray is zero. Under paraxial approximation, the forward scattered wavefield from a point far from the central ray, arrives outside of the narrow time window of the cross-correlation. Thus, the width of the cross-correlation window affects the width of the kernels. In other words, the effective medium, in which the constructive

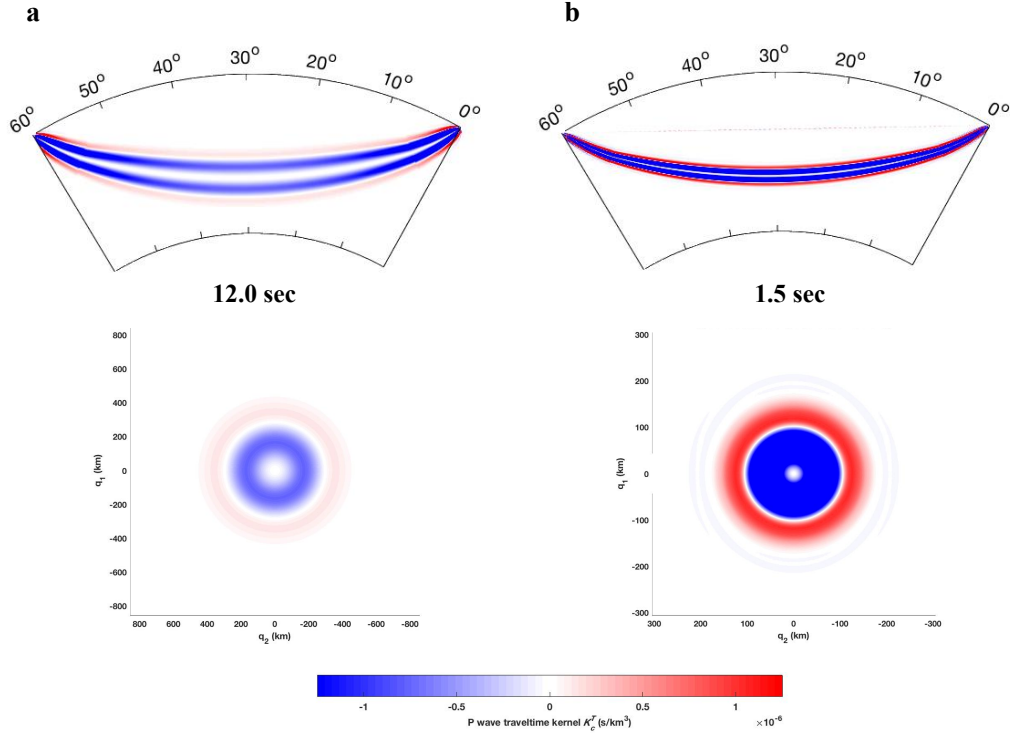


Figure 2.9: a) Cross-section through sensitivity kernels with dominant periods 12.0 and 1.5 seconds at epicentral distance 60 degree. b) Cross-sectional view perpendicular to the ray at 30 degree.

scatterings happen is narrowed to a volume as wide as the first or second Fresnel zone (or influence zones for band-limited traveltimes). Figure 2.9 indicates how the width of the kernels change by the dominant periods of the cross-correlation window. The sensitive area along the central ray is wider in the lower frequency kernels. In this study, I used the paraxial finite-frequency technique by Dahlen et al. (2000) to calculate the kernels. Sensitivity kernels were calculated through IASP91 velocity reference model (Kennett and Engdahl, 1991), completed with the Q model from PREM (Dziewonski and Anderson, 1981), and attenuation. To calculate the kernels, the reference model was discretized using a tetrahedral mesh, which is discussed in the following section.

2.3.1 Parameterisation

To solve the forward and inverse problems, we need to discretise the velocity model. The choice of the parametrisation influences the resolution limits of the tomography. Traditionally, the spatial variations in the rendering power of the data was ignored and the model space was simply discretised in a uniform mesh, i.e. cubic cells or spherical harmonics. However, there is a risk of losing some features if the cell size is coarser than the rendering power of the data. On the other hand, over-parametrisation increases the computational costs of the inversion. In addition, poorly constrained parts of the model will have a higher error if they are over-parametrised. This compromise between the minimum length scale of the mesh, constrained by the data, and the computational cost has been the primary motive to develop irregular meshes, in which the size of model parameters is adjusted to the geographical variations in the data (ray or kernel) density (Rawlinson et al., 2014). Pioneering works by Sambridge et al. (1995) and Sambridge and Gudmundsson (1998), were first to propose the application of Delaunay tetrahedra and Voronoi polyhedra in the tomography. Since then, utilization of the tetrahedral meshes were the subject of different studies (Montelli et al., 2004; Sigloch, 2011).

Hosseini (2016) built a tetrahedral mesh using DISTMESH tool (Persson and Strang, 2004), adjusted to the data density in his global tomography. The mesh was designed based on the Delauney criterion, in which no tetrahedron contains internally a vertex of another tetrahedron. The velocity model at any point between four vertices of a tetrahedron was defined by linear interpolation. The mesh, which is also used in this study, was created in two steps: first finite-frequency kernels were calculated for a global dataset in a uniform mesh with 200-km edge length. In the next stage, the number of vertices were modified according to the kernel density obtained in the first step. This has decreased the length of tetrahedron's edges to the minimum length of ~ 66 km under the regions with densest kernel density, e.g. North America. Under South America, the minimum edge length is ~ 75 km, whereas maximum edge covers ~ 418 km distance. The total number of mesh vertices is 396501, 15865 of which are located under South America.

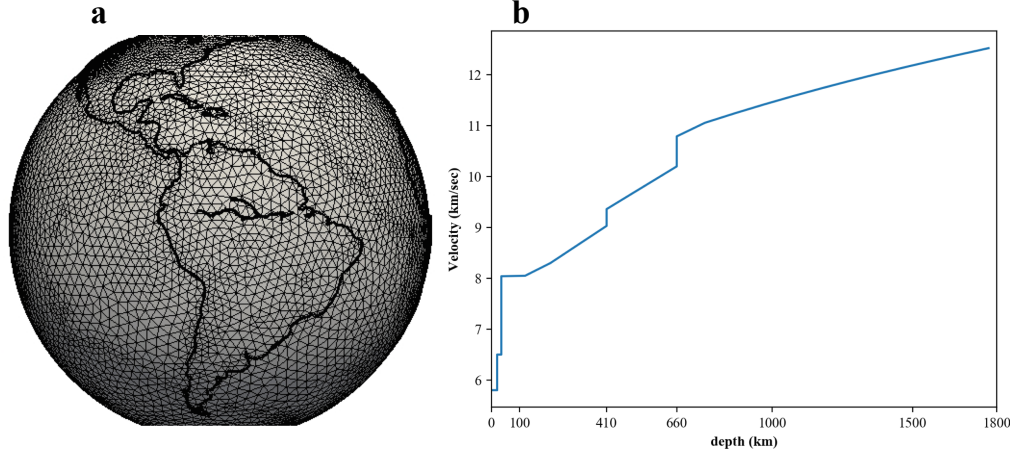


Figure 2.10: a) Global tetrahedral mesh, and b) the modified IASP91 velocity model, used in the forward and inversion computations.

Once the type of the basis function was chosen, the model space is represented in terms of N number of parameters as follow

$$m = \sum_{j=1}^N m_j h_j \quad (2.10)$$

where, h_j is the j^{th} basis function. This allows us to formulate eq. (2.6) in the following matrix form

$$\delta T_i = \int_{\oplus} K_i(x) m(x) d^3x = \sum_{j=1}^N m_j \int_{\oplus} K_i(x) h_j(x) d^3x = \sum_{j=1}^N G_{ij} m_j \quad (2.11)$$

where i and j refer to the index of the measurement and model parameter respectively. G_{ij} represents elements of G matrix as

$$G_{ij} = \int_{\oplus} K_i(x) h_j(x) d^3x \quad (2.12)$$

which represents the value of the i^{th} sensitivity kernel at the j^{th} node of the mesh. The sum of all elements of the G matrix corresponding to that model parameter is defined as the column density, which is indicated in 2D views at figs. 2.11 and 2.12 under South America at selected depths.

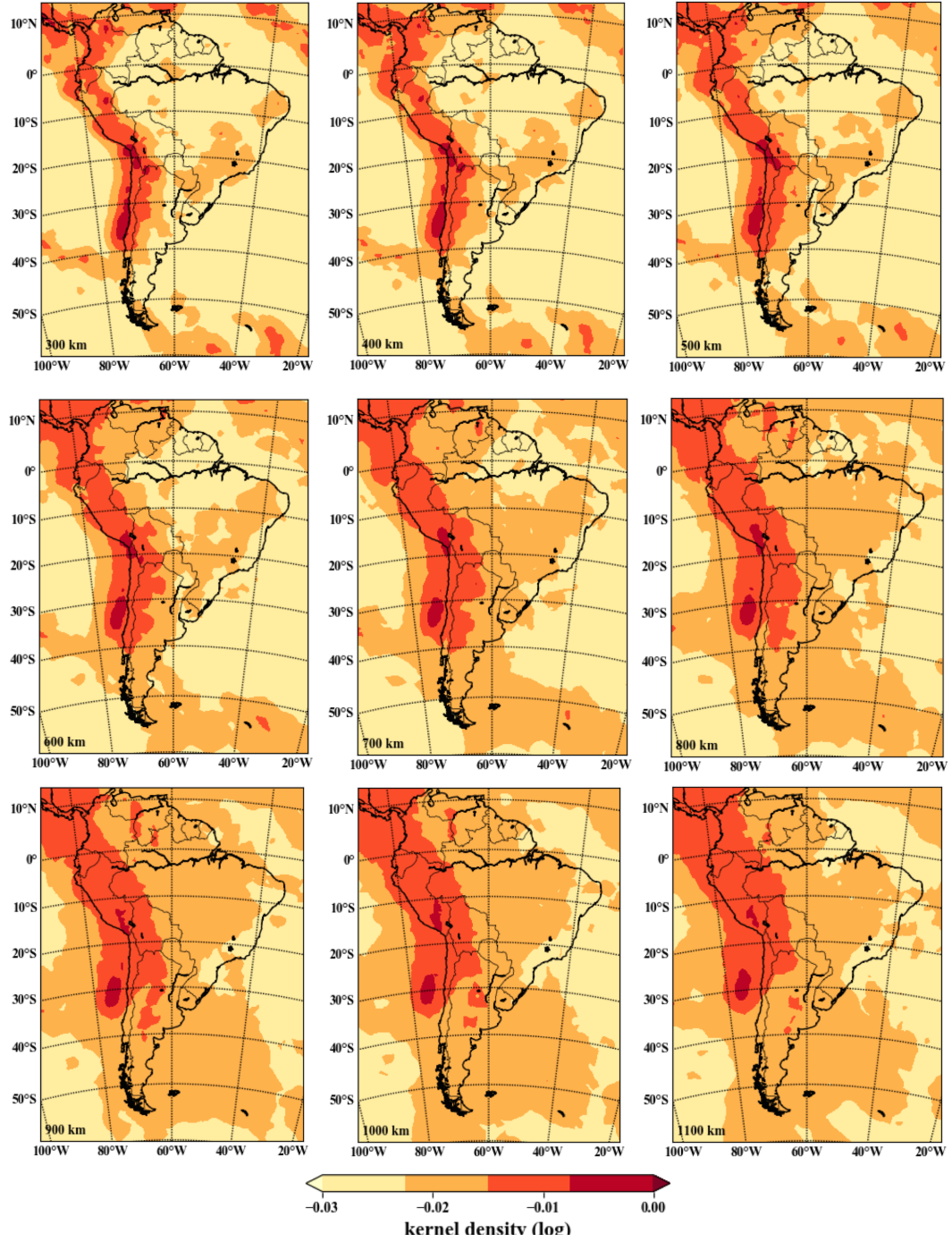


Figure 2.11: 2D map views of the logarithmic (base 10) density of the sensitivity kernels between 300 and 1100 km depths under South America. Larger values of density reflect the denser coverage of western margin.

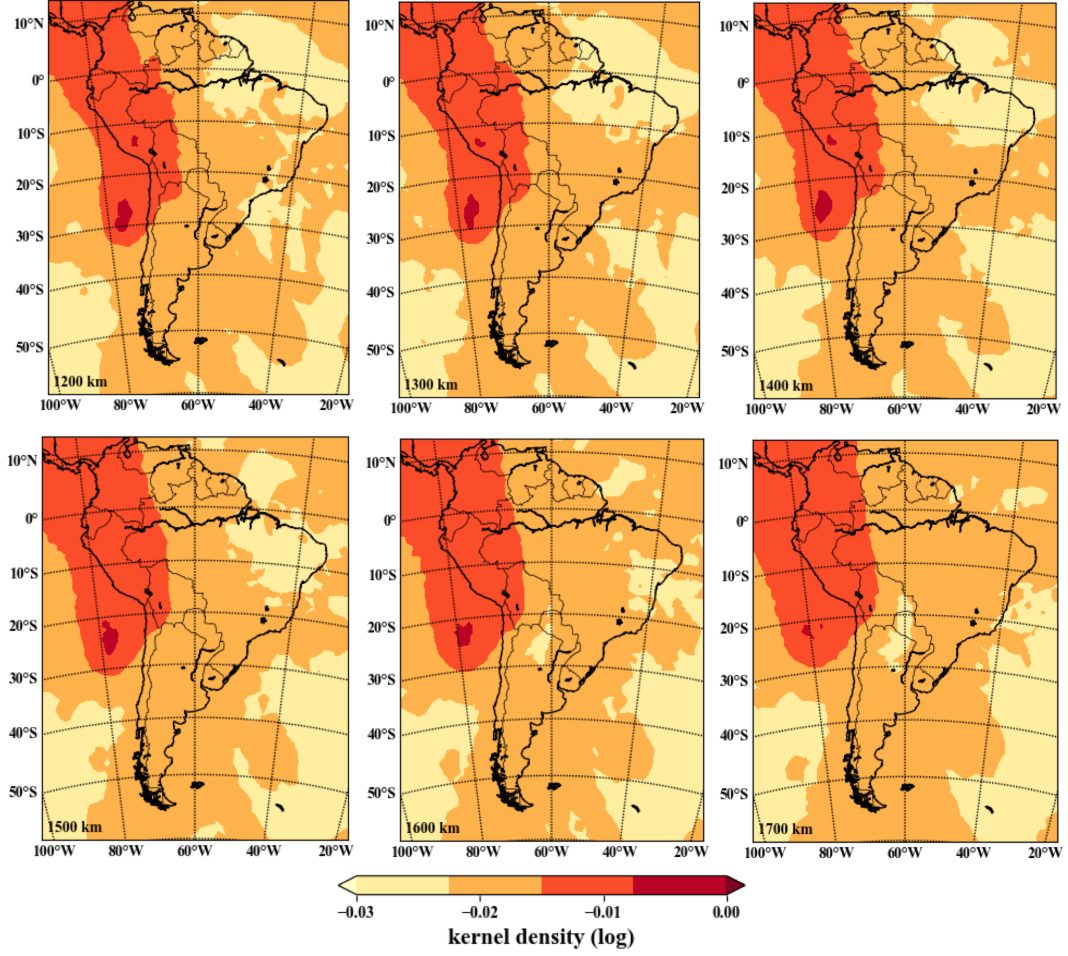


Figure 2.12: 2D map views of the logarithmic (base 10) density of the sensitivity kernels between 1200 and 1700 km depths under South America.

It is evident that there is a sharp discrepancy between kernel density in the western and eastern parts of the continent, which is more visible in shallower depths < 600 km. This is caused by the concentration of the broadband stations along the western border. Furthermore, thicker finite-frequency kernels cause the higher amplitudes to cover a wider area in the deeper depths.

2.4 Linear inversion

As it was explained above, we formulate a linear relationship between the data, i.e. traveltime difference ($d = \delta T$), and model parameter, i.e. P-wave velocity perturbations ($m = \frac{\delta c}{c}$) using the parameterisation. I also inverted source parameters

simultaneous to the velocity model (see section 4.1). It does not violate the linear relationship, as the combination of the two unknown vectors is still a vector.

$$Gm = G_1m_1 + G_2m_2 \quad (2.13)$$

where m_1 and m_2 refer to velocity perturbations and source corrections. The final linear relationship is a large linear system of equations, in the following form

$$Gm = d \quad (2.14)$$

in which, the d vector has ~ 8.5 M elements, including ~ 2.5 M multi-frequency and ~ 6.0 M short-period traveltime anomalies. The final unknown vector includes 396,501 elements for velocity anomalies, 75,312 entities for the hypocentre corrections and 36,885 elements of origin time corrections.

This equation is formulated as a minimisation problem, which satisfies data to a certain level. However, the solution to this criterion is not unique, due to the following issues:

1. the presence of noise in the seismic data, which is often difficult to precisely quantify, is ubiquitous (Rawlinson et al., 2014). This causes the equation to be ill-posed, in that final solution is highly sensitive to the changes in the data. Consequently, the number of models which fits the data is increased by the level of the noise. In contrast to the cross-correlation measurements, that are not picked explicitly, the picking uncertainty in short-period traveltime is unavoidable, causing higher uncertainties.
2. most of the time the problem is under-determined, meaning that the number of unknowns is higher than that of the observations. Despite including multi-frequency measurements, many equations are dependent (Nolet, 2008).
3. the uneven distribution of sources and receivers is another reason that produces non-uniqueness. Particularly, adding contribution from similar sources and stations, may not necessarily lead to a better solution, increasing the number of dependent equations.

With regard to the above issues, the imposition of a priori constraint on the minimisation criterion is unavoidable. In this case, the inverse problem is solved by a search to find that part of the solution, which fits both the data and the penalty function. In this case, the solution can be formulated as

$$\underset{m}{\text{minimize}} \quad \|Gm - d\| \quad \text{subject to a constraint.} \quad (2.15)$$

The way that the regularisation is chosen influences the inversion results considerably. The traditional way of constraining the minimisation problem is to find the solution with minimum Euclidian length ($\|m\|$), the method which was also applied to the inversion in this study. Therefore, the optimisation problem in eq. (2.15) can be written as

$$\underset{m}{\text{minimize}} \quad \|Gm - d\| + \epsilon\|m\| + \eta\|Dm\| \quad (2.16)$$

or

$$\begin{bmatrix} K_V & K_S \\ \epsilon I & \epsilon I \\ \eta D & 0 \end{bmatrix} \begin{bmatrix} V \\ S \end{bmatrix} = \begin{bmatrix} dT & 0 & 0 \end{bmatrix} \quad (2.17)$$

where D is a smoothing operator. ϵ and η refer to damping and smoothing parameters. K_V is the sensitivity kernel for velocity model, and K_S is an ad hoc sub-matrix corresponding to the source corrections S (see section 4.1), included in the sensitivity kernel matrix.

Since the distribution of the data is not even, some part of the model in eq. (2.16) might be damped more than others, which can introduce a bias in the model towards an apparent structure. To deal with this effect, we introduce the Bayesian method into the equation by considering a prior knowledge for the model parameters. The prior knowledge here refers to a general idea about the deviation of model parameters from the background model, i.e. the covariance matrix C_m , the diagonal elements of which are the allowed variance in the model parameters. The off-diagonal elements can account for the possible smoothness between parameters.

Assuming that the probability for the data and model parameters are independent, eq. (2.16) can be modified using Bayesian theorem as

$$S(m) = (Gm - d)^T \cdot C_d^{-1}(Gm - d) + m^T \cdot C_m^{-1}m + \eta m^T D^T Dm \quad (2.18)$$

where C_d and C_m are priori covariance matrices for the data and unknown vectors. Since we do not have any idea about the correlation between stations, the data covariance matrix is considered diagonal with σ_i as elements. The diagonal elements of C_m are considered as the allowed values for the variation of the velocity parameters from the initial model, which is ~ 0.1 %. The off-diagonal element, which may be a subjective choice, here is considered to be the correlation between model parameters, i.e. smoothness. To solve this system, we use LSQR (Paige and Saunders, 1982), a variant of the conjugate gradient method, minimising both data fit and model norm terms in the eq. (2.16), which is equivalent to

$$S(m) = \chi^2 + m^T \cdot C_m^{-1}m \quad (2.19)$$

where the 'chi-square', χ^2 , which is a statistical term used to obtain the data misfit, is defined as follows

$$\chi^2 = \frac{1}{N} \sum_{i=1}^N \left(\frac{G_{ij}m_j - d_i}{\sigma_i} \right)^2 \quad (2.20)$$

in which, N is the number of the data and σ_i refers to the standard deviation in datum i . The maximum likelihood solution, i.e. the model with the highest probability for its predicted data vector, is associated with the minimisation of the 'chi-square', provided that the data errors are uncorrelated and normally distributed. The major disadvantage of chi-square is dependency on the normal distribution, which might not be always the case. One attempt to get closer to this assumption is to truncate outliers, the data which suffer from lower probability, i.e. larger deviation from the mean traveltime residual (Nolet, 2008). Here, this criterion was done by a trial inversion, after which the data $> 3\sigma$ were rejected, which formed 4 percent of the data.

If the measurements' uncertainty is well known, a good data fit has a $\chi^2 = 1$. Therefore, we looked for solutions with the closest misfit to one. Since we only have one observation per datum, I used some educated magnitudes of the errors, which were attributed to the data based on the cross-correlation coefficients. This means that the higher the cross-correlation was, the lower the error was considered. In the case of ISC-EHB data, I experimented with $\sigma_i = 1.0$ and $\sigma_i = 0.7$ seconds, which led to the inversion results with almost similar roughness, although with different χ^2 values. The solution which is obtained by minimising chi-square is called "least square solution", which involves minimising the data fit. However, equation 2.19 contains an extra term for the minimisation of the solution norm as well, which increases the data error. This trade-off between the two terms is controlled by the choice of damping and smoothing factors. Higher values of these parameters decrease the length of the solution, increasing the fitting error. Choosing the lower values of them minimises the data misfit, creating an under-regularised solution with higher norm. The optimal value of damping factors is usually selected by the L-curves, the method that I also used to navigate the model norms versus the data misfits χ^2 by experimenting various ratios of damping and smoothing. Considering a fixed $\frac{\epsilon}{\eta} = 0.3$, I have finally chosen a solution with the $\chi^2 = 1.152$. I preferred a solution with less order of the norm (roughness) for the interpretation purpose. This choice seems reasonable, since the uncertainty in the measurements of the standard errors might make χ^2 up to 20 % far from the ideal value of 1 (Nolet, 2008).

Figure 2.13 shows the trade-off curve and compares the preferred model with some other solutions at a selected depth.

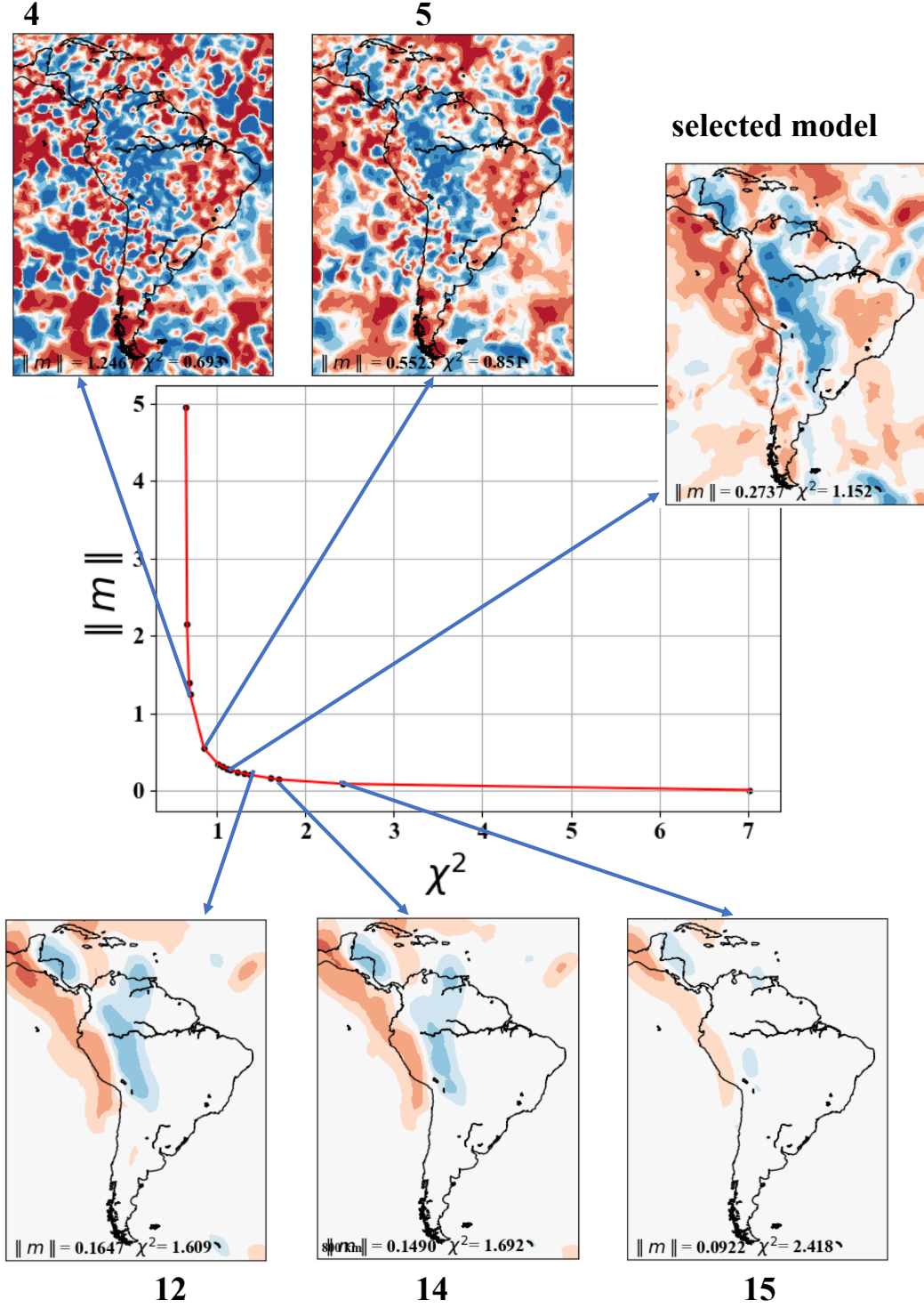


Figure 2.13: The trade-off curve for the real inversion with fixed $\frac{\epsilon_1}{\epsilon_2} = 0.3$. Black circles on the curve spot solutions with different values of ϵ_1 or ϵ_2 factors. Solution 4 and 5 at extreme ends are unwanted, since we fit the data error, leading to erratic features in the model, whereas we over-regularised the solutions at 15 and 14.

2.4.1 Source corrections

During the traveltime measurements, we assume that the source locations and origin times are known. Source parameters can be estimated reasonably from the stations nearby, provided that a suitable knowledge of the Earth is available. Source locations published by the National Earthquake Information Service (NEIS) or the International Seismological Centre (ISC) are usually estimated through 1D models such as IASP91 (Kennett and Engdahl, 1991) and AK135 (Kennett et al., 1995), in which shallow structures are biased towards a continental structure, as most of the stations are inland. In addition, source locations published by these agencies are traditionally measured using hand-picked onset point of seismic waves, representing the beginning of the rupture. This is not consistent with cross-correlation traveltimes, which is measured from the part of the waves with the highest energy, hence originated from a later location. This miscalculation may lead to a 10-km error, even larger in oceanic sites, at each coordinate of the source and several seconds change in the origin time approximation (Nolet, 2008). Obviously, an inaccuracy in the location, hence the origin time, causes an error in the calculation of the traveltime shift.

One common strategy to correct this problem, is to include source parameters in the inversion as unknowns, incorporating a prior knowledge of uncertainties in the linear system of equation. A Δh -miscalculation in the hypocenter can produce a change in the traveltime as follows

$$dT = \Delta T_o + p_x \Delta x + p_y \Delta y + p_z \Delta z \quad (2.21)$$

where p refers to the slowness, and ΔT_o , Δx , Δy and Δz show the unknown error caused by miscalculation of the origin time and hypocentre coordinates. To invert these anomalies, for every earthquake we add four columns to each row in the kernels matrix, and four entities to the unknown vector correspondingly.

The correction was implemented heuristically. Starting with initial guess of 8 km for each component of the source location and 2 seconds for the origin time, I changed the corrections gradually.

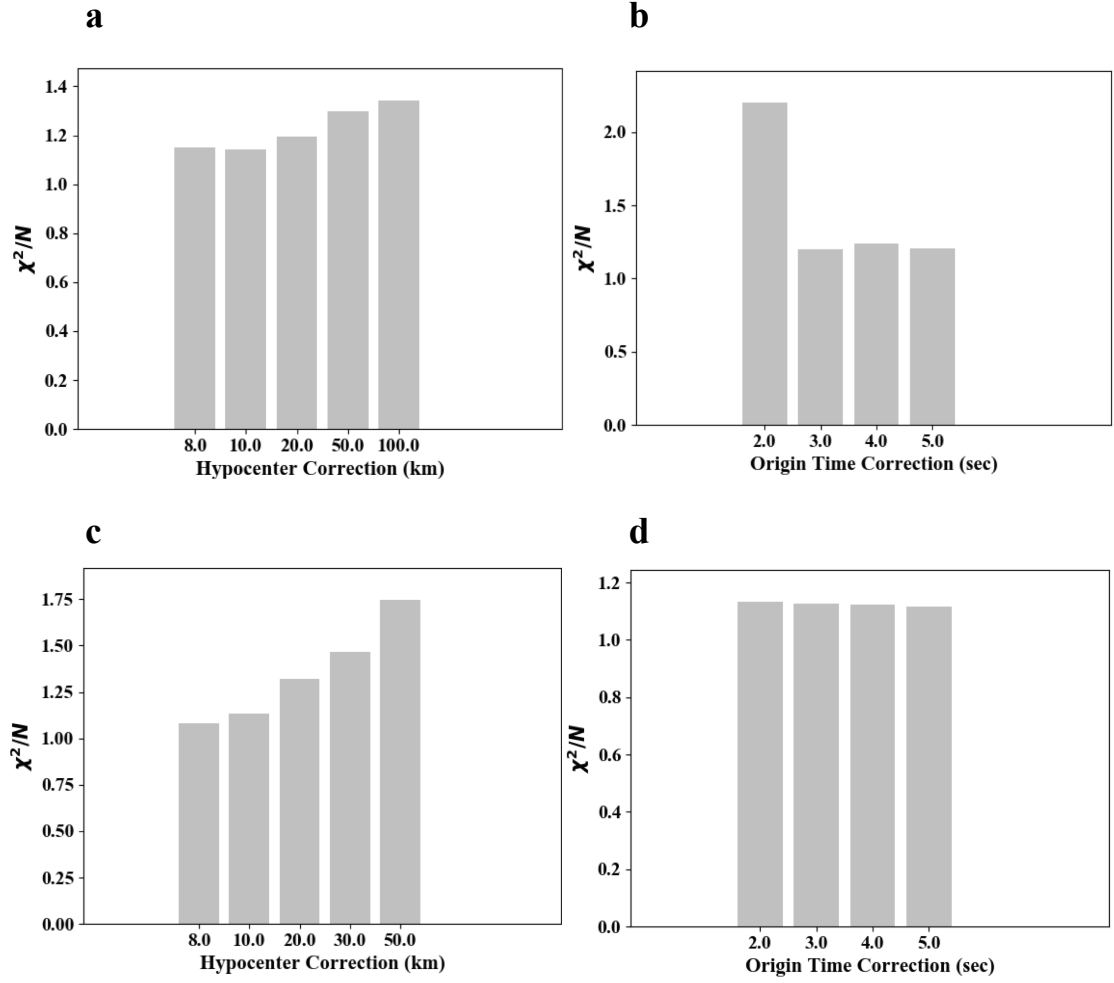


Figure 2.14: Comparison between Chi-squared values from the inversions of ISC (a, and b) and multi-frequency (c, d) data with different source parameters. The regularisation values for inversions at each dataset remains fixed, but differs between ISC and multi-frequency due to different noise levels in the two datasets. In both datasets, solutions with closest chi-squared values to 1.0 and smoothest (less roughness) have been chosen.

The values, which led to optimised solutions, were chosen as the final parameter. As I explained earlier, the preferred solution is considered as the solution with a chi-square value closer to 1.0 as well as less roughness. The source parameters for ISC-EHB and multi-frequency datasets were investigated separately.

Figure 2.14 shows comparison between the effects of the different source parameters. The inversion results of multi-frequency dataset seem not to be sensitive to the origin time range that I investigated, i.e. between 2.0 and 5.0 seconds. Whereas in the ISC-EHB case, the best result was achieved with the origin time correction

of 3.0 seconds. Source location corrections have a stronger influence on the final results of both trials, in which 10 km correction per each coordinate of the source location led to a sufficient output. Considering the results of these experiments, I concluded that 10 km correction per each component of the source location, and 3.0 seconds for the origin time uncertainty can be a reasonable choice.

2.5 Assessment of the inversion results

The last, but most important, part in the computation of any tomography is to judge the uncertainty of the final solution. To do so, one of the earlier techniques has been computation of the Resolution matrix, which is defined as

$$R = G^{-g}G \quad (2.22)$$

where G^{-g} refers to the “generalised inverse”, which serves an inverse in some sense for the rectangular G matrix. To recognise the effect of the resolution matrix, provided that the data is error-free and the parametrisation is perfect, the final solution error is defined as

$$m^{pre} - m^{true} = G^{-g}d - m^{true} = (G^{-g}G - I)m^{true}, \quad (2.23)$$

$$m^{pre} = Rm^{true} \quad (2.24)$$

The eq. (2.24) implies that the estimated model is called true if the $\sum_j R_{ij} = 1$. However, in practice, resolution matrix reflects both the bias and model resolution. As a result, it is always < 1 . This is one of the shortcomings of the Resolution matrix. In addition, in global tomography, computation of the Resolution matrix is too expensive.

Another important limitation of this method is that the identity Resolution matrix is possible only if the problem is over-determined, in which case the G matrix has the left inverse and $G^{-g}G = I$. Otherwise, the computation of the Resolution matrix does not lead to an identity matrix when the problem is under-determined,

even if the data is noise-free.

A quick way of estimating the influence of damping on the solution is by implementing synthetic tests, which is the most common method. In these tests, an input model is designed, through which artificial data is calculated. The synthetic data, which is contaminated to a certain level of random noise, is then inverted with the same regularisation as one used in the real problem. By comparing the output and input models, one can inspect the reliability of the recovered model. Checkerboard-like patterns, in which the model is defined by a positive and negative pattern of anomalies, are common features employed as the input model. The synthetic reconstructions are criticised for different reasons, including 1) the simplified noise added to the data cannot be representative of the actual noise, 2) using the same forward method to create the synthetic data and the inversion will ignore the possible error in the computation of forward method (Rawlinson et al., 2014).

Although resolution tests are not the most proper tools, they are the only means of analysing the resolution in our large-scale tomography. To this goal, I performed several synthetic tests using equally spaced Gaussian balls, in which the amplitude of the anomaly decreases through a Gaussian curve. In order to examine the effect of the noise, different levels of random noise ranging from (0.3 to 0.5 ms) was added to the theoretical data produced by finite-frequency methods in 4 frequency bands through the synthetic model. The resolving power of the inversion is estimated by $\frac{1}{e}$ width of the Gaussian ball in km, further away than which we assume the amplitude of the anomaly is negligible. The resolving length in my study, i.e. $\frac{1}{e}$ width of the ball, has found to be 200 km at all depths. This is the minimum length scale I recovered the Gaussian pattern with a suitable resolution, with 0.35 sec noise added to the synthetic data. Figures 2.14 and 2.15 illustrate resolution tests for 0.35 s noise with the ball diameters 200.

The results show a good agreement between the site of recovered Gaussian balls and that of real anomalies. In shallower depths < 600 km, the balls are not well recovered under northern Brazil, because of the sparse sampling caused by the lack of seismograms, and seismicity in the region. In contrast, in deeper depths, the

synthetic balls in the majority of depths are sharply retrieved, due to a denser data coverage therein. However, at 1500 km depth, the Gaussian pattern is smeared north-westward at north-western latitudes, reflecting the fact that most of the waves are lined up in this direction. It can be observed in figs. 2.11 and 2.12, where the kernel density at different depths are displayed. Thus, we will take this effect into account in our interpretation, as similar smearing effect is apparent in the real images at 1500 km depth. A north-eastern smearing under eastern longitudes of Brazil can be observed at and below 800 km depth, which does not affect our interpretation, since no significant anomaly is recovered at this region.

Overall, the synthetic tests indicate our inversion is capable of imaging structures under the north-western and western margins, where mantle structures are located in the real model. In the next chapters, I present the preferred models and give a detail explanation of the structures and their connection to the surface geology.

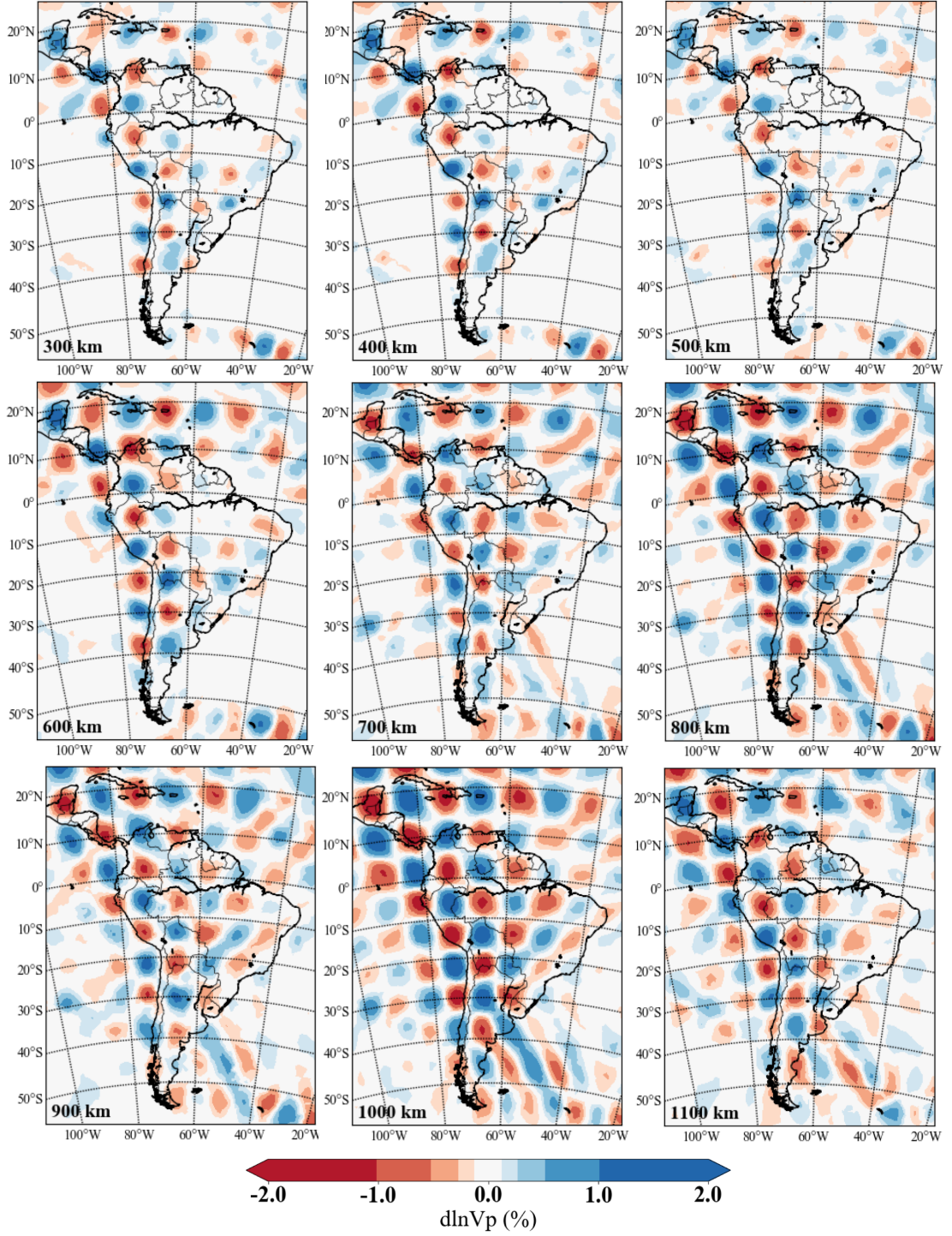


Figure 2.15: Lateral resolution tests between 300 and 1100 km depths. At each map the input patterns are Gaussian balls of 200 km diameter. Recovery is very good under the western part of the continent but deteriorates towards the east and into the oceans due to a paucity of seismic stations in these regions.

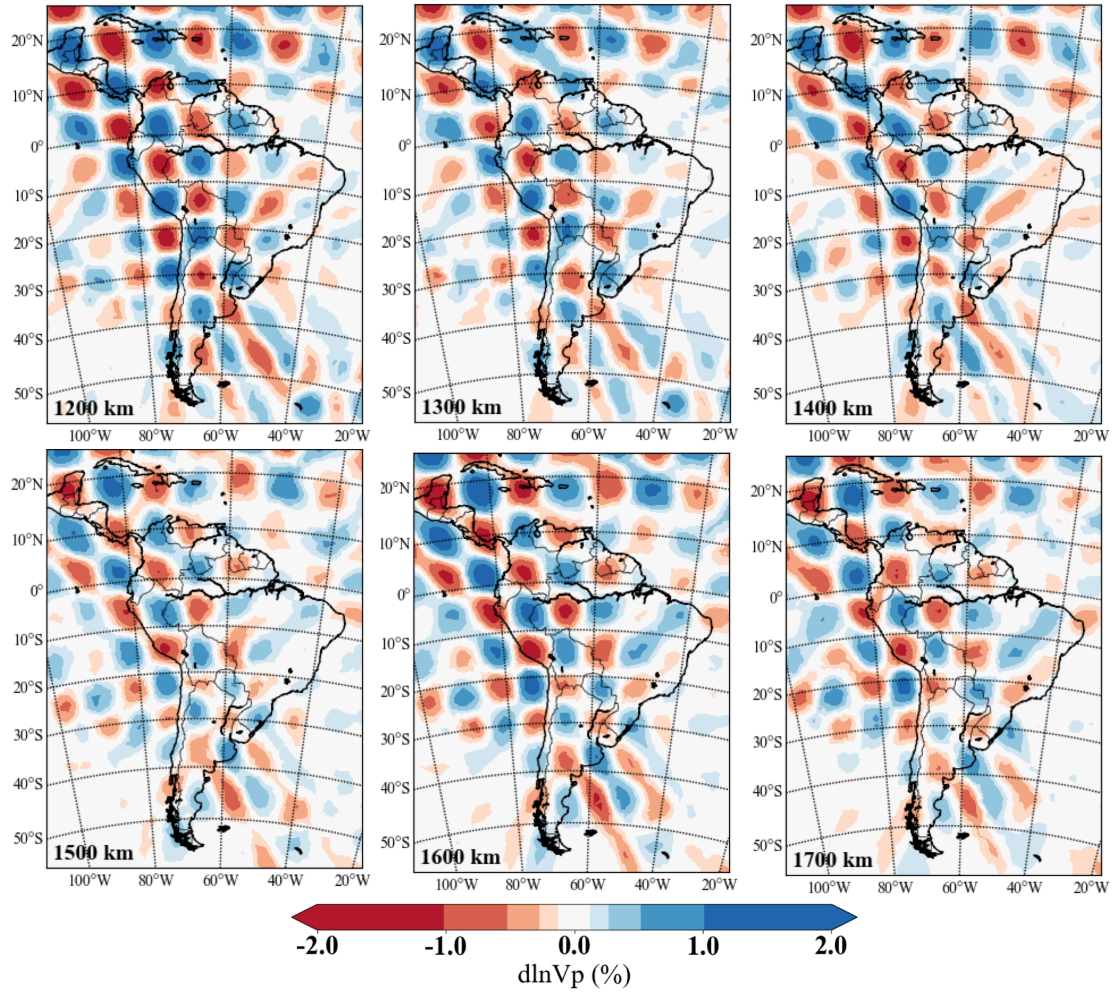


Figure 2.16: Lateral resolution tests between 1200 and 1700 km depths. At each map the input pattern is equally spaced Gaussian balls with 200 km diameter. The spheres are retrieved well in deeper depths, due to the higher kernel density.

3

Tomographic results

This chapter presents the preferred solutions of the current tomography under South America. Presenting the selected model at every 50 km depth, the following section explains the geometry of the mantle structures. In section 3.2, tomographic image of flat subduction, which is one of the significant features in South America, are presented. Finally, the preferred model is compared with previous P-wave tomographic studies in section 3.3.

3.1 Geometry of mantle structures under South America

2D map views of preferred P-wave velocity model are shown in Figures 3.1 to 3.5 in depth increments of 50 km. Seismically fast anomalies (shaded blue in figs. 3.1 to 3.5) fall into four main categories according to the variations in their depth, strike and the latitude. These slab units are explained from deep to shallow in this section.

Northern latitudes ($\sim 9^\circ$ S to $\sim 9^\circ$ N) are dominated by a complex seismically fast anomaly from ~ 1800 to 1200 km depth, while other mantle regions beneath the continent are slab-free in this depth range. This unit comprises two anomalies, including the north-eastern trending MEZ, and the north-south striking Fa anomalies.

Starting at ~ 1800 km depth along the Peruvian Andes at latitude $\sim 6^\circ$ S, the west-dipping slab (MEZ) stretches to higher latitudes in Caribbean plate. Almost at the same depth an east-dipping slab (Fa) joins to the east of MEZ. The two slabs display a north-western trend at 1500 km. However, it seems to be a smoothing effect, as it was evident in the resolution tests at this depth. The top end of the two slabs are connected between 1500 and 1200 km, displaying a horizontal subduction from the west to the east (figs. 3.4 and 3.5).

Higher up in the mantle, the second slab category appears between depths of 1200 and ~ 900 km. Figure 3.3 shows this subdivision, in which the western MEZ is still evident, although receding upward. While Fa slab is piling almost vertically along a NW-SE striking trench between latitudes 7° N and $\sim 15^\circ$ S, a NE-striking anomaly (GAC) appears northeast of it at ~ 1200 km. Migrating north-eastward, the GAC slab is becoming sharper by retreating MEZ subduction zone, which is truncated at 900 km depth.

The third partition contains a single Fa anomaly, which is observed to be the shallower part of the Fa slab discussed earlier. It is visible from ~ 900 to ~ 600 km depth, stretching a long distance (~ 3800 km) along its NW-SE trend from 5° N to 28° S latitudes. Figures 3.2 and 3.6a indicates how the steep Fa slab wall begins an east-dipping subduction at ~ 950 km depth, which lasts up to the surface. The slab's rollback coincides with northward migration of west-dipping GAC anomaly. No connection between the Fa and GAC slabs is evident from depth ~ 900 km upward.

The fourth slab units appears from 600 km depth to the surface. In contrast to the above segments, this division covers a wider area along Andes, containing Fa slab between $\sim 4^\circ$ S and $\sim 35^\circ$ S latitudes, which is intersected by a large gap (P). Appearing under Peruvian latitudes ($\sim 9^\circ$ S to 1° N), the (P) gap is visible up to the surface. However, due to the lack of seismic stations in northern Peru, the shallower part of the gap (< 300 km) seems to be poorly recovered. Recent surface studies (e.g. Schaeffer and Lebedev, 2013) have demonstrated high velocity anomalies in this region. Thus, my interpretation in Chapter 4 is restricted to

the presence of the gap between 300 and 600 km depths. An indentation (JFP) is observed along Fa anomaly under Central Chile. This fragmented slab gap occupies the area between latitude $\sim 29^\circ$ S and 19° S at ~ 650 km depth. Narrowing towards the surface, it is observed under Chile at $\sim 27^\circ$ S latitude. Map views at fig. 3.1 demonstrates this feature. The split of the Fa anomaly by the P gap resulted in a NE oriented slab (SB) under Colombia, which is linked to a NE-trending anomaly underlying western Venezuela (Ca) above ~ 600 km. From the tomographic images in fig. 3.1, one might infer an overlap between the two anomalies, which dip in the direction of south-east. However, it is caused by the projection of the 3D anomaly on the map view. The SB anomaly is connected to the Fa anomaly under ~ 600 km depth. In cross-section views at fig. 3.6b, no link between SB and Ca is visible. The last structure is a significant band of slow materials parallel to the Fa anomaly between ~ 900 and ~ 300 km. This broad province underlies several bathymetric highs in the eastern Pacific including Cocos, Carnegie, and Nazca ridges. The Carnegie and Nazca are aseismic ridges on the Nazca plates that are being subducting under South America. The large scale of the anomaly, which is commonly recovered in seismic tomography of South America, was linked to the fast rate of the spreading of the Pacific ridges, leading to the younger, hence buoyant, lithosphere by Heintz et al. (2005).

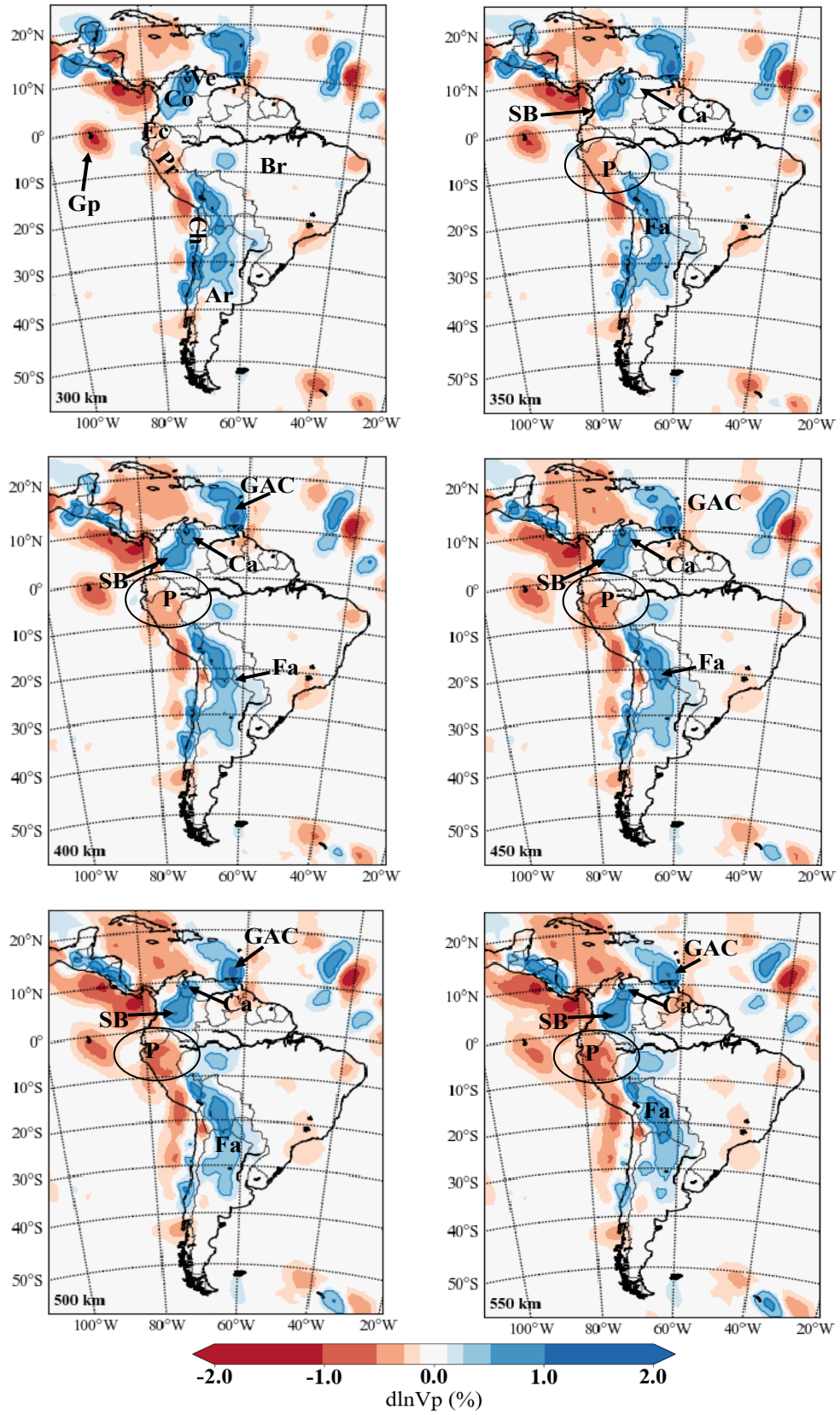


Figure 3.1: Map view of V_p anomaly at selected depths between 300 and 550 km. Ar: Argentina, Br: Brazil, Ch: Chile, Co: Colombia, Ec: Ecuador, Gp: Galápagos, Pr: Prue, Ve: Venezuela.

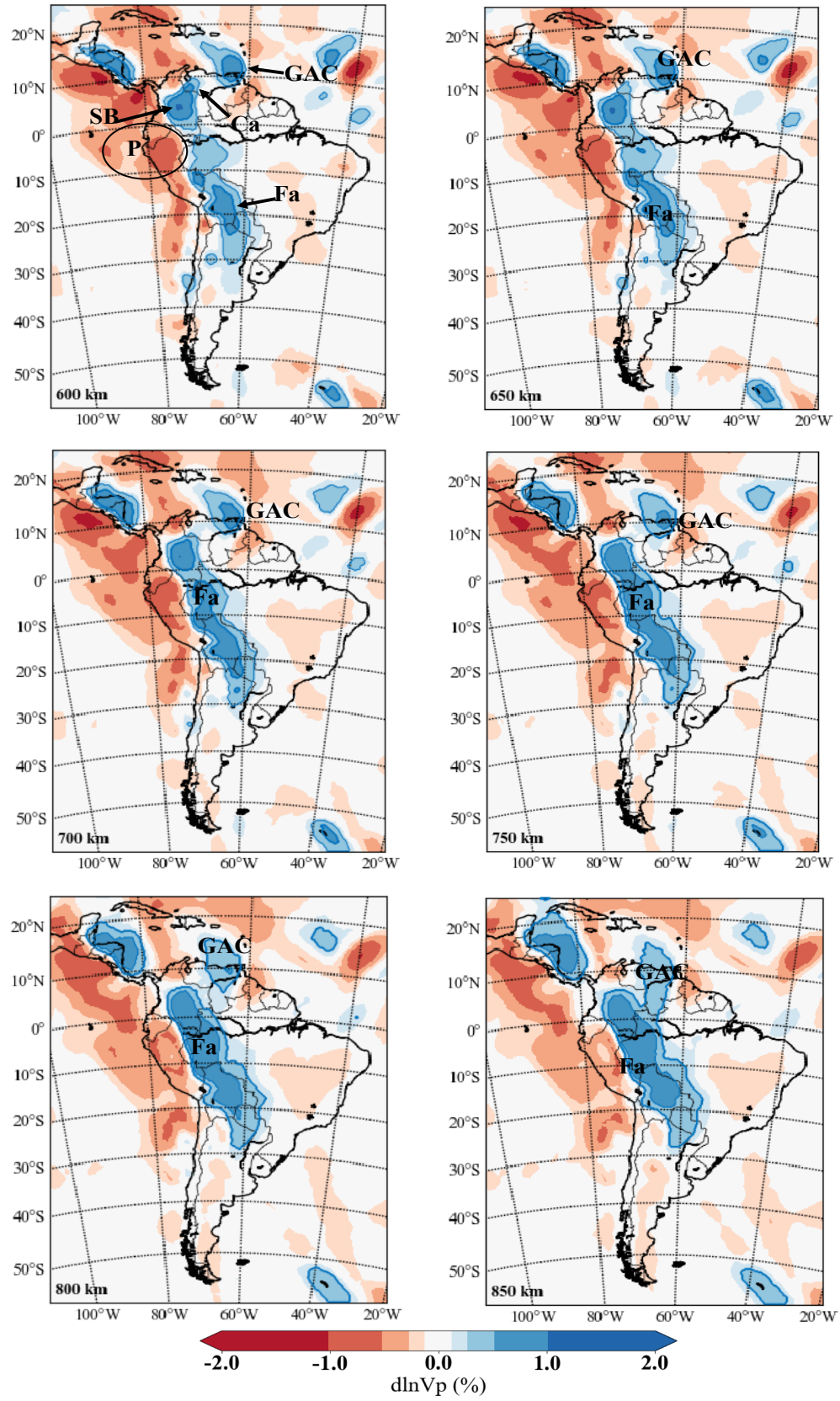


Figure 3.2: Map view of V_p anomaly at selected depths between 600 and 850 km.

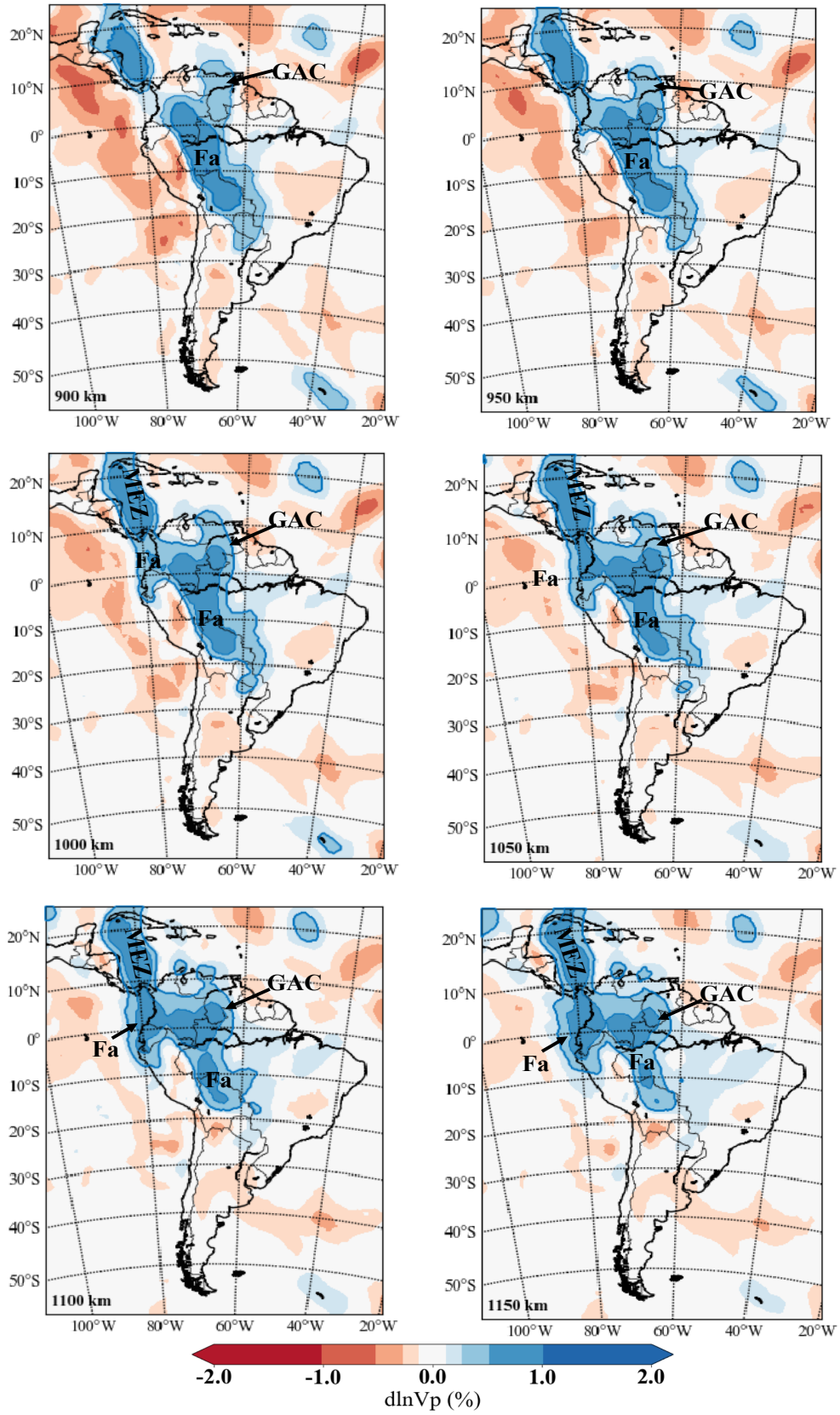


Figure 3.3: Map view of V_p anomaly at selected depths between 900 and 1150 km.

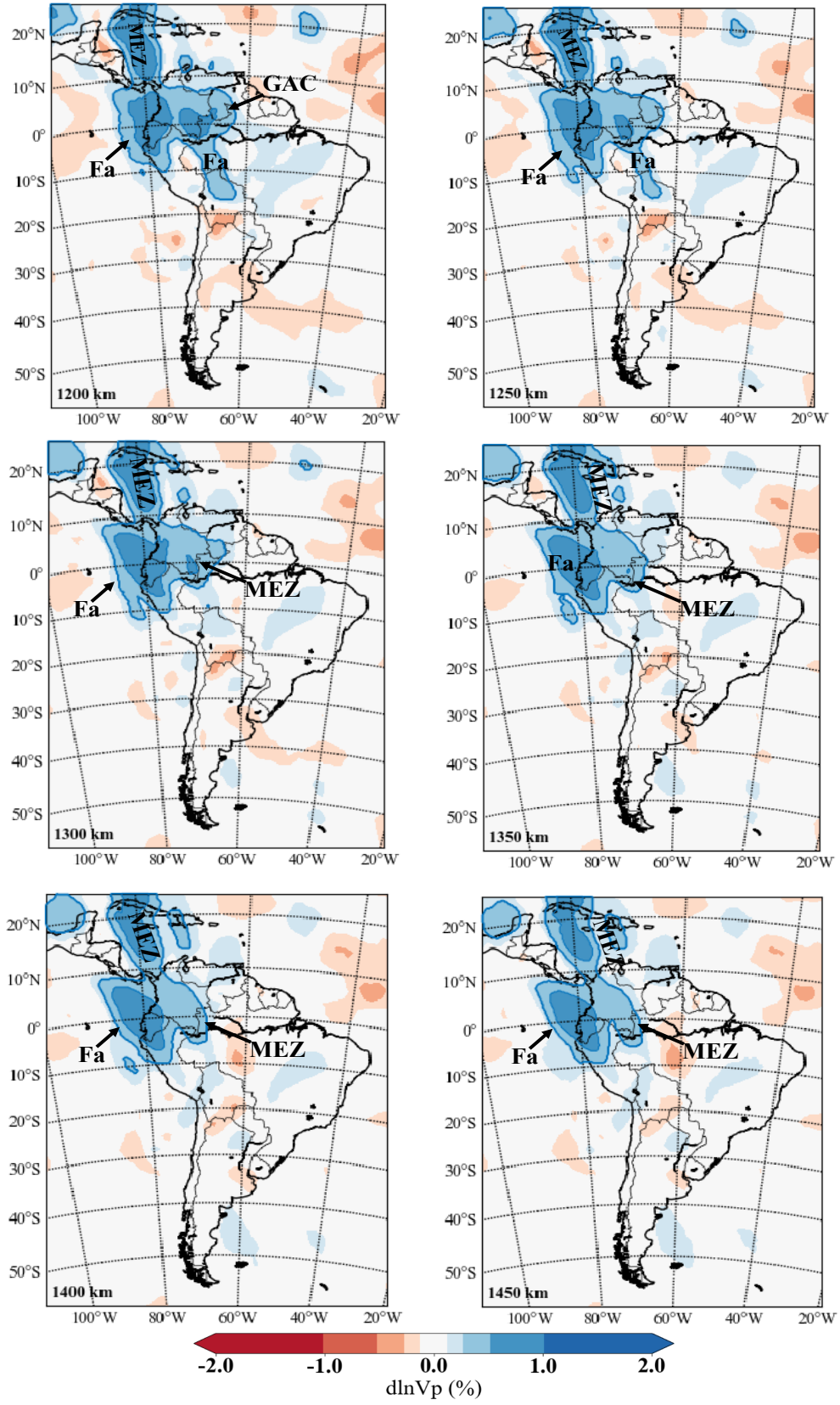


Figure 3.4: Map view of V_p anomaly at selected depths between 1200 and 1450 km.

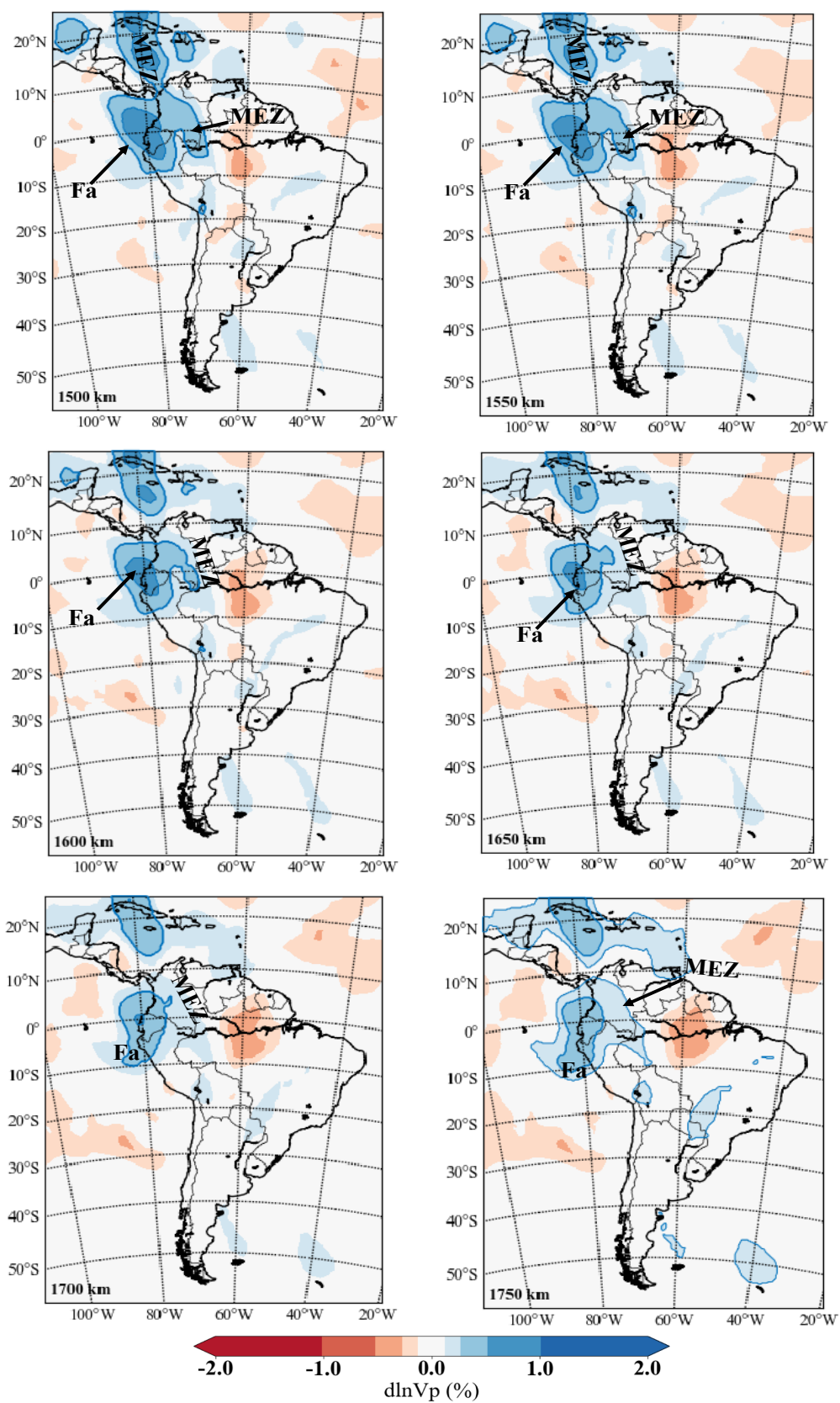


Figure 3.5: Map view of V_p anomaly at selected depths between 1500 and 1750 km.

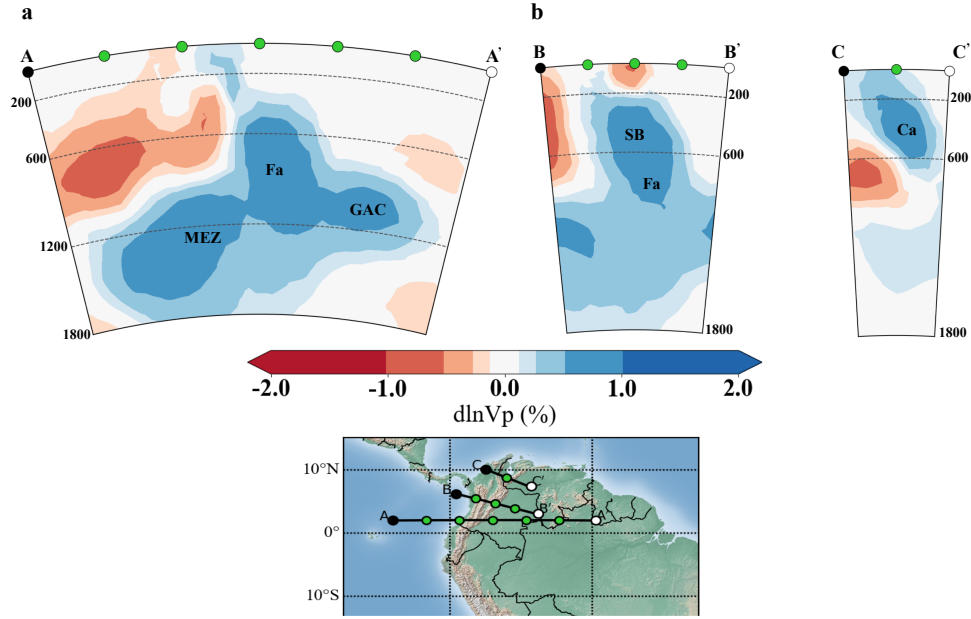


Figure 3.6: a) AA' cross-section of MEZ, Fa, and GAC slabs. b) BB' indicates SB connected to the north-western tail of Fa after 600 km depth, CC' section reveals the SE-dipping Ca slab.

3.2 Flat versus steep subductions

Flat subduction zones are region, where oceanic lithosphere subducts horizontally under the overriding plate. They are associated with the absence of related Quaternary volcanism as well as deformation in the overriding plates (Espurt et al., 2008; Martinod et al., 2010; Gutscher et al., 1999). Regarding the spatial correlation between the Andean deformation and the variations in the subduction dip, in this section I present tomographic signatures of two important horizontal subduction systems under Peru and Chile.

3.2.1 Peruvian flat slab

One of the largest examples of horizontal subduction is located under Peru between the 5° S and 14° S latitudes (Barazangi and Isacks, 1976), where it occupies as far as 700 km inland (van Hunen et al., 2002). Wadati-Benioff zone in this region shows that the depth of earthquakes increase gradually from west to east at an angle of $\sim 30^\circ$ to a depth of ~ 100 km where it levels out for several hundred kilometres, before steepening again (Manea et al., 2017). Figure 3.7 displays cross-sections

perpendicular to the trench along the Peruvian Flat slab. To have a better inference of the slab geometry, seismic sources from EHB-ISC catalogue (between 1967-2014) are overlaid on the cross-sections. As it is visible, the spatial distribution of the horizontal seismicity overlaps with a region of no significant structures < 200 km depth. In deeper depths between 200 and 600 km, however, there is a remarkable slow anomaly underlying the flat seismicity. In higher latitudes, $\sim 10^\circ$ to $\sim 14^\circ$ S (fig. 3.7, cross-sections 1-4), the slow anomaly occupies almost a flat regions under the seismicity. Towards the southern latitudes, $\sim 17^\circ$ to $\sim 23^\circ$ S (fig. 3.7, cross-sections 5-8), the slow anomaly is still evident under the flat seismicity, although with a different geometry. It is tilted and almost vertical. stretching along the steep slab. The Fa slab overlaps with the steep seismicity (cross-sections 5-8), while the slow blanket in all the cross-sections in fig. 3.7 underlies the flat seismicity.

3.2.2 Pampean flat slab

Another region subjected to the flat subduction is the Pampean segment of central Chile between $\sim 27^\circ$ S and 33° S (Barazangi and Isacks, 1976). The subduction starts with a 15° dip down to depth ~ 100 km, where it flattens for ~ 300 km. This flat-slab is characterised by a double seismic zone located within the subducting Nazca slab at intermediate depths (50–200 km) and separated by a 20–25 km zone with low seismicity (Marot et al., 2013).

Figure 3.8 presents tomographic cross-sections at the corresponding latitudes. As the structures at the Chilean latitudes, obtained from the preferred model, show low amplitudes, Appendix.A displays the cross-sections from a less regularised inversion with $\chi^2 = 1.108$. Under central segment of the Central Andes (fig. 3.8, cross-sections 1-3), the steep seismicity parallel the high velocity anomaly. Cross-sections 4-8 indicates Chilean flat seismicity overlying the tomographic structures.

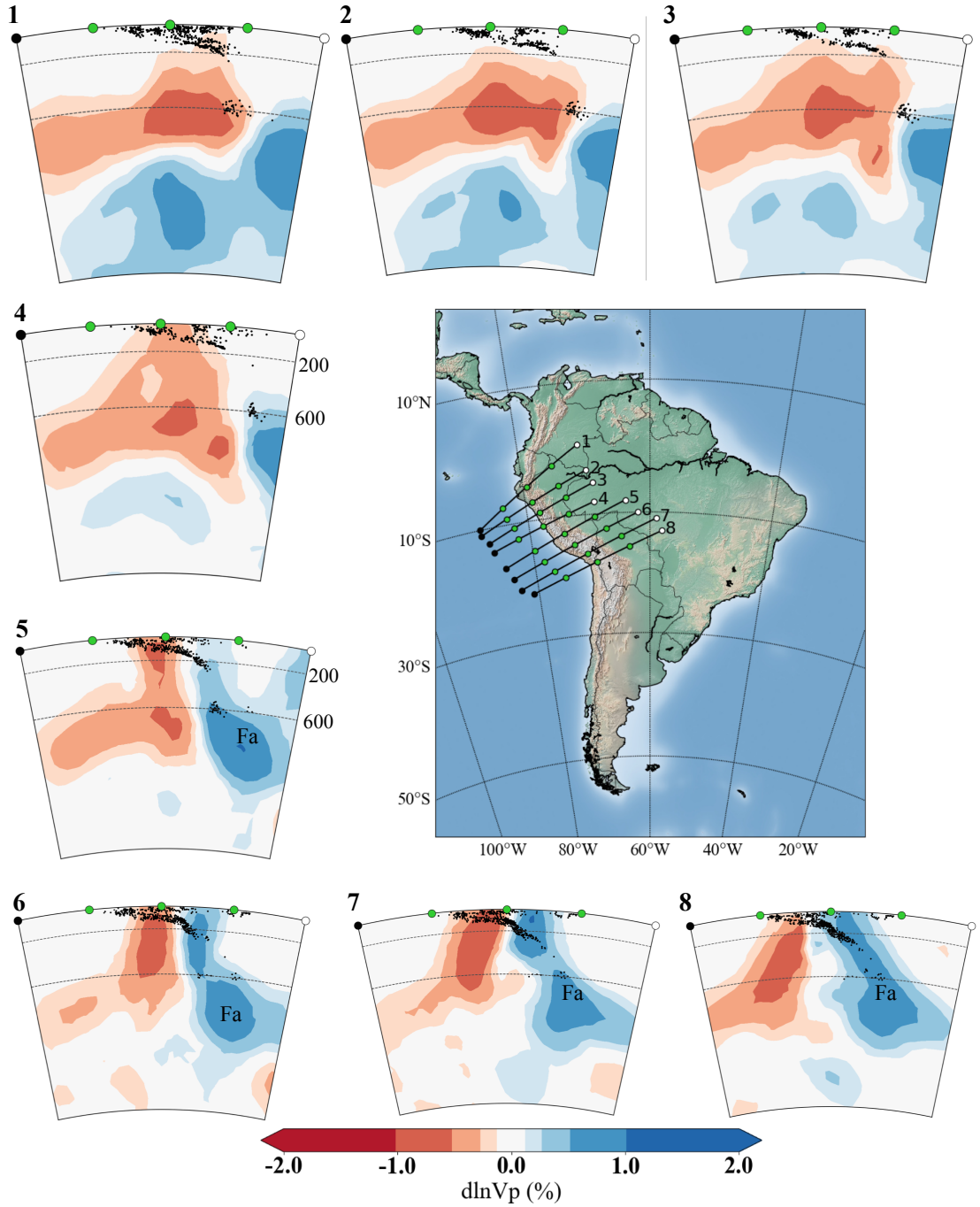


Figure 3.7: Cross-sections of the Peruvian flat subduction through the current model. Left: map view of South America, showing location of 8 cross-sections through tomographic model of South America. The dashed lines in all cross-sections indicates 200 (shallower line), and 600 km depths. Black circles represent location of seismic sources.

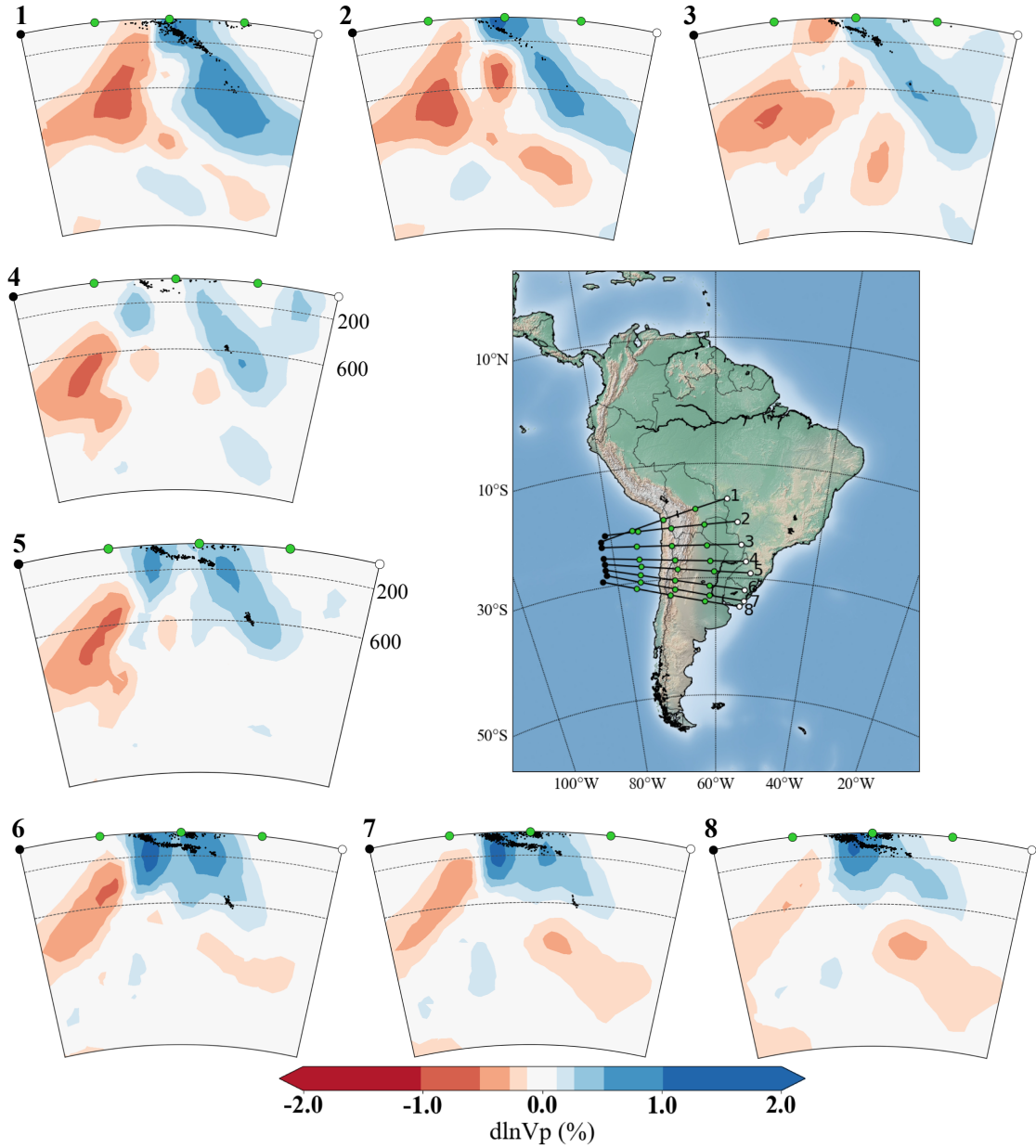


Figure 3.8: Cross-sections of the Chilean flat subduction through the current model. Left: map view of northern-Chile region, showing the location of 8 cross-sections through tomographic model of South America. The dashed lines in all cross-sections indicates 200 (shallower line), and 600 km depths. Black circles represent location of seismic sources.

3.3 Comparison with other models

Figures 3.9 to 3.20 compare the new model with previous studies. The most important function of this comparison is to ensure that the slab categories, which were explained above, are robust features to be interpreted. The images were created

using SubMachine online tools (Hosseini et al., 2018). Some general characteristics can be inferred from the figures. As it is evident, tomographic models by Hosseini (2016) seems under-regularised in the majority of depths. Whereas, the “GAP-P4” model by Obayashi et al. (2013) is too smooth. The smooth and large-scale anomalies recovered at shallow depths (< 600 km) in SPani-P (Tesoniero et al., 2015) might reflect the impact of long-period surface waves. The model seems under-regularised in deeper depths. MITP_USA_2016MAY model (Burdick et al., 2016) contains significantly low-amplitude at all depths, particularly in deeper depths (> 1200 km).

Figures 3.9 to 3.11 compare the first slab segment (1800 to 1200 km), which was explained in section 3.1, in different models at selected depths. High velocity anomalies of this segment were commonly imaged in the north-western regions by previous global P studies (e.g Simmons et al., 2012; Montelli et al., 2006). However, the geometry of the slabs is different at each study. In all models, the presence of the two Fa and MEZ anomalies are clearer in 1500 and 1300 km depths. One interesting feature in this unit is the presence of the fast anomalies in the eastern parts of the continent, which is evident in some models (e.g. Montelli et al., 2006; Amaru, 2007; Tesoniero et al., 2015). However, this anomaly is not constructed in the current model, probably due to the lower kernel density in the eastern regions. The second division of the slabs (from 1200 to 900 km) is compared in figs. 3.12 to 3.14. The southward propagation of Fa anomaly can be inferred from most of the studies, although with different amplitudes and thickness. In contrast GAC anomaly has been recovered sharply in fewer models (e.g. Montelli et al., 2006; Amaru, 2007). The disappearance of MEZ anomaly in this unit is evident in almost all models.

The comparison of the third slab partition (900 to 600 km) in different tomographic models can be observed in figs. 3.15 to 3.17. The single Fa anomaly is evident in all models, although its boundary is not as sharp as Fa segment in the current images. For example, the Fa anomaly in the study by Houser et al. (2008) seems too thickened and smoothed, while it is fragmented in the study by Hosseini (2016).

In the southern latitudes, my model does not show any robust fast feature, whereas Fa anomaly stretches southward in some models (e.g. Amaru, 2007; Obayashi et al., 2013; Simmons et al., 2012). The fourth category (600 to 300 km depth) is compared in figs. 3.18 to 3.20. It is clear that the southern segment of the Fa slab is curved in the majority of the models. The slow anomaly blanket, which is parallel to the Fa anomaly, can be observed in the tomographies by Montelli et al. (2006) and Amaru (2007).

The advantages of the present model over the previous studies, can be understood better by comparing cross-sections of the subducted slab in the western border, where is well sampled, hence well recovered in all models. Cross-sections in figs. 3.21 to 3.24 indicate how well flat and steep seismicity overlaps with the recovered slab in the current tomographic image, while the variation of the slab's dip is not recovered in most of the previous studies. Overall, my model agrees with the previous P-wave tomographies. However, the structures in the current model indicate sharper geometries, which enables me to confidently divide the slabs into sub-units. In the next chapters, I will interpret the four slab segments described in section 3.1 and explain how this slab distribution correlates with segmentation of the geological records along the strike of the Andean mountain belt.

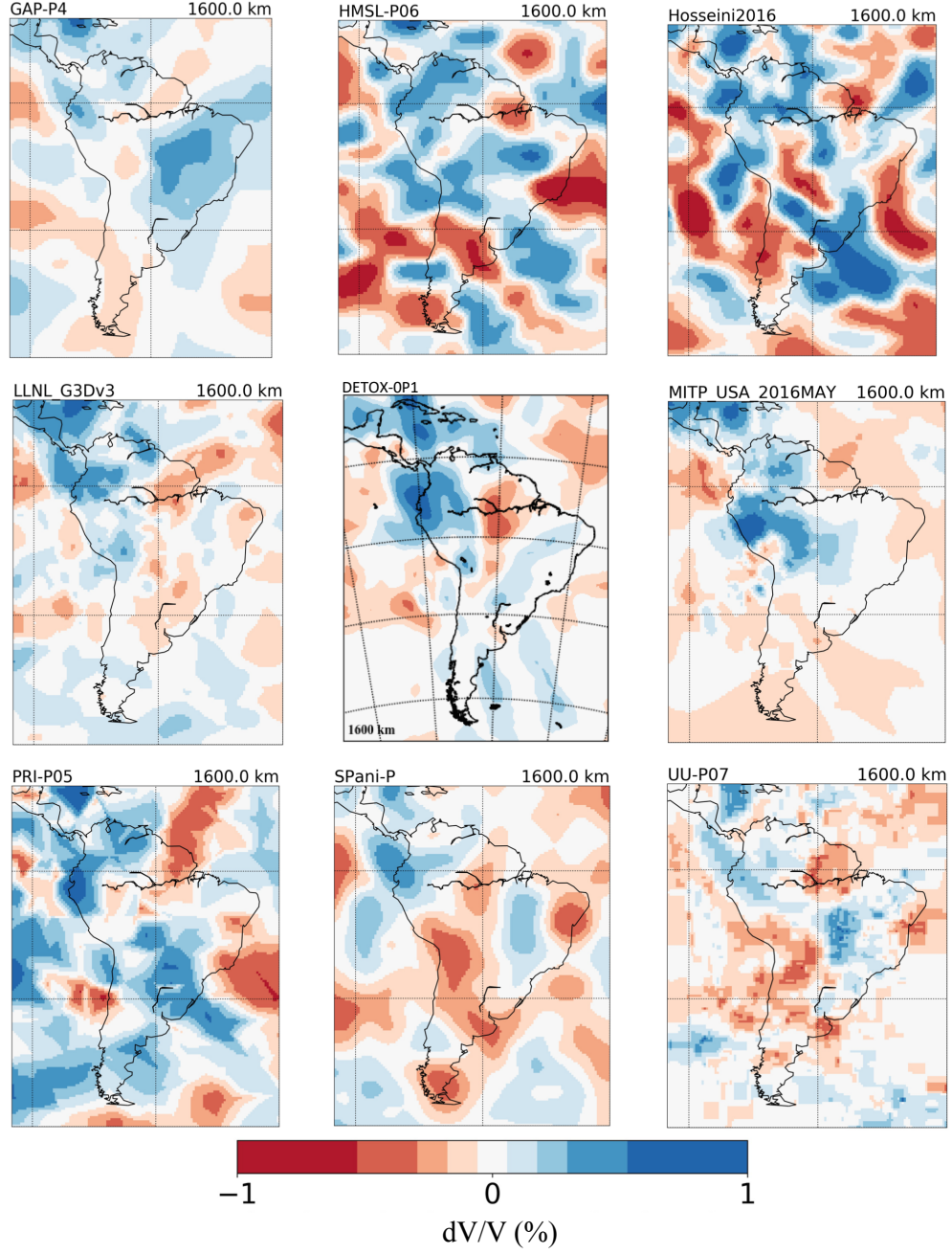


Figure 3.9: Comparison of the first slab unit in the different tomographies at 1600 km. DETOX-0P1 (current model), GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

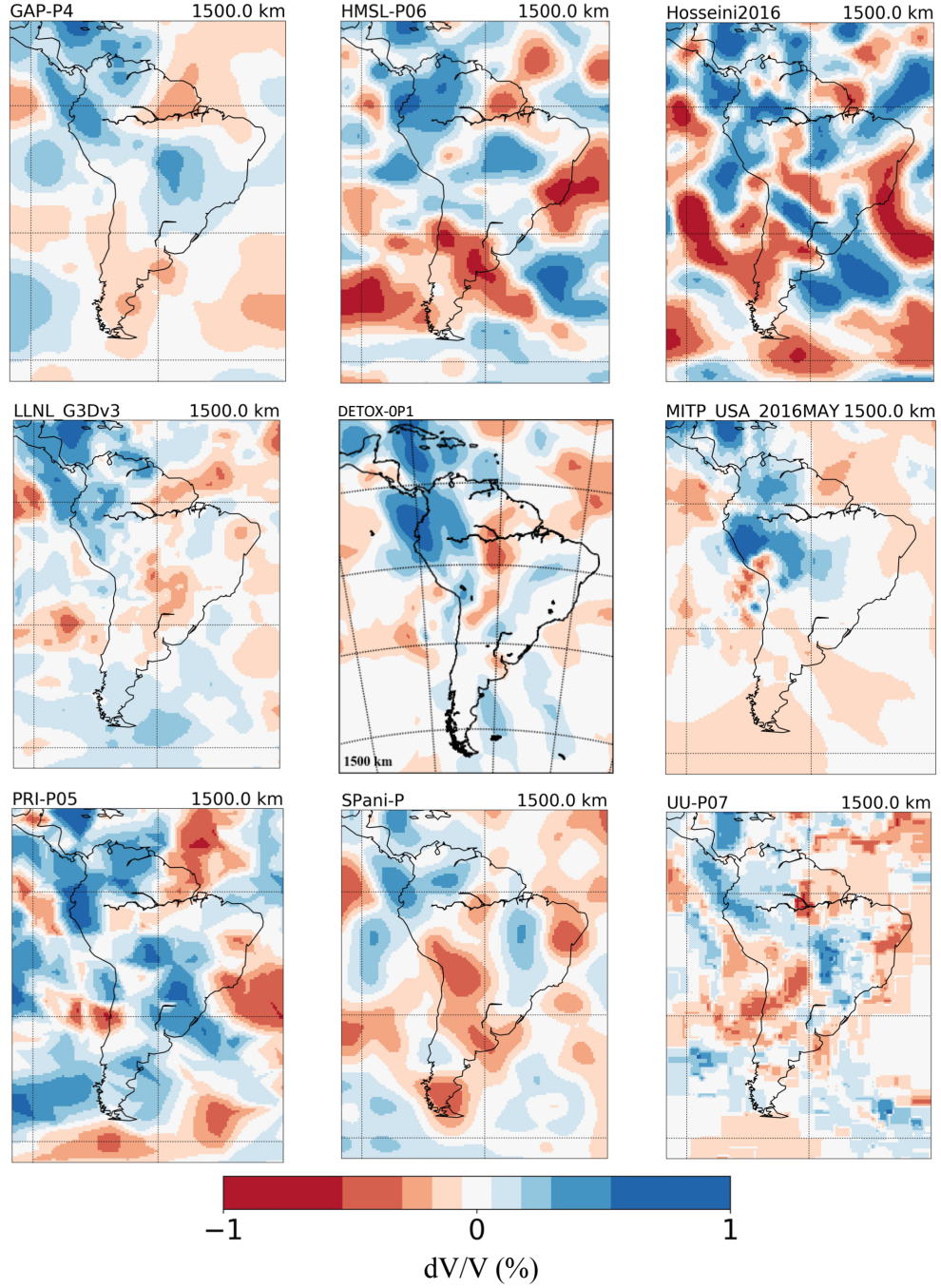


Figure 3.10: Comparison of the first slab's unit in the different tomographies at 1500 km. DETOX-0P1 (current model), GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

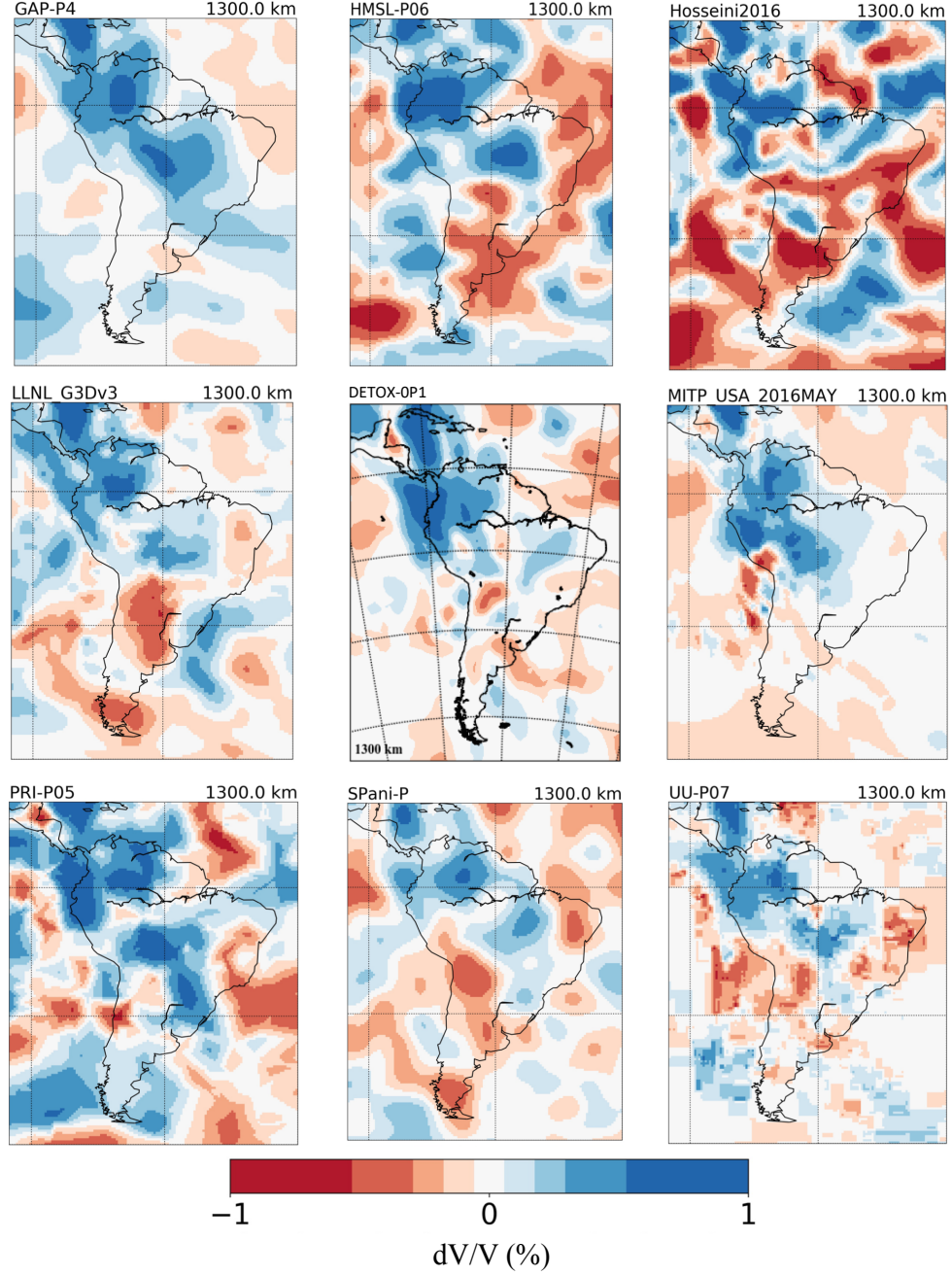


Figure 3.11: Comparison of the first slab's unit in the different tomographies at 1300 km. DETOX-0P1 (current model), GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

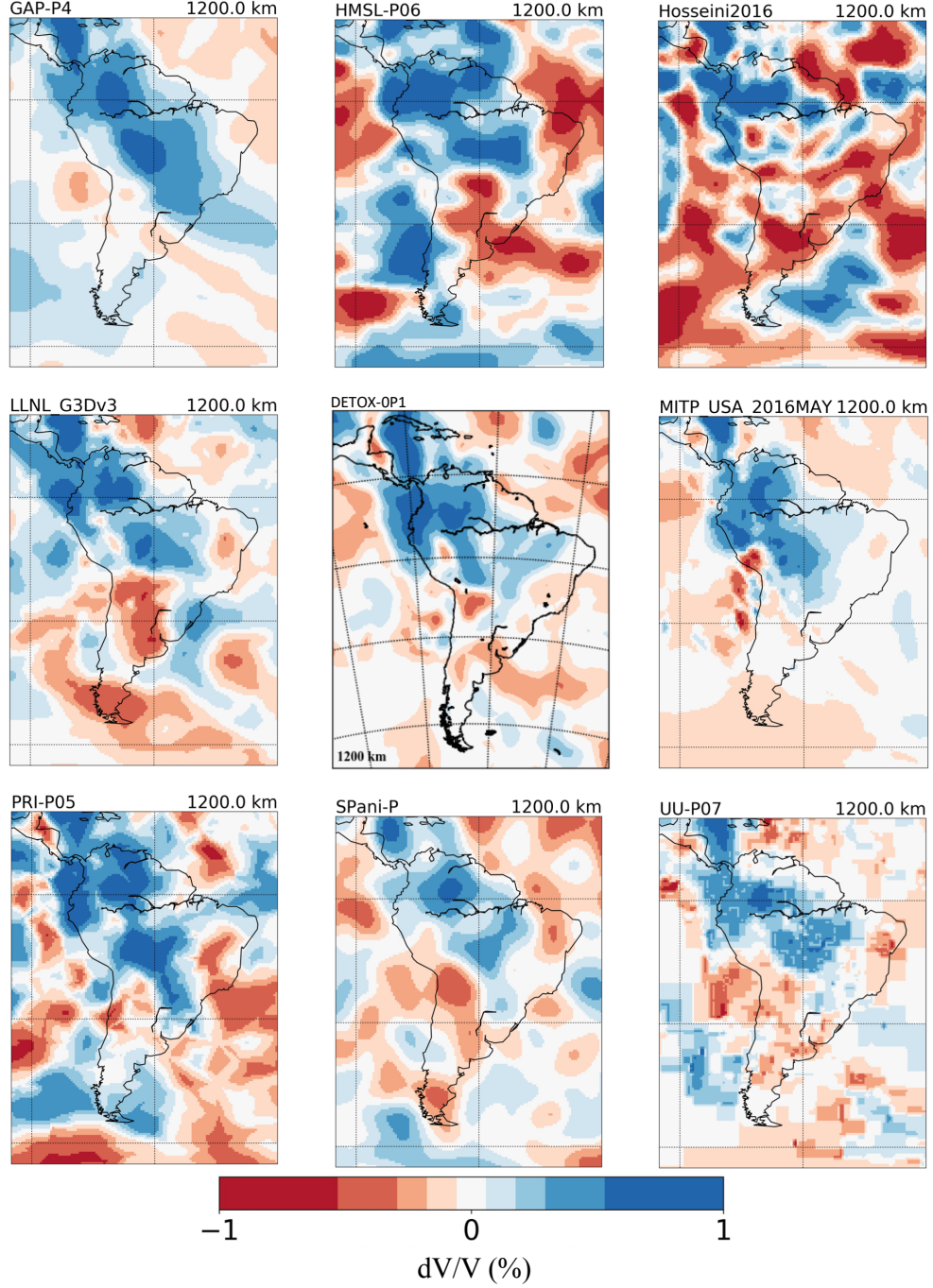


Figure 3.12: Comparison of the second slab segment in the different tomographies at 1200 km. DETOX-0P1 (current model), GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

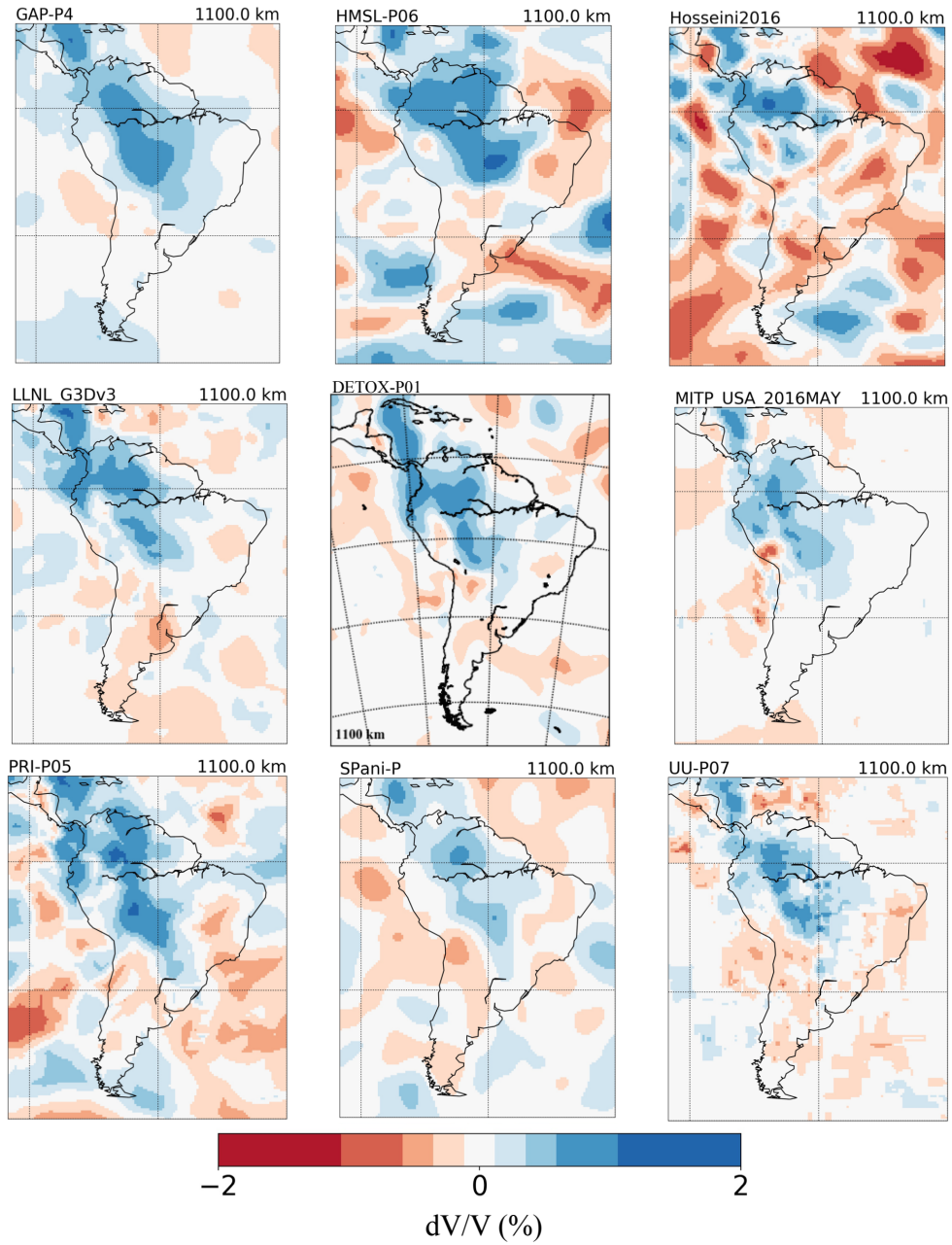


Figure 3.13: Comparison of the second slab segment in the different tomographies at 1100 km. DETOX-0P1 (current model), GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

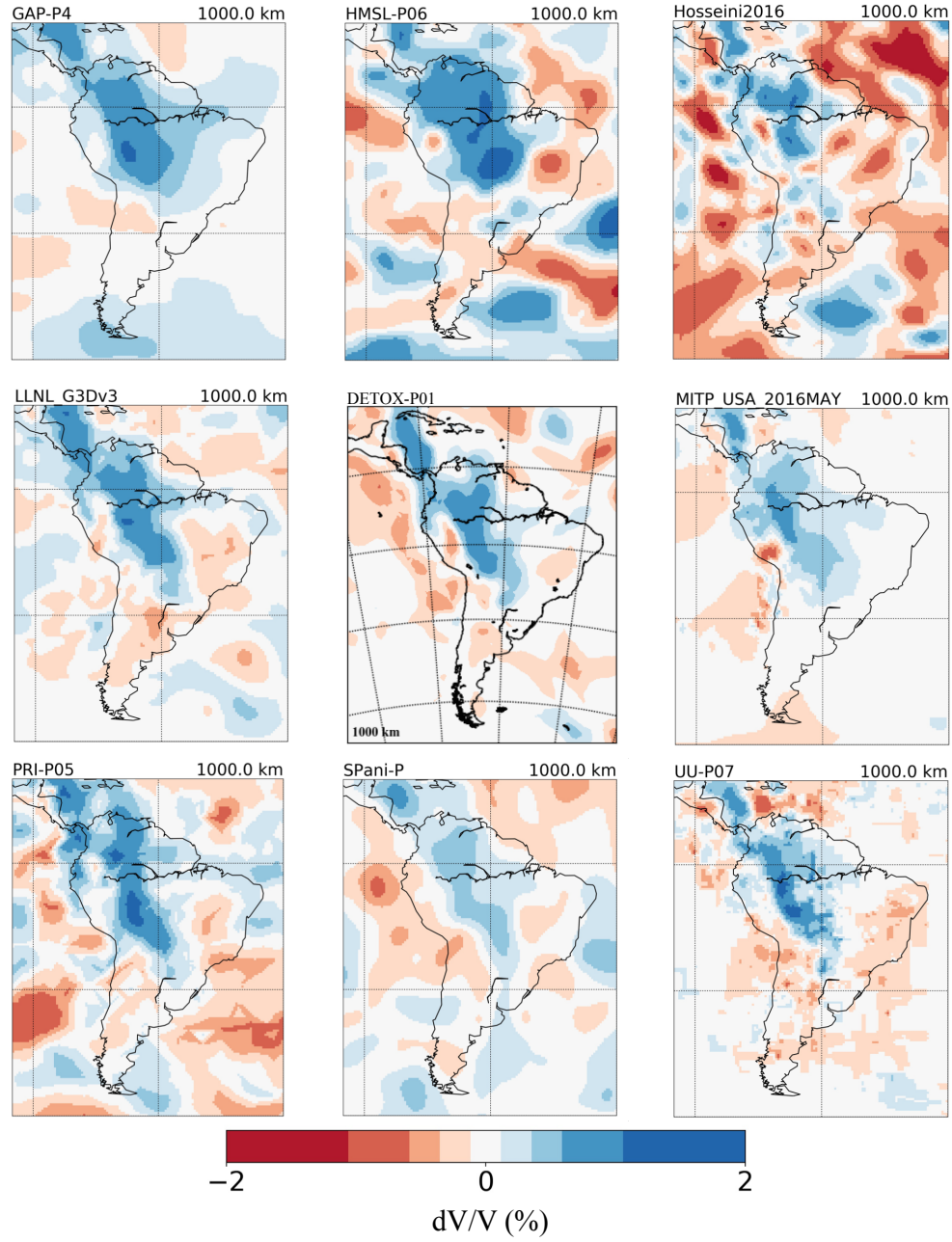


Figure 3.14: Comparison of the second slab segment in the different tomographies at 1000 km. DETOX-0P1 (current model), GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

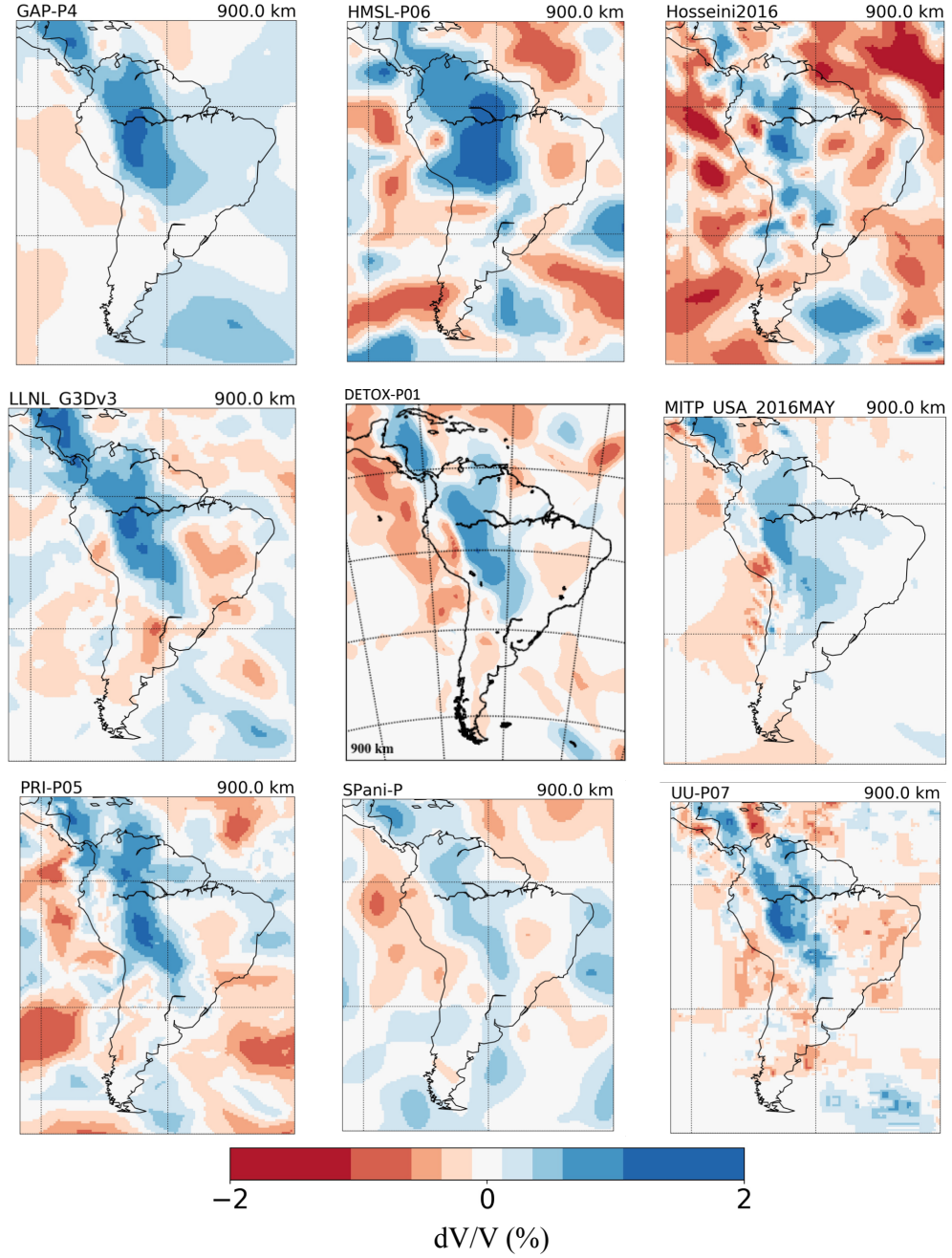


Figure 3.15: Comparison of the third slab division in the different tomographies at 900 km. DETOX-0P1 (current model), GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

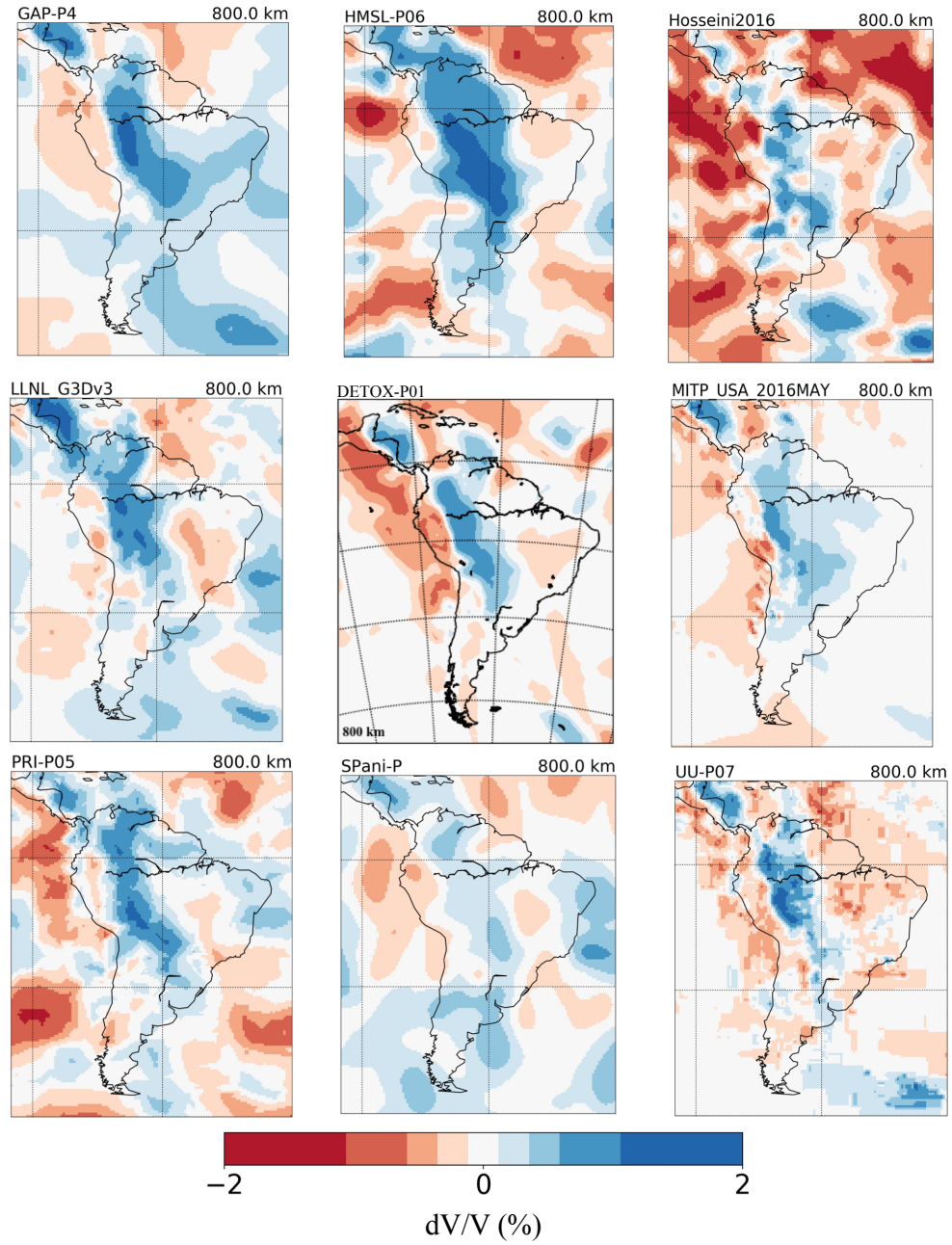


Figure 3.16: Comparison of the third slab division in the different tomographies at 800 km. DETOX-0P1 (current model), GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

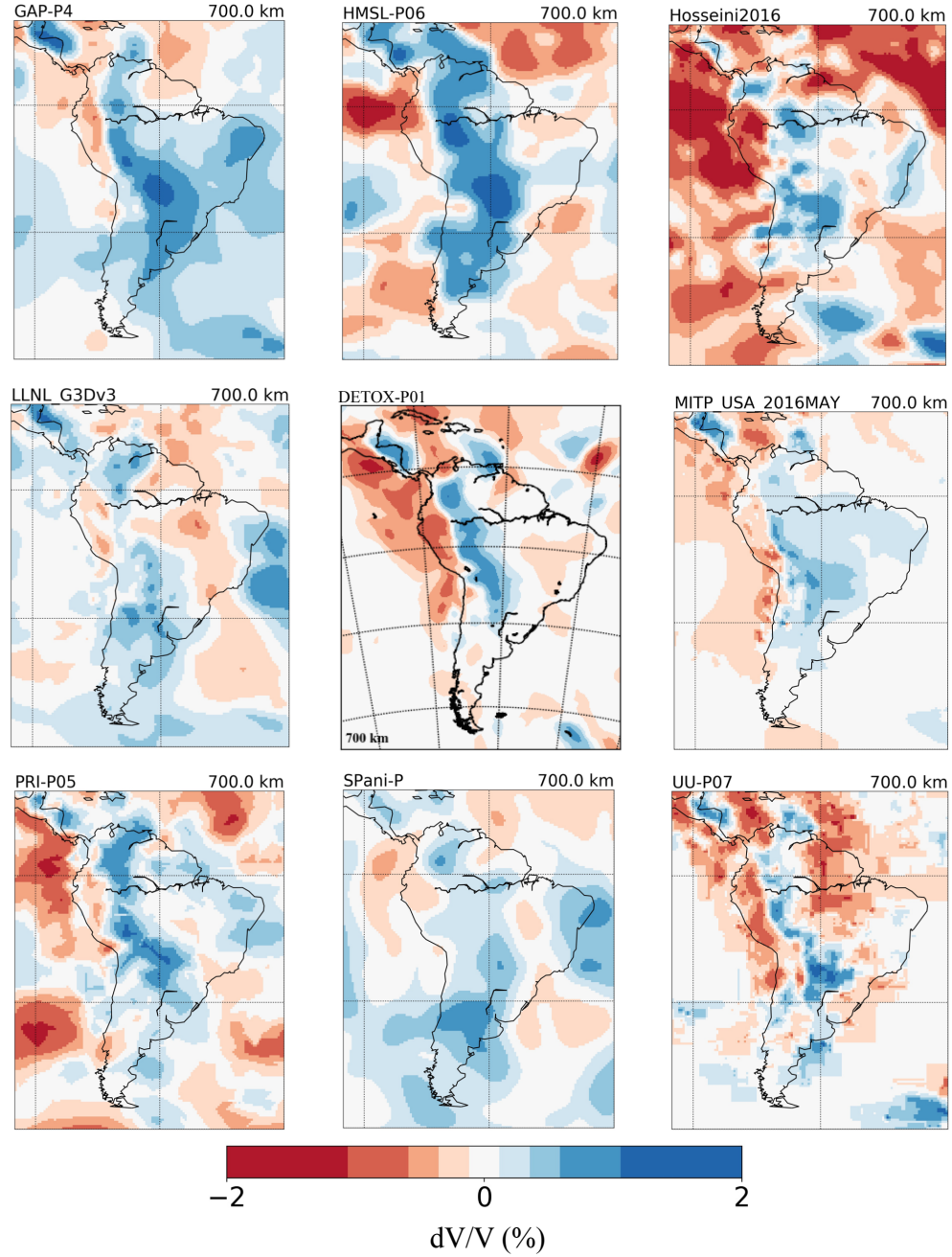


Figure 3.17: Comparison of the third slab division in the different tomographies at 700 km. DETOX-0P1 (current model), GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

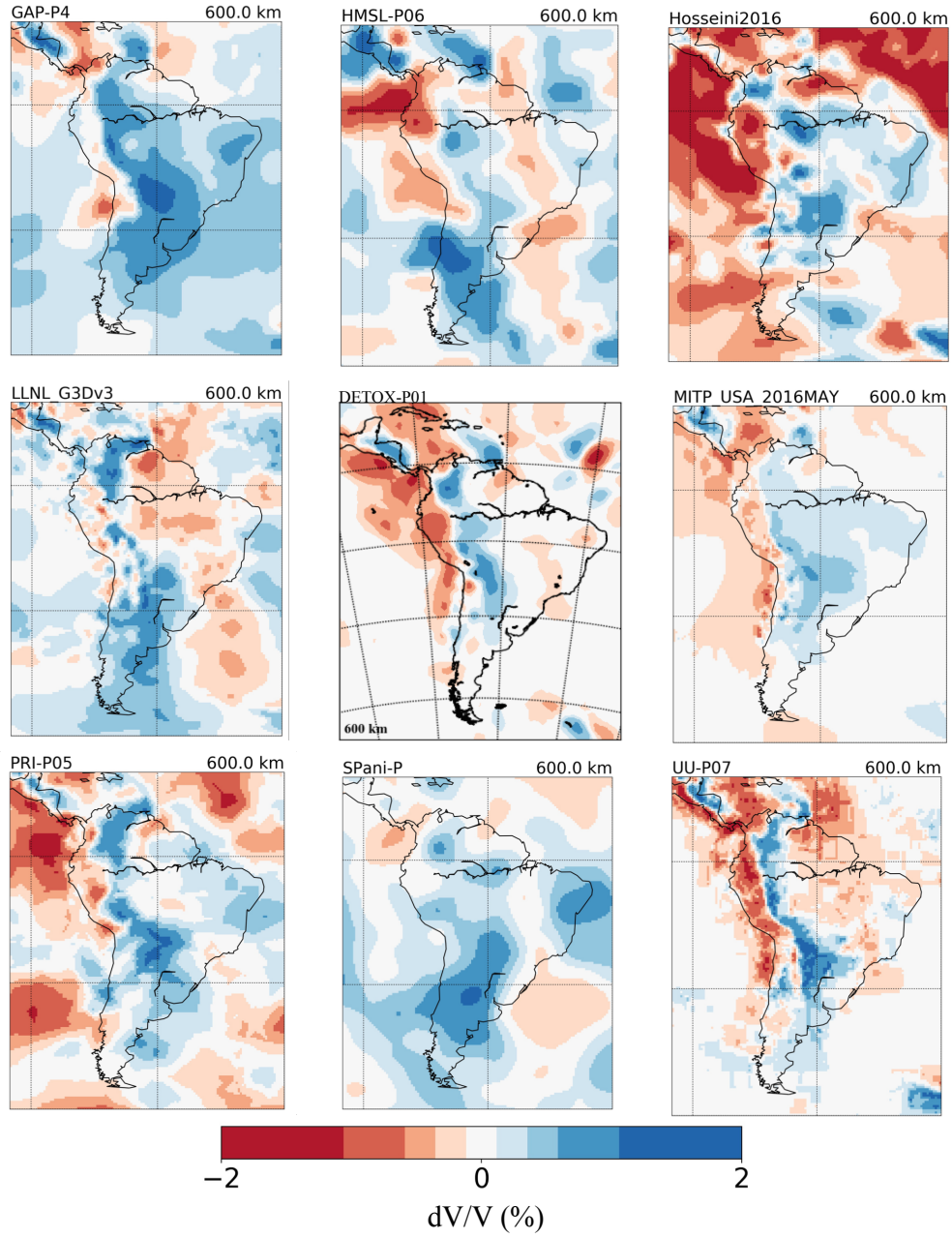


Figure 3.18: Comparison of the fourth slab segment in the different tomographies at 600 km. DETOX-0P1 (current model), GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

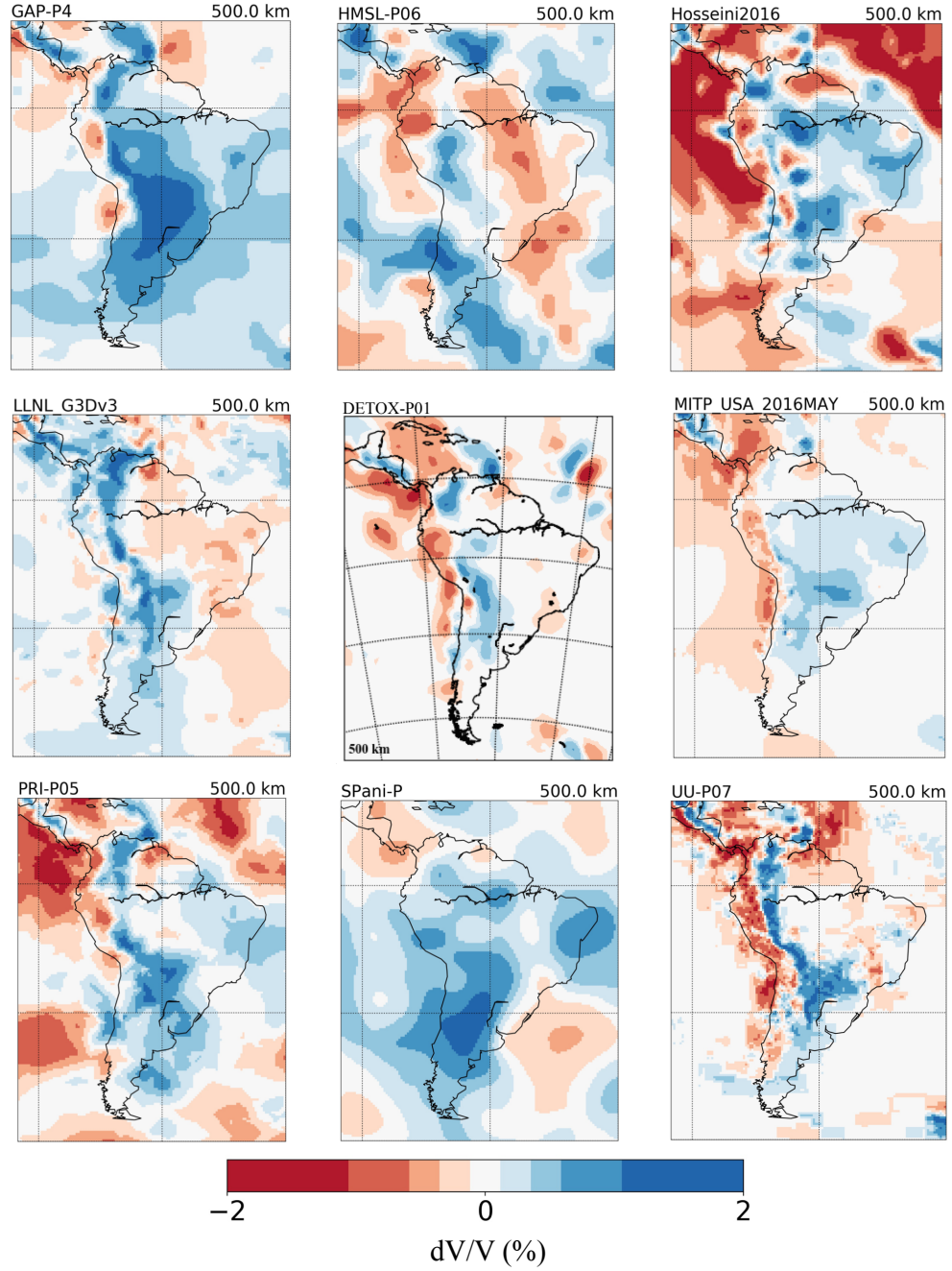


Figure 3.19: Comparison of the fourth slab segment in the different tomographies at 500 km. DETOX-0P1 (current model), GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

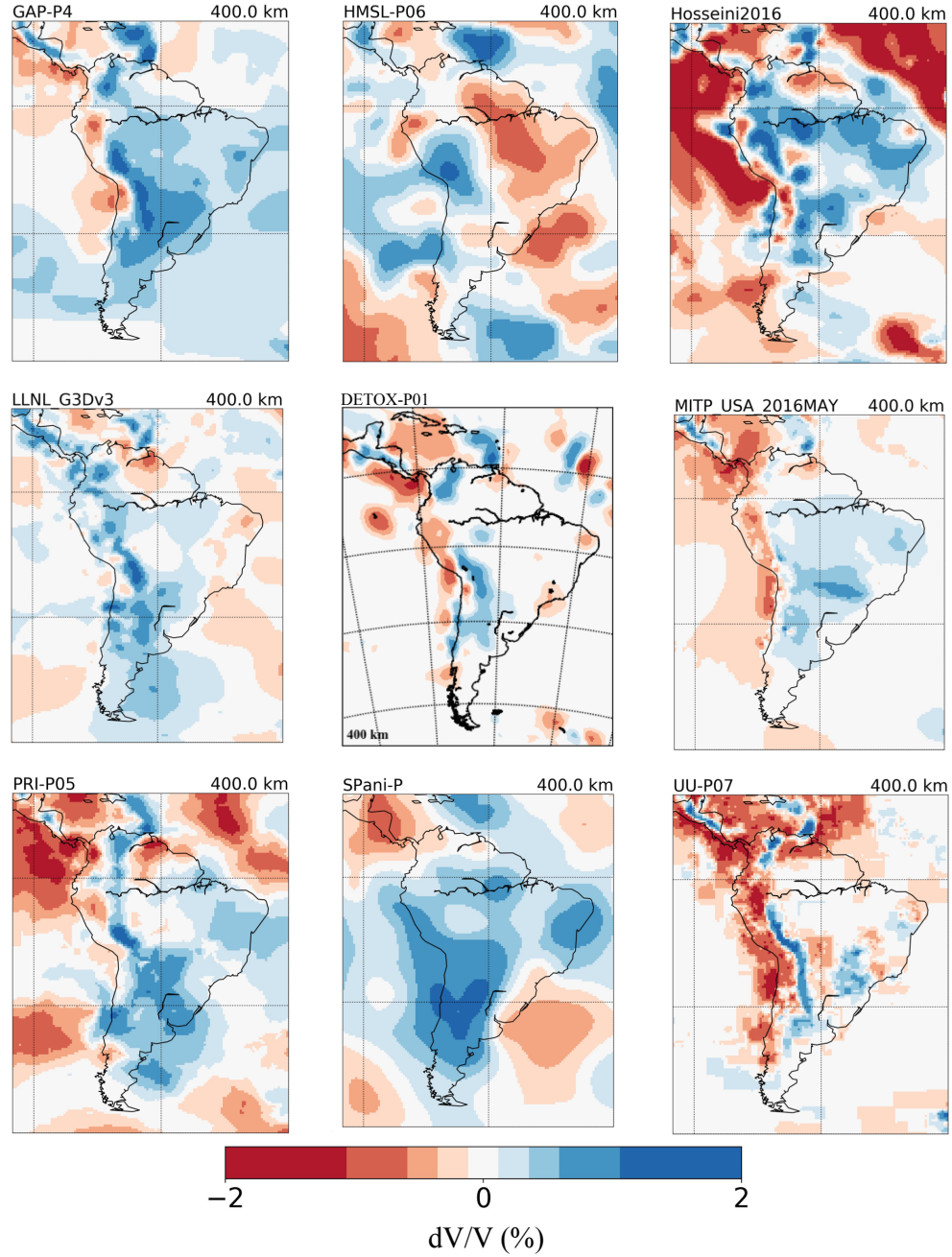


Figure 3.20: Comparison of the fourth slab segment in the different tomographies at 400 km. DETOX-0P1 (current model), GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

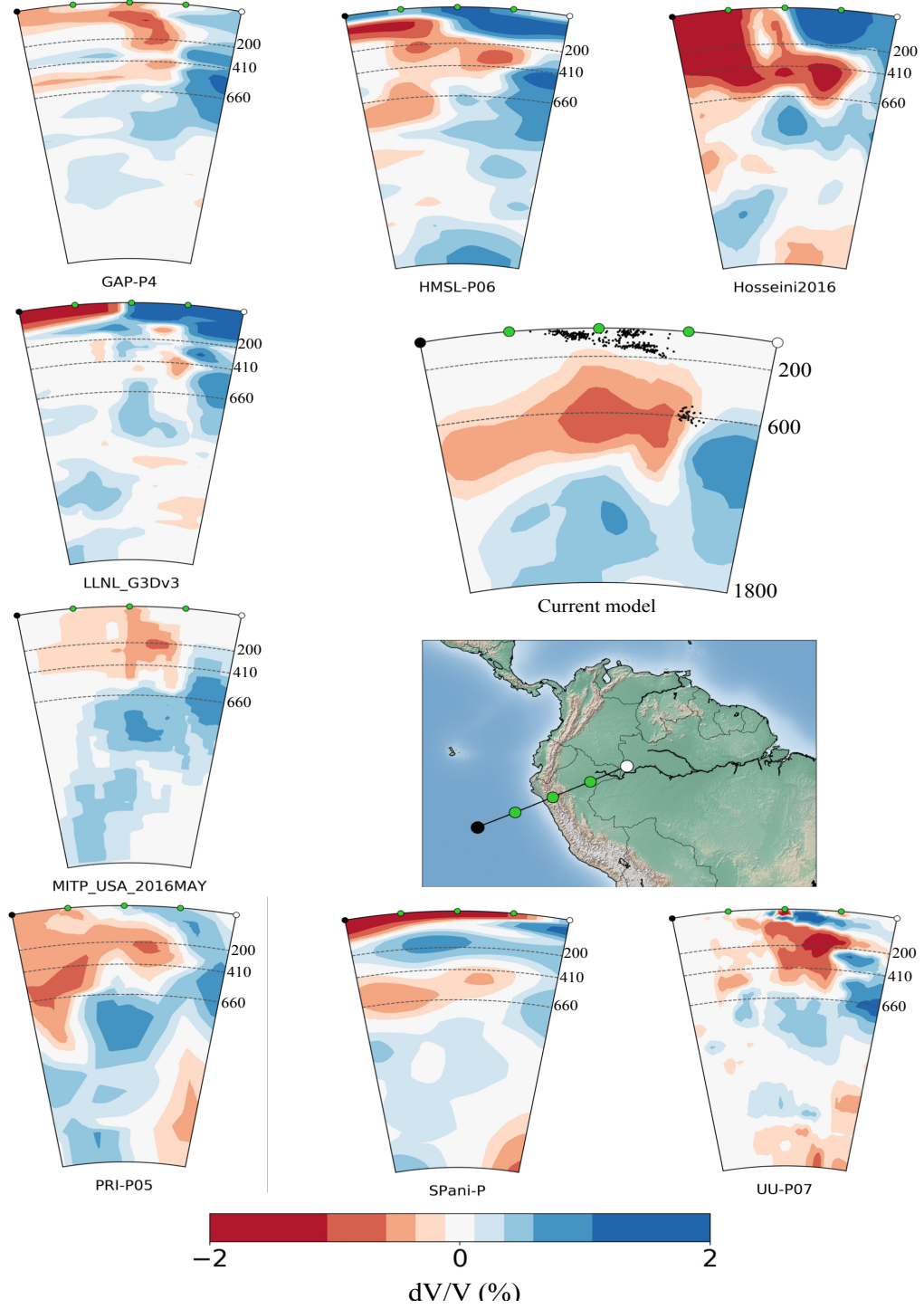


Figure 3.21: Vertical sections of Peruvian flat slab at northern latitude, recovered by different tomographies. The deepest depth in previous models show the core-mantle boundary. Black circles on the current model represent location of seismic sources. GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

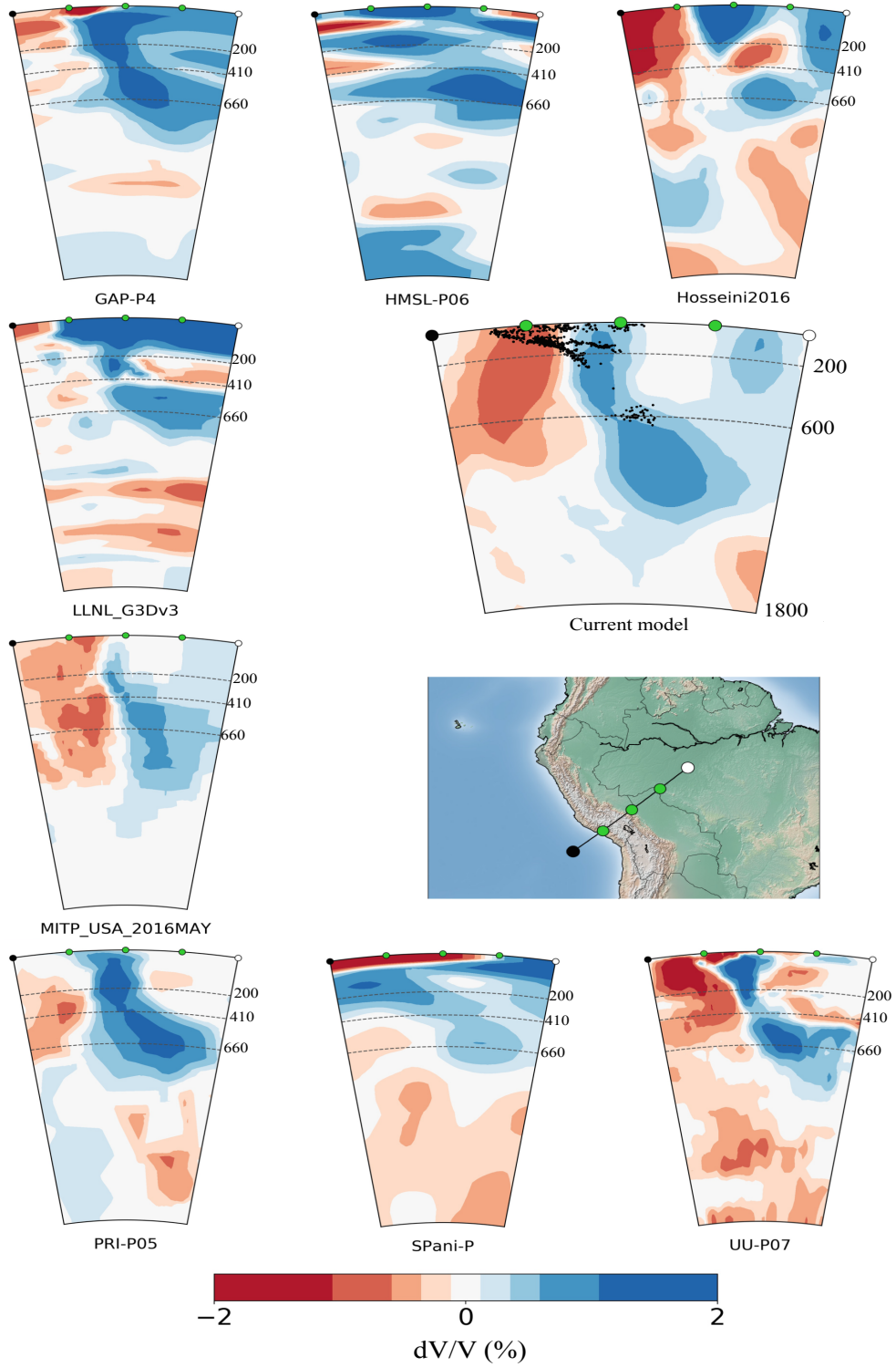


Figure 3.22: Vertical sections from surface to the core-mantle boundary, showing Peruvian flat slab at southern latitudes. Black circles on the current model represent location of seismic sources. GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

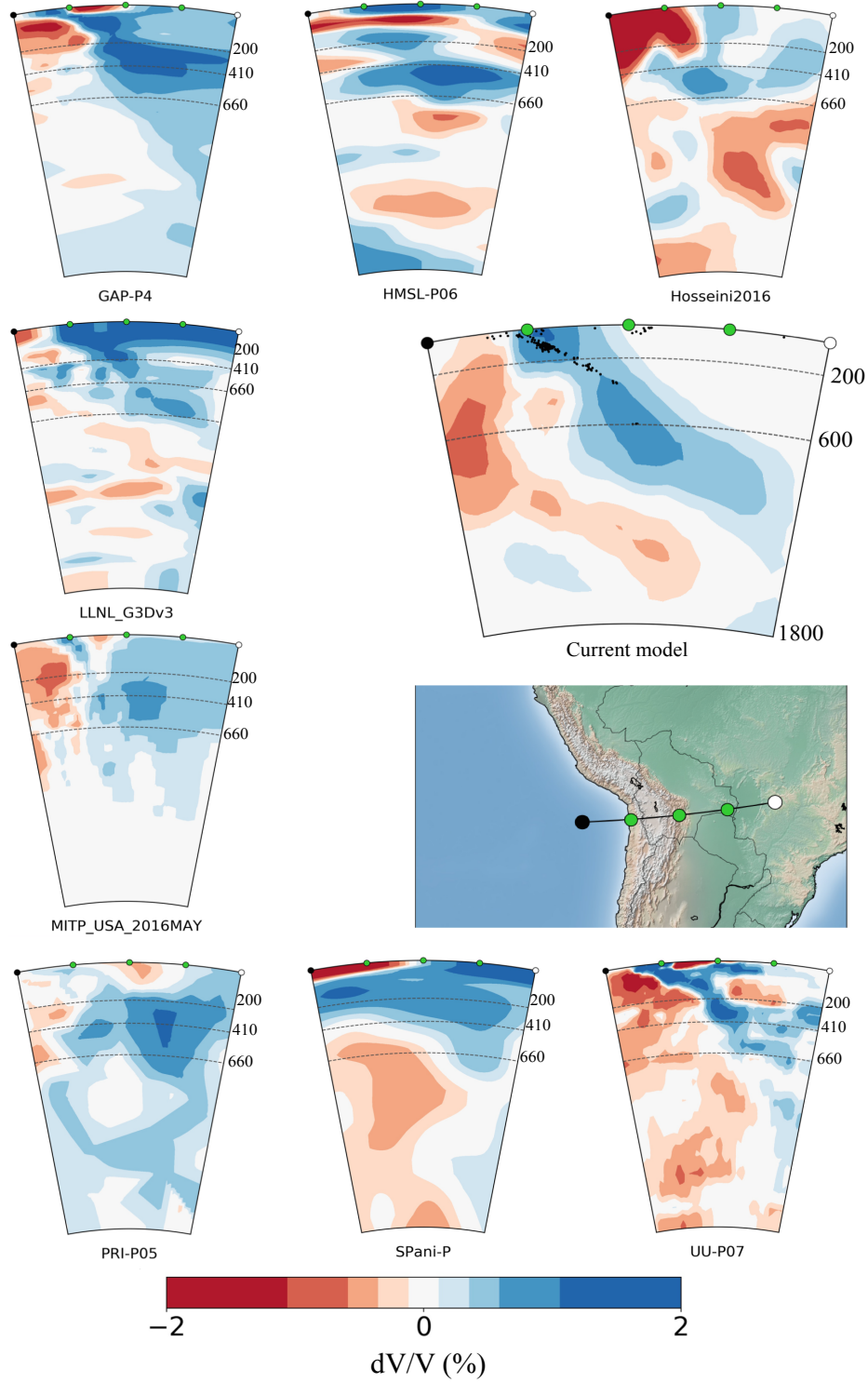


Figure 3.23: Cross-sections of the steep subduction under Central Andes, recovered by the P-wave tomographies. The deepest depth in previous models show the core-mantle boundary. Black circles on the current model represent location of seismic sources. GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

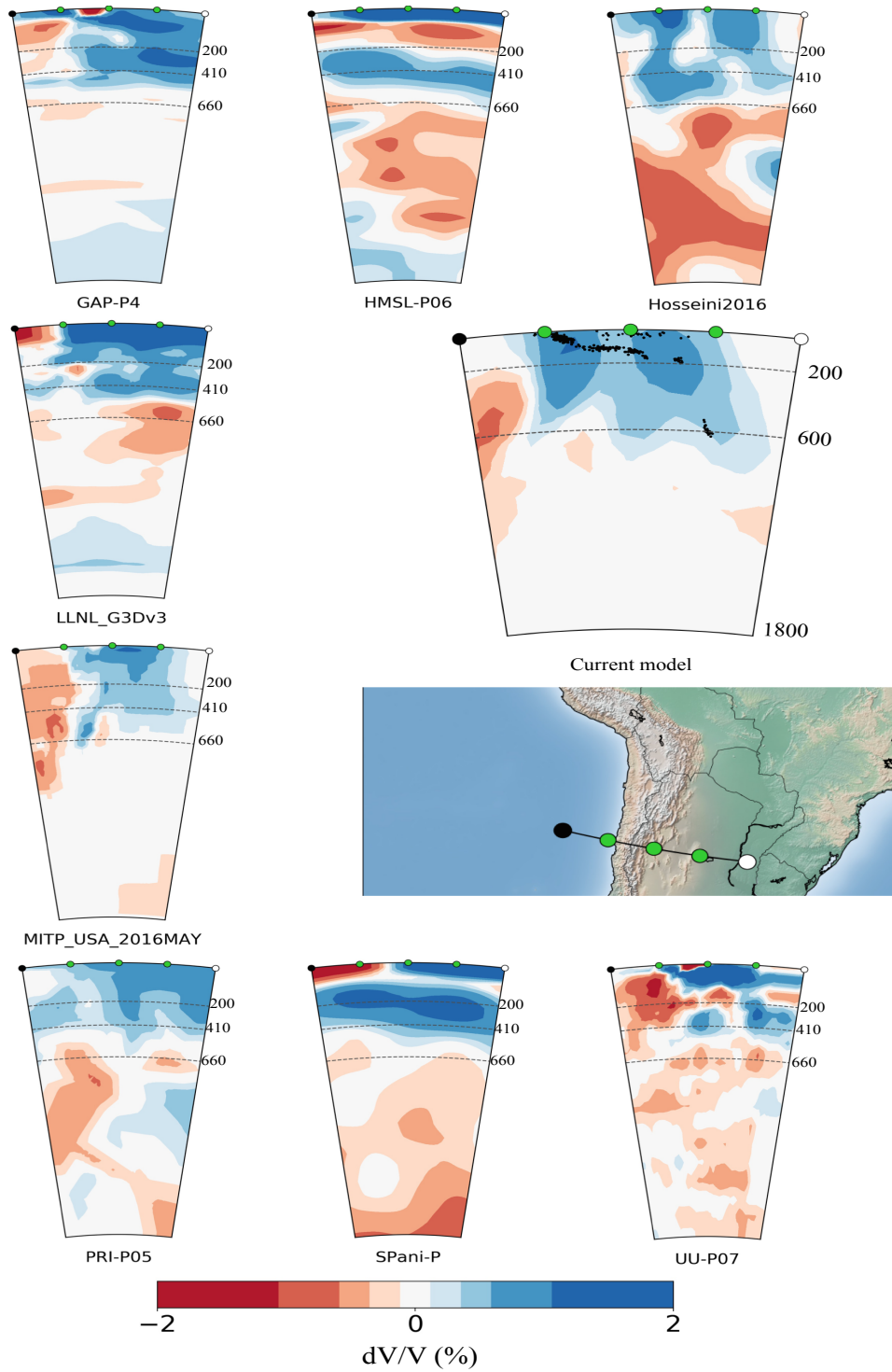


Figure 3.24: Cross-sections of Chilean flat subduction, recovered by P-wave tomographies. The deepest depth in previous models show the core-mantle boundary. Black circles on the current model represent location of seismic sources. GAP-P4 (Obayashi et al., 2013), HMSL-P06 (Houser et al., 2008), Hosseini2016 (Hosseini, 2016), LLNL_G3Dv3 (Simmons et al., 2012), MITP_USA_2016MAY (Burdick et al., 2016), PRI-P05 (Montelli et al., 2006), SPani-P (Tesoniero et al., 2015), UU-P07 (Amaru, 2007).

4

Mantle structures under South America

In this chapter, I integrate mantle structures under South America with plate tectonic reconstructions and make an interpretation for the development of the Northern and Central Andes from Cretaceous to Oligocene. Due to the lack of resolution in the P wave tomography of southern segment of the Andes, I will not consider it here. I discuss mantle provinces from deep (~ 1800 km) to shallow (300 km) depths, which are interpreted in terms of old to young subduction. Features shallower than 300 km are not discussed here. In the following section, I first present a brief overview of the regional geology in the north Andes, as this segment is the main focus of this chapter. This is followed by section 4.2, in which the details behind tomo-tectonic features, i.e. the combination of the mantle structures and tectonic models, are explained. Sections 4.3 to 4.7 contain interpretation of the reconstructions. This is followed by section 4.8, which summarises the evolution of mountain chains since early Cretaceous.

4.1 Regional geology of the North Andes of Colombia and Ecuador

The northern segment of the Andes is a mosaic of allochthonous, parautochthonous, and autochthonous crustal segments, which are divided into four separate ranges

(Pepper et al., 2016), including the Eastern, Central, Western and Coastal Cordilleras. The focus of this chapter is the Central, Western, and Coastal Cordilleras, which are characterized by the presence of oceanic terranes.

The Central Cordillera is mainly formed by Jurassic–Cretaceous igneous rocks intruded into and resting on both oceanic and continental crystalline basements (Pepper et al., 2016). Romeral and Peltetec terranes form ~ 1000 km of the western flank of the cordillera. This tectonic *mélange* comprises a mixture of high-pressure metamorphic rocks, island arcs, mafic and ultra-mafic complexes (Cediel and Shaw, 2003). Geological observations by Aspden and Litherland (1992) proposed the Ecuadorian segment of the suture zone, i.e. Peltetec terrane, was collided obliquely during the Early-to-Middle Cretaceous.

The Western Cordillera contains oceanic magmatic and marine sedimentary rocks of mainly Cretaceous age (Pindell and Kennan, 2009). The basement of the Western Cordillera is formed by the Late Cretaceous oceanic rocks in Colombia (e.g. Calima Terrane) and Ecuador (Pallatanga-Piñon Terrane) (Spikings et al., 2015). Geochemical, isotopic and geochronological data suggested that the mafic and ultramafic rocks in the cordillera are equivalent to those of the Caribbean Oceanic Plateau (COP) (Kerr et al., 1997). Oceanic plateaus are large portions of the ocean floor that are significantly thicker than the normal oceanic crust. These regions, which are classed as Large Igneous Provinces (LIPs), are generally believed to be formed by mantle plumes. The Caribbean LIP formed during the Cretaceous time when the other major phase of LIPs in the Pacific, including the Ontong Java, and Manihiki Rise were produced (Kerr and Tarney, 2005).

4.2 Tomo-tectonic reconstructions

4.2.1 Vertical sinking

The main challenge for integration of plate reconstruction modelling and subduction history is the dynamic of sinking slabs. Wall-like geometry of the slabs recovered by tomographic studies supports the idea that the slabs sank almost vertically. This can be interpreted to imply that slab walls have deposited beneath stationary

trenches from the moment of sinking. Thus, paleo trenches can be placed above the present-day location of the slabs, although they can be horizontally moved to the extent of uncertainty in the rheology of the mantle, and uncertainty in the plate reconstruction. Numerical modelling by Steinberger et al. (2012) argued that subducted slab might experience a few hundred kilometres lateral displacement from the surface to the core-mantle boundary in a standard rheology. Sigloch and Mihalynuk (2013) and Sigloch and Mihalynuk (2017), also estimated that the cumulative uncertainty may relocate the presumed paleo-trench of Mezcalera paleo-ocean a few hundred kilometres from their estimated reconstruction of tectonic history of North America. Vertical sinking is also inferred from the massive volume of the slabs in the current tomography (e.g Fa slab, which is still connected to the present-day western trench). However, for now this simple geophysical observation is considered as a null hypothesis and is tested by geological observations along the Andes in the following sections.

4.2.2 Summary of the reconstructions and figures

Figures 4.1 to 4.5 represent four-dimensional superimpositions of slab depositions, plate motion and seafloor spreading since early Cretaceous. To make these reconstructions, P-wave velocity anomalies with amplitude $> 0.25\%$ were contoured out and coloured by depth. The colour changes every 200 km in order to relate slab depth with timing of initial entry into the mantle via subduction. Slabs are assumed to sink vertically in a constant average rate of $\sim 10 \text{ mm yr}^{-1}$ in the mantle. This sinking is supported by previous studies of Sigloch and Mihalynuk et al, (2013) and (2017), which tested geological observations in North America against mantle structures modelled by tomography. In the first 200 km (from the surface to 200 km depth) the slabs' rate is presumed to be $\sim 20 \text{ mm yr}^{-1}$, as they are supposed to be mechanically attached to the plates. This rate is similar to the typical plate motion's speed Steinberger et al. (2012). Thus, the slabs above 200 km are assigned to 5 Ma, while below 200 km, every 100 km is assigned to 10 Ma. Section 4.8 discusses the choice of the sinking rate in more detail. This null hypothesis is

testable by geological records along the Andes, such as subduction of Farallon under the Central Andes during Cenozoic. In sections 4.3- 4.7 and chapter 5, I focus on the geological records of the Andes, which are independent data for examining the proposed sinking rate regime.

Every figure in figs. 4.1 to 4.5 contains two reconstructions of the same time. The larger-size reconstruction indicates the coloured slab at the corresponding time, overlain on the deeper (older) slabs. To have a better inference of the distortion of the slab geometry at each time slot, a smaller-size reconstruction of the same time is added, indicating the current slab alone, without laying on older slabs.

The vertical dimension of the slab is referred to as the ‘height’ of the slab, while its ‘length’ and ‘width’ refer to the trench-parallel and trench-perpendicular dimensions, respectively. Solid lines with barbs on represent the paleo-trench positions, which were inferred from the shape of the slabs at a 50-km depth increment. The polarity of subduction is inferred from the geometry of the slabs. Barb colours correspond to particular subducting oceanic plates: Farallon (light green), Mezcalera (light orange), Atlantic (yellow), Caribbean (purple). A trench line is assumed to be active, while the corresponding slab is overlain by slab materials in the next time slot. If no new slab is added to the current slab, the trench is assumed to be extinct. Present-day location of coastlines are drawn with a light silver solid line, whereas reconstructed coastlines and magnetic reversal isochrones at each respective time interval are drawn with a dark grey solid line. Slab colour changes every 200 km, which is corresponding to 20 Myr of subduction, based on a presumed slab sinking rate of 10 mmyr^{-1} in the mantle. The depth-dependent slabs are overlain on the ocean floor isochron dataset by Muller et al. (2016) for spreading history, and plate rotation model by Matthews et al. (2016) using the GPlates software. In the following sections, four slab units, which were described in Chapter 3, are interpreted within the above kinematic framework.

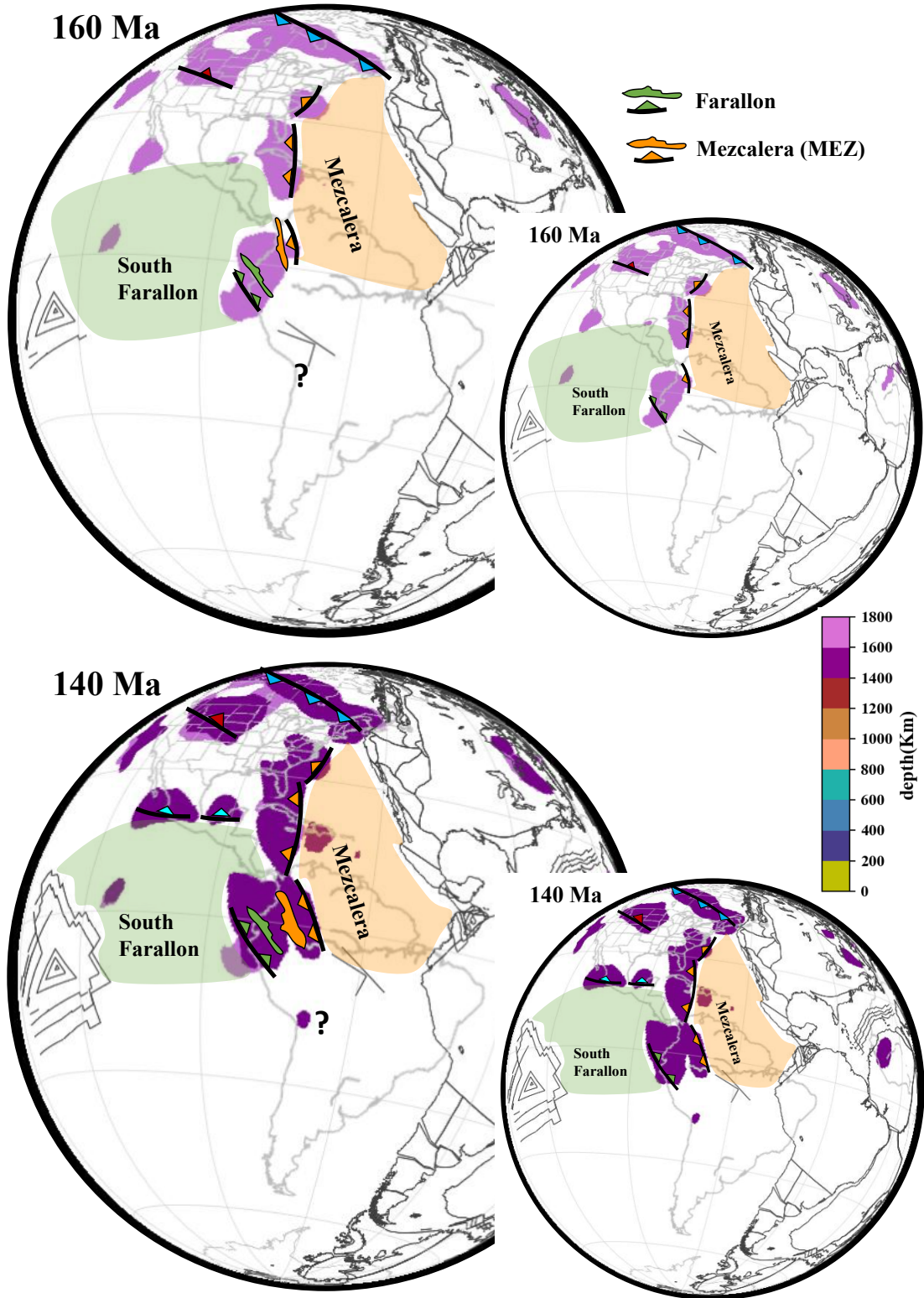


Figure 4.1: Tomo-tectonic synthesis of Andes from Late-Jurassic to Early-Cretaceous. MEZ: Mezcalera slab, Fa: Farallon slab. The right-hand side column explains the cartoons used in the reconstruction.

4.3 Fast anomalies from 1800 to 1200 km depth: Double-sided subduction zone (~ 180 - ~ 120 Ma)

Figure 4.1 superimposes slabs at 1800 km to 1700 km depth and 1700 km to 1500 km depth on the plate reconstructions at ~ 160 Ma and ~ 140 Ma, respectively. During this period

1. The Farallon plate moves eastward relative to the mantle. It is constrained by the spreading isochrones.
2. North America drifts westward, constrained by the Atlantic paleo-magnetic isochrones.
3. The seismic anomalies are all vertical implying a stationary trench.

To establish the positions of paleo-trenches, I assume a lower mantle reference frame, which locates the massive slab to the west of the South American margin. This choice seems sensible, since it 1) supports the subduction of Farallon seafloor isochrones in the west of the subduction zone and 2) is in good agreement with the geological records, which does not offer any evidence of Early-Cretaceous subduction under the north-western margin of South America (Jaillard et al., 2009). Pindell and Kennan (2009) argued against the development of autochthonous Early Cretaceous continental arcs in Ecuador and Colombia. If the slabs subducted later, it would be deposited under north-western margin of South America, which contrasts the latter. Accepting the paleo-position of the intra-oceanic subduction thousands of kilometres to the west of South America, the question remains how to draw the trenches relative to the west coast of the continents. Here the paleo-positioning is obtained under the vertical sinking assumption, which is inferred from the vertical slab walls. Under this approximation, two slabs are inferred including the most westerly Fa, constrained by eastern Farallon isochrones, being subducted since +180 Ma and the most easterly MEZ. In fact, there must be two subduction zones to ensure the trenches are stationary. I propose the MEZ slab is the southern extent of subducted

Mezcalera ocean, which was suggested to be overridden by North America and closed by Cretaceous (Sigloch and Mihalynuk, 2013; Sigloch and Mihalynuk, 2017). The strong link between westward dipping MEZ slab and west-dipping Mezcalera ocean, has motivated me to identify it as the southern verge of the ancient ocean. This double-sided subduction might indicate an environment in which Romeral and Peltetec terranes were formed. These terranes are located in the western flank of the Central Cordillera. The timing of the subduction is close to the Late Jurassic to Early Cretaceous age of the oceanic sediments in the Romeral terrane (Cediel and Shaw, 2003).

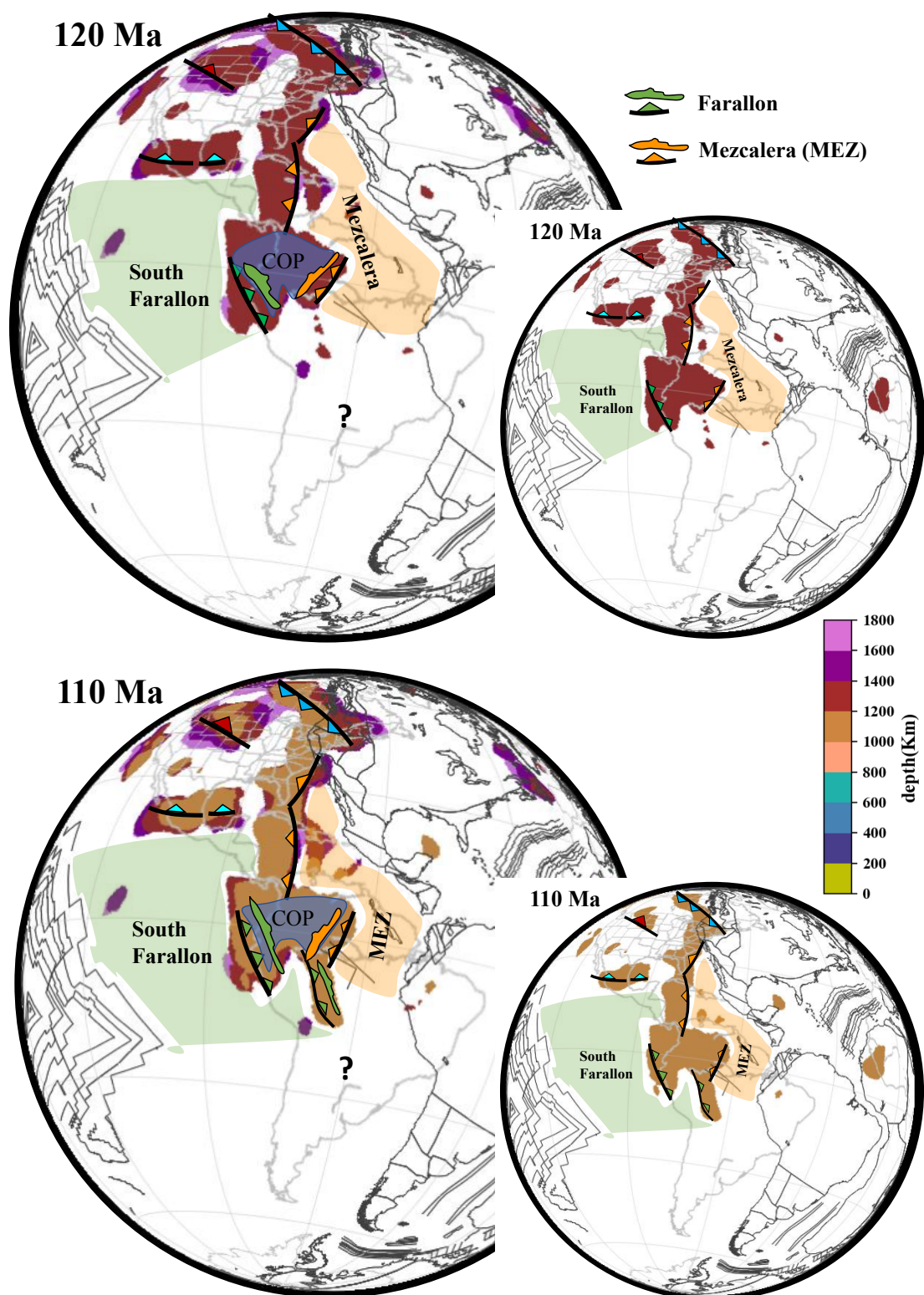


Figure 4.2: Lower Cretaceous constructions of Andes. MEZ: Mezcalera slab, Fa: Farallon slab, COP: Caribbean Ocean Plateau.

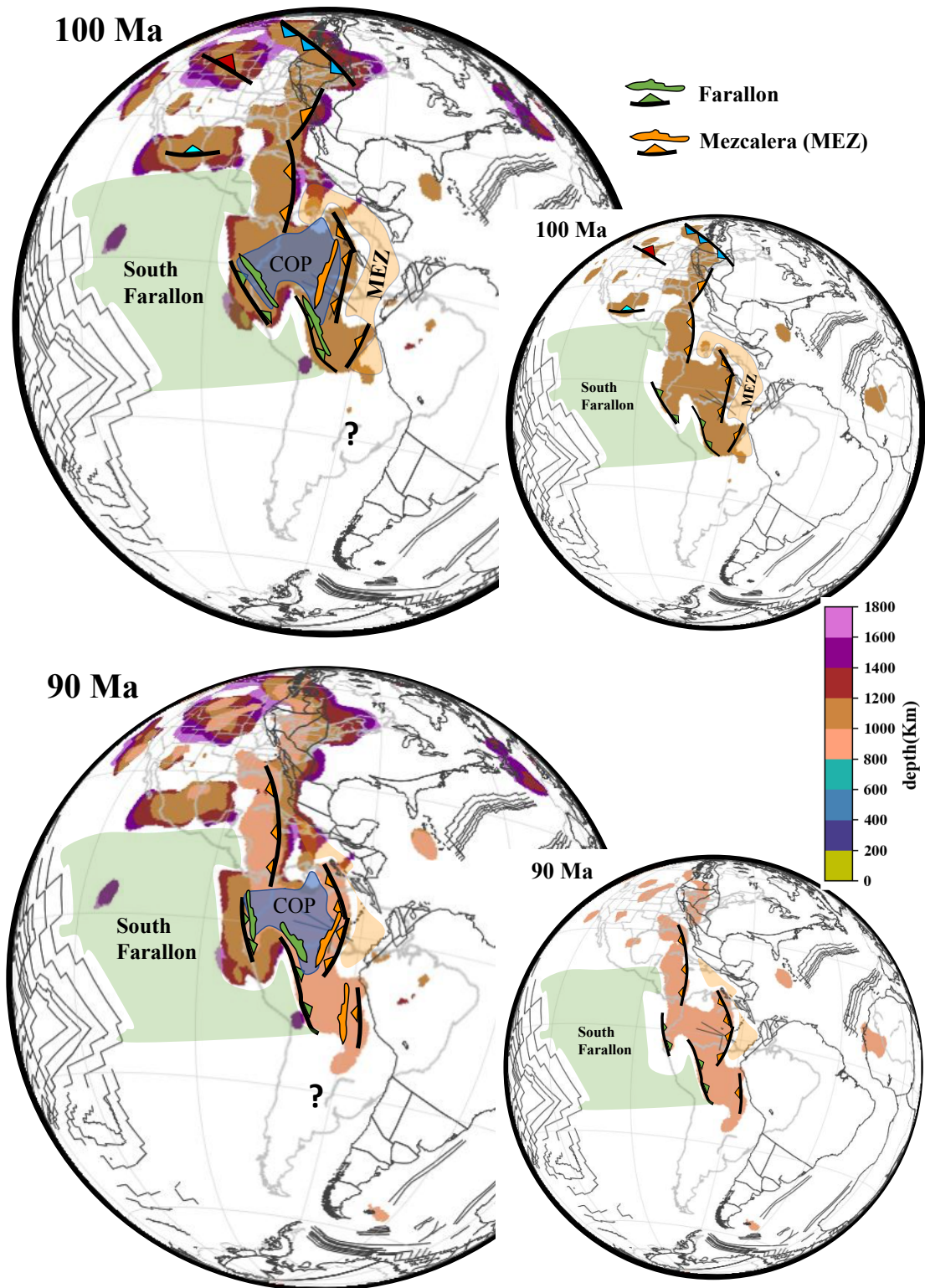


Figure 4.3: Mid-Cretaceous synthesis of Andes. MEZ: Mezcalera slab, Fa: Farallon slab, COP: Caribbean Ocean Plateau.

4.4 Fast anomalies from 1200 km to 800 km depths

Tomo-tectonic reconstructions of 120-70 Ma are illustrated in figs. 4.2 and 4.3. The figures show the subduction initiation in the south or southwards growth of the lateral extent of the trenches. Of the lower mantle slabs in these figures, the Fa slab wall which is already migrated to the eastern trench and connected to the present-day active subduction of Nazca can be reliably linked to the Farallon lithosphere, with which we can identify other slabs. The Farallon plate reconstruction implies that the easterly trench of MEZ should have migrated eastward as well, since the slab mass closer to the Pacific-Farallon spreading ridge, i.e. Fa slab, should be considered as the Farallon. Therefore, I interpret the most easterly oblique slab, which appears to the eastern edge of COP in fig. 4.2 at 120 Ma time frame, as the rollbacked MEZ. The opposite scenario, i.e. if the oblique slab were created from the rollback of Farallon slab, while Mezcalera remained in the old westerly trench, is not supported by the spreading isochrones.

This sudden change in the slab geometry at ~ 120 Ma is coincident with the age proposed for the inception of Caribbean LIP by Pindell and Kennan (2009). I propose that the eastward migration of MEZ-Fa may mark the location of the plateau's opening, provided that I assume the buoyancy of the plateau caused the intra-oceanic subduction of MEZ and Fa after 120 Ma. Earlier works by Nair and Chacko (2008) and Leat and Larter (2003) hypothesised that the gravitational instability due to the compositional and density contrast between the normal lithosphere and a plateau may produce intra-oceanic subduction systems. Thus, here, I suggest that the thickened plateau triggered the subduction since 120 Ma. This tomo-tectonic driven location of the plateau, which is close to the present-day location of Ecuador, is consistent with paleomagnetic results, which revealed that the Caribbean plate was derived from near-equatorial latitudes (Acton et al., 2000). The outlines of Fa and NE-trending MEZ slabs between 120-75 Ma conform to the convex margins of South America and may highlight the location of the collision of the arcs to the continent. If this is the case, between ~ 120 and ~ 75 Ma two

accretionary events along the north-western margin of the Andes occurred, which are explained in the following two sub-sections.

4.4.1 Collision of Romeral-Peltetec (~ 120 Ma)

Figure 4.2 shows the sudden change in the geometry of subduction system initiating at ~ 120 Ma, which is interpreted as the opening of the Caribbean plateau. This change may also mark the end of the double-sided subduction setting explained in section 4.3. I propose the island arcs that were produced from the double-sided subduction zone since ~ 160 Ma, began to collide with the north-western margin of South America at ~ 120 Ma. This collision may have caused Romeral and Peltetec terranes to amalgamate along the Central Cordillera. Aspden and Litherland (1992) also suggested an Early-to-middle Cretaceous age for the collision of the Peltetec block.

4.4.2 Evolution of the Western Cordillera of Colombia and Ecuador

The volcanic rocks, which have been developed along the Caribbean plateau edges since ~ 120 Ma, began to collide with the north-western margin of South America and formed the Western Cordillera of the Andes during the Late Campanian, when Mezcalera ocean is closed (Figures 4.3 and 4.4).

Consequently, the collision led to a prolonged assemblage of NNE trending island arcs to the Colombian and Ecuadorian Andes, which preserved un-subductable thickened plateau material in tectonic slices. Buoyancy analyses by Cloos (1993) indicated that island arcs need to be active at least 20 myr so that they cause an orogenic collision. Thus, the intra-oceanic island arcs, which had been erupting for more than 20 m.y over the Caribbean plateau, were buoyant enough to resist the subduction under the margin of South America at Campanian interval.

I suggest that this phase of accretion lasted until ~ 50 Ma, when SB and Ca slabs appear and a third phase of accretions (see section 4.6) begins. Geochemical analyses indicate that the Western Cordillera comprises volcanic rocks, with island arc

affinities, which formed during late Cretaceous (Reynaud et al., 1999). In addition, geochemical, isotopic and geochronological data suggest that the Cordillera contains ultramafic and mafic rocks, the composition of which resemble that of the Caribbean plate (Kerr et al., 1997). Due to this similarity, researchers have been arguing that the plateau collided with the north-western margin of South America. The proposed collision time ranges from $\sim 85\text{--}80$ Ma (Kerr et al., 2002) or ~ 75 Ma (Vallejo et al., 2006; Spikings et al., 2015). Stratigraphic (Jaillard et al., 2009) and geochemical (Kerr et al., 2002) studies proposed that the accretion of the oceanic terranes lasted until the late Paleocene. Similar to my proposal, U-Pb zircon dating by Vallejo et al. (2006) suggested that the oceanic plateau was intruded by a west-dipping intra-oceanic arc prior to their collision with South America in the Campanian. My interpretation correlates well with the geological timing including 1) Late Cretaceous formation of the island arcs overlying the plateau, 2) collision of the oceanic complexes with South America during Campanian , and 3) termination of the accretion in the Late Paleocene.

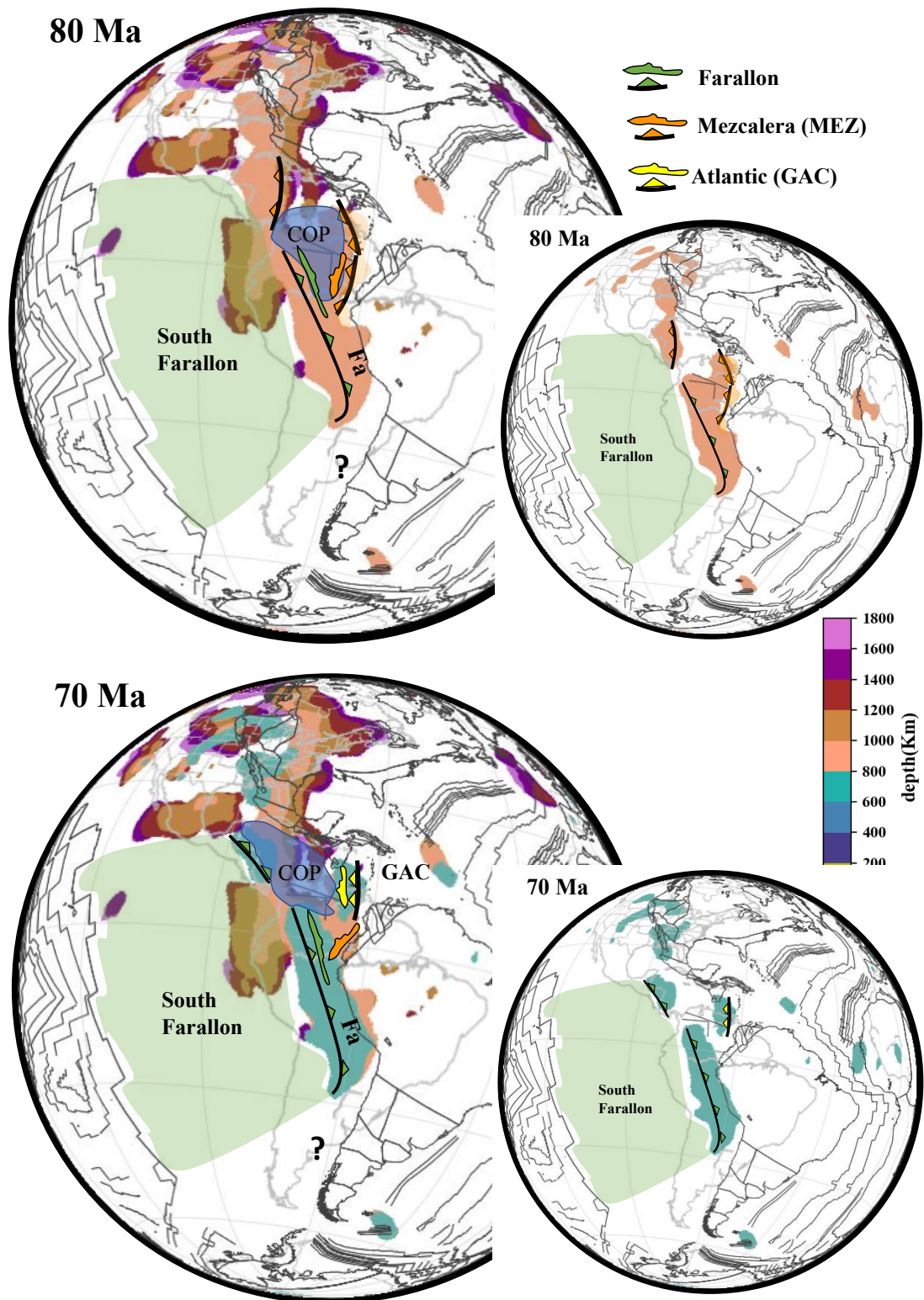


Figure 4.4: Upper-Cretaceous reconstructions of Andes. MEZ: Mezcalera slab, Fa: Farallon slab, COP: Caribbean Ocean Plateau, GAC: Atlantic ocean.

4.5 Fast anomalies from 900 to 600 km depths: initiation of the Farallon subduction under the Central Andes (~ 70 Ma)

Figure 4.4 shows that the Fa slab developed along a longer NW-SE trench since ~ 70 Ma, which is parallel to the western margin of South America. Since then, a rollback in the subduction of Farallon is visible, which led the slab to dip eastward. The eastward pushback is evident in figs. 4.4 and 4.5, where older slabs are protruding along their width perpendicular to the trench. This sudden change in the subduction of Farallon, which coincides the western margin of the continent overriding the slab, might be interpreted as the beginning of a new period of magmatism along the Central Andes. Presence of the Fa slab under the Central Andes is consistent with the abundant magmatism in the mountain belt $< \sim 100$ Ma, which is shown by U-Pb geochronology and Hf isotope analysis of zircon (Pepper et al., 2016). However, the lack of slab prior to this period is questionable, as magmatism were reported to be active in the Central Andes since Jurassic (Pepper et al., 2016). Geochronological analyses show a strong periodicity in the ages of magmatism in the Central Andes during the past 200 m.y (Pepper et al., 2016). Haschke et al. (2002) found four episodes of magmatism for northern Chile, occurring at 195–129, 123–82, 76–37, and 22–0 Ma, which were separated by periods of quiescence. They proposed the periodic magmatism may be caused by a cyclic change in the slab dip, in that termination of subduction-related magmatism is connected to the onset of flat subduction, followed by slab-steepening due to the eclogitisation of the slab tip. Thus, one possibility for the lack of slabs prior to ~ 70 Ma may be the occurrence of a period of flat subduction, which may have pushed the slab eastward, where my tomography might have not identified it. The Cenozoic subduction of Farallon under Central Andes and the lack of the subduction during the Cretaceous are analysed in detail in chapter 5.

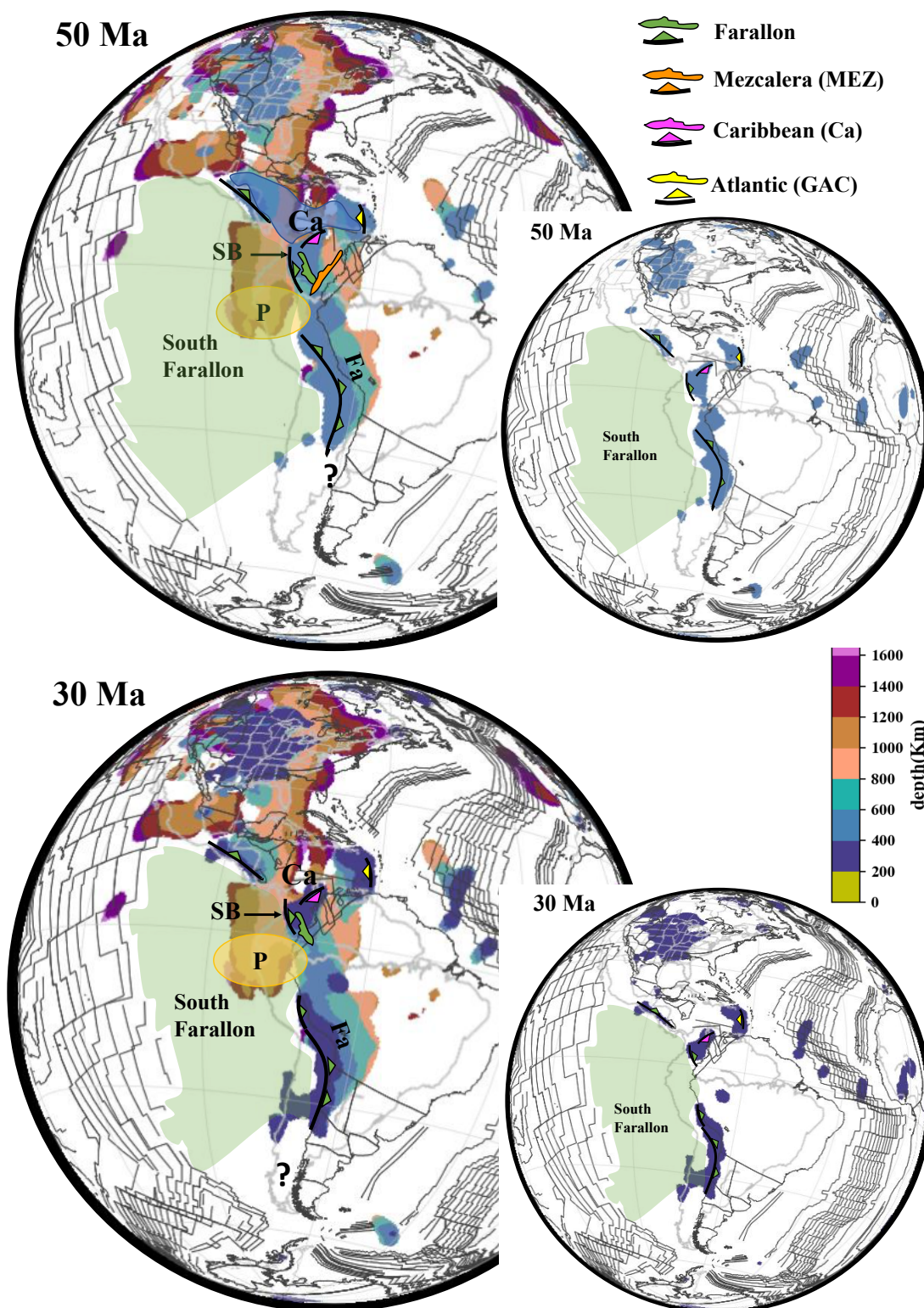


Figure 4.5: Tomo-tectonic synthesis of Andes during Paleogene. Ca: Caribbean slab, Fa: Farallon slab, SB: Serranía de Baudó terrane, P: Piñon Plateau. Orange circles display location of the collision of the hypothetical plateaus.

4.6 Fast anomalies from 600 to 300 km depths

The fourth division of the slabs, as it was described in Chapter 3, comprises the break-up of the Fa slab by P gap and advent of SB and Ca anomalies under North Andes. Figure 4.5 illustrates 50 and 30 Ma reconstructions of the east-dipping subduction of Farallon under Central Andes. Taken together, they indicate that the initial NW-SE strike of Farallon at ~ 70 Ma (fig. 4.4) is turning toward NS gradually. This is consistent with the Eocene change in the Farallon plate motion, recorded in magnetic stripes, suggesting that southern flank of the northeast-moving Farallon plate has veered toward the more easterly motion of the future Nazca plate well before 23 Ma (Lonsdale, 2005).

The narrow Ca anomaly underlies Northern Venezuela, orienting toward SW. Here I associate the slab, which appears between ~ 55 and 25 Ma, with the Eocene subduction of the Caribbean plate under northern Venezuela. This is consistent with, recent geological evidence of subduction-related magmatism found in north-western Andes (Cardona et al., 2011; Bayona et al., 2012), which were related to the subduction of Caribbean edge. Earlier work by Taboada et al. (2000) also identified a N-NE trending subducted slab under the Eastern Cordillera of Colombia, plunging toward the E-SE, which was thought to be a remnant of subducted Caribbean Plateau.

The SB anomaly in fig. 4.5, on the other hand, is a fragment of Farallon, constrained by the seafloor isochrones. This subduction is discussed in the following section. The southern gap, which is in fact an indentation in Fa slab, precedes the Chilean flat (Cross-sections 4-6 in fig. 4.6) subduction. This feature, which might indicate a heterogeneity on the Farallon plate, is discussed in chapter 5. In the following section, I discuss the presence of the P gap, which is depicted by an orange circle in fig. 4.5.

4.6.1 Peruvian Gap and Piñon Plateau

Under the current time-depth correlation, the P gap appears under Ecuador during Eocene time. The subduction of ocean floor beneath a continent is considered as a steady-state process. It ends when either the subducting plate is prevented from

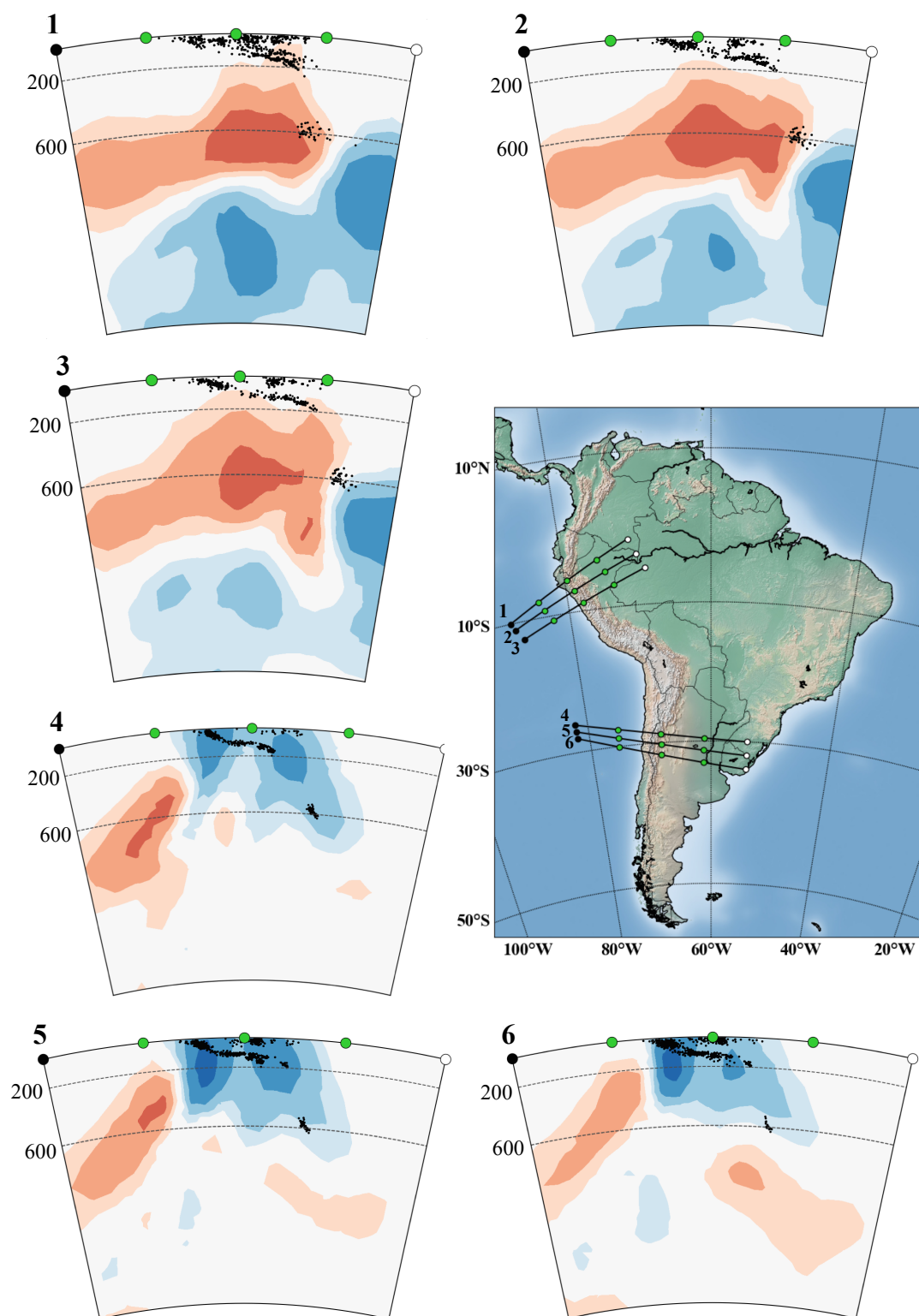


Figure 4.6: Tomographic cross-sections under Peru (1-3) and Chile (4-6), superimposed by seismic sources (black circles).

sinking (e.g., collision of a continent or large oceanic plateau) or the slab-pull force is no longer transmitted to the surface plate (e.g., ridge subduction or slab break off). The observed gap cannot be a slab detachment, the process used to explain the syn-collisional magmatism (Davies and von Blanckenburg, 1995), due to the lack of direct evidence of any subduction along the Ecuadorian margin between late-Jurassic and late Eocene (Jaillard et al., 2009). The question remains why the slab gap in the foundering Farallon took place under Ecuador.

I propose that the collision of the Ecuadorian coastal Plateau, which was recognised by geochemical studies, created this pause in the current reconstruction of the subduction of Farallon during Eocene time. This implies that no direct subduction happens under Ecuador at this period, but rather the hypothetical plateau was accreted. The collision might have pushed the subduction back to behind the trailing edge of the plateau, as was suggested by Kerr et al. (2002). Although, the back-stepped subduction is not recovered in the present tomography, it might be the reason why no slab exists under the western Ecuador in my model.

The region of the slab gap is consistent with the aseismic regions in the seismicity sections of fig. 4.6 (cross-sections 1-3). No earthquake with magnitude > 3 has been recorded between years 1967 to 2014 (from ISC-EHB catalogue). The region of no seismicity coincides with the lower (600 km) and upper (250 km) limits of the slab gap in my model, which is interpreted as the corresponding time (~ 55 -25 Ma) of the plateau collision.

The accretion of the plateau should be completed by early Oligocene (~ 23 Ma), when subduction of Nazca plate commenced (Hey, 1977). The subduction of Nazca can be observed in recent surface tomographic studies (Schaeffer and Lebedev, 2013), where the gap is filled with high velocity anomalies above 250 km depths. However, in the current P-wave tomography, the gap is visible up to the surface due to the limited resolution at shallow depths < 300 km.

This interpretation is consistent with geochemical studies that suggested the coastal terranes of Ecuador were derived from a separate plateau, having collided to Ecuador during Eocene time (Kerr et al., 2002). The igneous sequences in both the

Western Cordillera and coastal terranes of Ecuador (Piñon unit) were suggested to be intruded at the same time. However, as they accreted at different periods, researchers suggested that they belong to two different plateaus. A chemical difference in the composition of the two units was also reported, as the coastal terrane of Piñon unit showed more depleted “light rare earth elements” (LREE) patterns (Kerr et al., 2002). In addition, since the igneous basements of western Ecuador are covered by Upper Cretaceous sediments, it cannot belong to late-Cretaceous Caribbean plateau (Reynaud et al., 1999; Kerr et al., 2002).

4.6.2 Peruvian Gap and SB slab

After the break-up of Farallon by the Piñon plateau, the northern segment of the Fa slab, named SB, was formed. This slab, which is located offshore, might refer to an intra-oceanic subduction. The lithosphere between the SB arc and the coastline might indicate the presence of the hypothetical Piñon plateau. If this is the case, the Farallon subduction occurred along the edge of the plateau and formed the SB arc. It collided with the Pacific coast of NW Colombia, and formed Serranía de Baudó later in the Early-Oligocene. The time of the collision of the terrane is inferred from when the SB slab underlays the north-western coastline.

This scenario raises some questions, including:

1. What evidence supports the idea of the SB slab being formed by Farallon?
2. Which evidence supports the theory that the subduction took place along the edge of the Piñon plateau?

First, the SB slab is constrained by the Farallon seafloor spreading, which signifies the idea that SB was formed by the subduction of Farallon.

The rather NS strikes of the SB slab, which appears near the gap encourages the idea that the subduction of Farallon took place along the western edge of the hypothetical plateau. The basement of Baudó terrane was suggested to be similar to the Piñon plateau of coastal terranes of Ecuador by geochemical studies (Kerr et al., 2002). Furthermore, the age of the subduction-related rocks overlying the

basement of the Baudó terrane is younger than that of the Western Cordillera (Kerr et al., 1997). Therefore, 1) the age of the subduction-related materials in the terrane and 2) the collision time of the plateau is rather close, which is almost contemporaneous to the subduction of the SB slab and advent of the P gap in my model. Thus, it seems reasonable to suggest the SB subduction occurred along the edge of the plateau in my model. Kerr and Tarney (2005) proposed almost similar time span $\sim 45\text{-}40$ Ma for the collision of the plateau.

4.7 Slab sinking rate

In sections sections 4.3 to 4.6, it was observed how the reconstructions made by a sinking rate of $\sim 10 \text{ mmyr}^{-1}$ leads to the appearance of the Farallon trench to the west of the North Andes at ~ 180 Ma. This hypothesis led us to recover the segmentation along the Andean belt. These two events, i.e. 1) Upper-Jurassic initiation of Farallon subduction, and 2) Cenozoic east-dipping subduction of Farallon under western margin of South America, which are consistent with geological records, are used as the proxies to evaluate other proposed sinking rates in this section.

Domeier et al. (2016) investigated the possibility of using seismic tomography in building a reference frame under vertical sinking assumption. Their results indicated a significant time-depth correlation and proposed a global average sinking rate of $11\text{-}19 \text{ mmyr}^{-1}$ in lower mantle. Here I examine these values of sinking rates. Figure 4.7 compares tomo-tectonic reconstructions, recovered from different sinking rates in this range. A constant rate of $\sim 20 \text{ mmyr}^{-1}$ in the lower mantle appears too fast, as we end up with a significant slab gap to the east of Pacific from $+180$ to ~ 85 Ma, while Farallon isochrones are spreading during this period. By reducing the rate, this period of subduction quiescence is decreased. A sinking rate of $\sim 11 \text{ mmyr}^{-1}$ leads to an inconsistency between the Farallon magnetisation and the subducted lithosphere between ~ 180 Ma and ~ 150 Ma. The sinking rate of 10.5 mmyr^{-1} might locate the east-dipping Farallon ~ 70 Ma under Central Andes. Yet, the advent of the Fa slab is delayed for ~ 20 Myr from onset point of Farallon

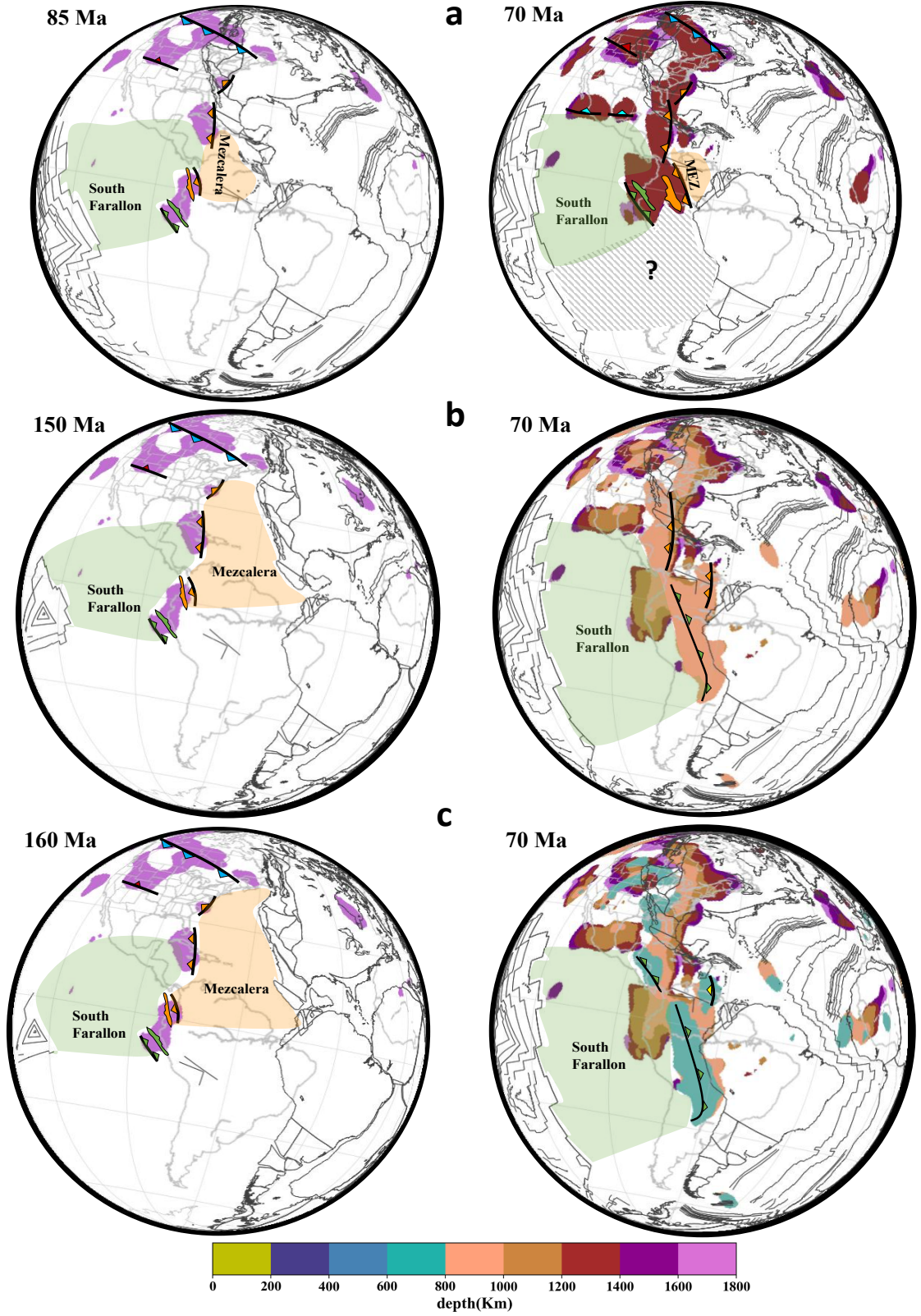


Figure 4.7: Comparison of tomo-tectonic reconstructions made with constant average sinking rates of a) 20, b) 11 and c) 10.5 mmyr^{-1} . Images in the left columns indicate the time slot when the Farallon slab first appears. The right column shows position of Farallon slab at ~ 70 Ma.

isochrones at ~ 180 Ma.

A constant sinking rate of $\sim 10 \text{ mmyr}^{-1}$ appears to support best both the Late-Jurassic Farallon isochrone as well as geological records along the Andes. It also seems to be consistent with the sinking rate of Farallon and its successors under North America, as Sigloch and Mihalynuk (2013) proposed $10 \pm 2 \text{ mmyr}^{-1}$ speed for lower mantle slabs. In all of the reconstructions, the sinking rate in the first 200 km is considered to be 20 mmyr^{-1} . This means that I assumed slabs become detached from the plate after 200 km. Increasing this depth to the upper mantle delays the appearance of Fa slab by ~ 150 Ma. In other words, the assumption of slabs being detached in the upper mantle leads to a slab gap between ~ 180 and 150 Ma.

4.8 Synopsis

Summing up this time-depth inventory reconstruction, I was able to explain the contrasted evolution of the Andean mountain chain. A simple assumption of vertical sinking, with an average constant rate of $\sim 10 \text{ mmyr}^{-1}$, leads to a sufficient reconciliation between geological records and kinematic constraints as follow:

1. Intra-oceanic subduction of Mezcalera and Farallon can be recognised from the deep slabs between ~ 1800 and 1200 km. The double-sided subduction initiated along a zone near the present-day location of Ecuador-Colombia.
2. Eastward rollback of the Mezcalera trench is inferred from the slab rearrangement at 1200-900 km, which is interpreted to mark the initiation of Caribbean plateau at ~ 120 Ma.
3. The presence of NE-trending MEZ and almost vertical Fa slabs between 1200 and 900 km, highlights two accretionary events with the north-western margin of South America including
 - (a) collision of the early Cretaceous island arcs formed in the double-sided subduction zone, which might have created the Romeral and Peltetec terranes.

- (b) collision of the intra-oceanic arcs erupted over the edges of the Caribbean, accreting the Western Cordillera of the North Andes during the Late Campanian.
- 4. Simultaneous to the collision, the eastward subduction of Farallon beneath the western margin of South America initiated ~ 75 Ma. This might show a new period of magmatism along the Central Andes and can be recognised from the NW-SE trending slabs appearing at and above 900 km.
- 5. A noticeable gap is imaged in the Farallon slabs under northern Peru above 600 km depth, separating the northern (SB) and central segments of Farallon (Fa) slabs. This gap is interpreted as the collision of the Piñon plateau to Ecuador and Colombia during Eocene time.
- 6. Two SE-NW dipping slabs (SB and Ca) appear under North Andes. SB anomaly, between 600 and 300 km depths, is associated with the subduction of Farallon under the Colombian-Piñon plateau, creating coastal Serranía de Baudó terrane. Whereas, Ca anomaly is interpreted as the subduction of the Caribbean under the Eastern Cordillera.

5

Evolution of the Central Andes: Insight from seismic tomography

In this chapter, I investigate the changes in the shape of the subducted plates and their connection with the evolution of the Central Andes. After a brief review of the regional geology in section 5.1, tomo-tectonic reconstructions are explained in section 5.2. I summarise the discussion in section 5.3.

5.1 Geological history of the Central Andes

The central Andes are believed to be a recent case of continental arc, built up by the subduction of oceanic plates under the western margin of South America during the Cenozoic period (Capitanio et al., 2011; Sempere, 2008; Allmendinger et al., 1997). The mountain belt contains one of the greatest plateaux in the world, the Altiplano-Puna, with ~ 4 -km height and 400-km width (Isacks, 1988). Crustal thickness in the central segment of the plateau is comparable with that of collisional orogens, reaching > 70 km near the western magmatic arc (Mamani et al., 2010).

5.1.1 Early Cretaceous tectonic setting

The Late Triassic–Early Cretaceous history of the pre-Andean is broadly explained as extensional and/or post-extensional neutral-stress conditions, after which a transition to the regional shortening led to the mountain building (Horton, 2018b). This abrupt transition is expressed in the sedimentary basins along the Andean belt, where the retroarc basins have been conserved. The lowermost stratigraphic sequences, which are preserved across a series of the structures in the retroarc basins of the Eastern Cordillera, Altiplano and Western Cordillera of the Central Andes exposed marine Maastrichtian-lower Paleocene deposits. In the central Andes of Bolivia, Upper Cretaceous–Cenozoic series, which are preserved in the Eastern Cordillera province of the Andean fold-thrust belt, lie on a lower Paleozoic marine strata. This sudden transition encouraged this idea among geologists that a period of neutral tectonic regime was active before the development of the central Andes, despite the fact that the isotopic aging of the rocks in the central Andes indicated the magmatism have been active during the past 200 myr (Pepper et al., 2016). Mpodozis and Ramos (1990) proposed that the Upper Jurassic-lower Cretaceous Chilean and Argentine Andes were dominated by backarc extension, resembling “Mariana-type” subduction (Uyeda, 1982). They linked the closure of the backarc basin during the mid-Cretaceous to a compressive condition, which resulted from the change in the convergent rate, due to the active drift of South America after being separated from Africa. On the other hand, recent analyses of volcanic rocks in northern Chile proposed that these rocks resembled those from modern island arcs, deposited in a backarc basin during Upper Jurassic times (Rossel et al., 2013).

5.1.2 Initiation of the Central Andes

The central Andes are widely accepted to be a product of the widespread crustal shortening and lithosphere thickening of the upper plate during Cenozoic. However, there is less consensus on the inception of the mountain building. While earlier studies estimated that the shortening, as the main mechanism for the uplift, initiated during Neogene (Isacks, 1988), later works estimated the onset of the shortening as mid-Eocene (~ 45 -50 Ma) (Faccenna et al., 2017; Oncken et al., 2006). The former's timing resulted from the study of the topography and structures of the eastern flank of the plateau, whereas latter's estimation of the shortening was based on the thrust belt shortening preserved east of the active magmatic arc. However, structural and stratigraphic studies reported that the known shortening at the surface accounts for 70-80 % of the crustal thickness of the plateau (Allmendinger et al., 1997). McQuarrie et al. (2005) argued that short-term duration for shortening does not account for the deformation of the Eastern Cordillera, due to an incomplete understanding of the stratigraphy, uplift and low preservation of Tertiary strata. They suggested that addition of ~ 30 myr of shortening as early as Late Cretaceous, which is constrained by eastward migration of a foreland basin/fold-thrust system, is required to take mid-Tertiary sedimentary history of Altiplano and Eastern Cordillera into account. Structural and stratigraphic evidence for the late Cretaceous shortening have been reported in the forearc regions of northern Chile (Arriagada et al., 2006). One of the northern Chile's basin is Salar de Atacama, which was inferred to be formed in a foreland setting (Arriagada et al., 2006). The presence of a thrust wedge in the western edge of the basin was interpreted to be coeval with a eastward motion of thrusting during mid-Cretaceous.

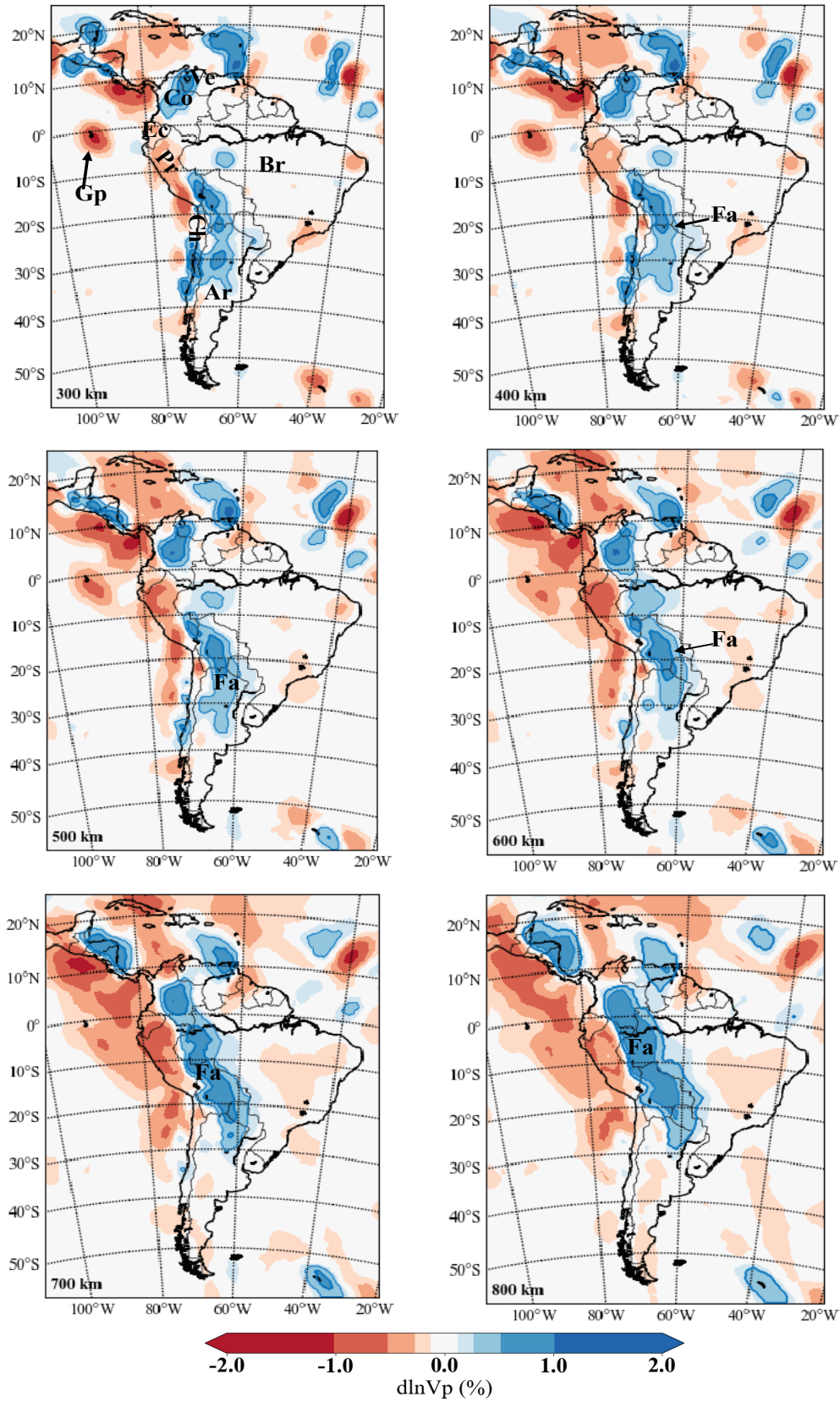


Figure 5.1: Map view of V_p anomaly at selected depths between 300 and 800 km. Ar: Argentina, Br: Brazil, Ch: Chile, Co: Colombia, Ec: Ecuador, Pr: Prue, Ve: Venezuela, and Fa: Farallon slab.

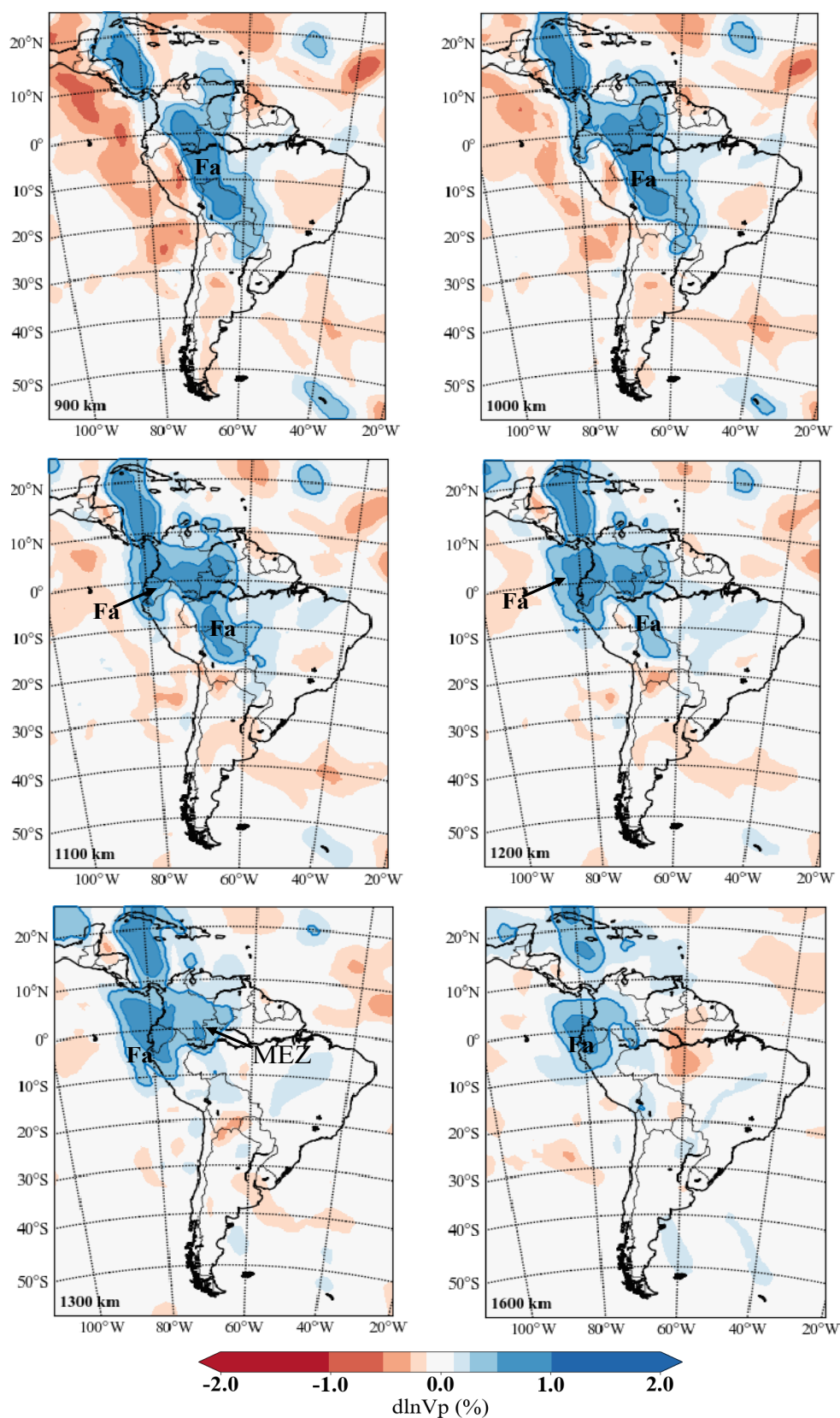


Figure 5.2: Map view of V_p anomaly at selected depths between 900 and 1600 km. MEZ: Mezcalera, and Fa: Farallon slabs.

5.2 Tomo-tectonic reconstructions

Most studies agree that the interaction between South America and oceanic plates of the eastern Pacific basin plays a key role in the tectonic setting of the Central Andes. Researchers who favoured the dynamics of subducted Nazca as the primary driver of deformation have studied the geometry of the Wadati-Benioff Zone (Gephart, 1994; Gutscher et al., 1999; Gutscher et al., 2000), in search of features that may express the fundamental tectonic processes. The Nazca subduction zone is a present-day feature, the past evolution of which needs to be constrained. Imposing kinematic constraints on the slab geometry may help to investigate dynamic changes in the geometry of the slab better. In the following sections, I compare the tomo-tectonic reconstructions with geological observations in the region and attempt to answer the following main questions:

1. Which constraints does mantle structure put on the start time of Andean mountain building?
2. What do tomo-tectonic reconstructions tell us about shortening and extension episodes in the Andes?
3. Do deep structures contain signatures about the formation of the Bolivian orocline?

Interpretation is structured in two main steps. First section 5.2.2 discusses the mantle constraints on the late-Cretaceous history of the Central Andes and compares them with geological data. In the second part, mantle structures are examined against the Cenozoic history of the Andes in section 5.2.3.

5.2.1 Summary of the reconstructions

I proceed with the working hypothesis that depth is linked to subduction time as follows (as detailed previously in Chapter 4):

1. Slabs are assumed to sink vertically in a constant average rate of $\sim 10 \text{ mm yr}^{-1}$ in the mantle.

2. In the first 200 km (from the surface to 200 km depth) the slabs' rate is presumed to be $\sim 20 \text{ mmyr}^{-1}$.

Thus, the slabs above 200 km are assigned to 10 Ma, while below 200 km, every 100 km is assigned to 10 Ma.

To make tomo-tectonic reconstructions in figs. 5.4 and 5.5, P-wave velocity anomalies with amplitude $> 0.25\%$ were contoured out and coloured by depth. The colour changes every 200 km. To better present the geometry of the slabs, only the slab layer that corresponds to the indicated time is displayed, but not the older slabs underneath. The slab depositions are superimposed on the plate motion and seafloor spreading. The vertical dimension of the slab is referred to as the 'height' of the slab, while its 'length' and 'width' refer to the trench-parallel and trench-perpendicular dimensions, respectively. Solid lines with barbs represent the paleo-trench positions, which were inferred from the shape of the slabs at a 50-km depth increment. The polarity of subduction is inferred from the geometry of the slabs. Present-day location of coastlines are drawn with a light silver solid line, whereas reconstructed coastlines and magnetic reversal isochrones at each respective time interval are drawn with a dark grey solid line.

5.2.2 Mid-Late Cretaceous reconstructions: Are the Andes collision-free?

Figures 5.1 and 5.2 indicate the current tomography between 300 and 1600 km depths. As is evident, no slab is observed under the Central Andes below ~ 800 km, which is inferred to correspond to times prior to ~ 70 Ma. This slab gap might result from the low resolution of the current P-wave tomography in the south-eastern region of the continent, as some models (Amaru, 2007) show the presence of slabs in this area. Figure 5.3 depicts different tomographies that recovered this anomaly. This mid-mantle slab, which is not recovered in my model, is not located along the western border, but rather is pushed towards the central longitudes ($\sim 59^\circ\text{S}$ - $\sim 50^\circ\text{S}$) of the continent. At ~ 1500 km depth, it occupies an area between $\sim 29^\circ\text{S}$ and $\sim 7^\circ\text{S}$ latitudes. Its northern end seems to be connected

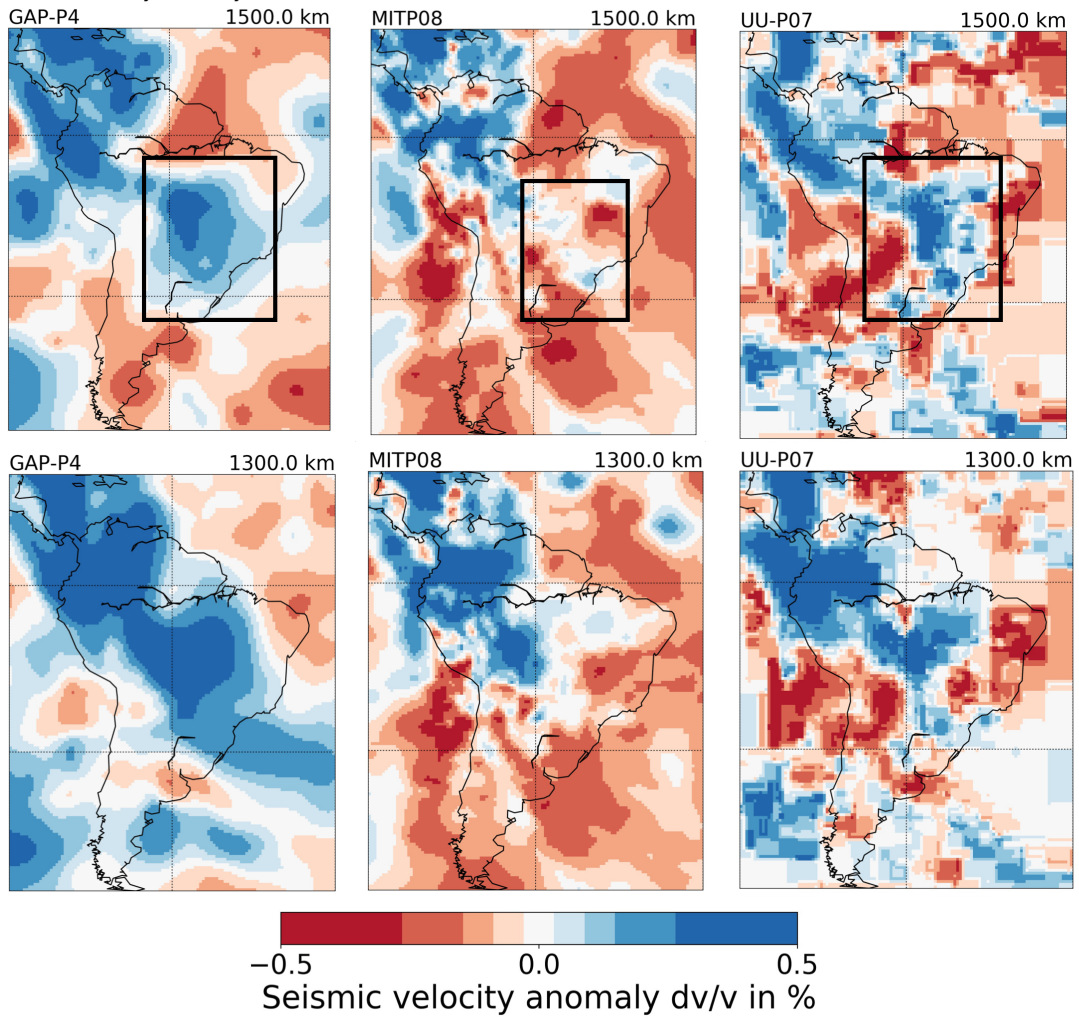


Figure 5.3: Tomographic maps at 1500 and 1300 km depth from previous global P-wave tomography studies: models GAP-P4 (Obayashi et al., 2013), MITP08 (Li et al., 2008), UU-P07 (Amaru, 2007). The images were created using SubMachine online tools (Hosseini et al., 2018). The black box shows the location of the intra-oceanic slab, explained in section 5.2.2

to the north-western double-sided subduction of MEZ-Fa in GAP-P4 (Obayashi et al., 2013), UU-P07 (Amaru, 2007) in fig. 5.3.

Under my vertical sinking assumption, this slab seems to be developed on an intra-oceanic subduction, rather than a continental arc along the Andes. Similarly, Chen et al. (2019) proposed a Mesozoic intra-oceanic setting for the presence of this south-eastern anomaly in MITP08 tomographic model (Li et al., 2008), using a different approach. They used an unfolding method to restore the subduction history.

If this slab shows an intra-oceanic setting, then it might have collided with the proto-Central Andes during late Cretaceous. These tomographic observations conflict with the traditional view about the Central Andes, which argues the mountain belt did not involve any collisional process (Sempere, 2008). However, I propose that geological observations i.e. 1) the sharp deviation in the sedimentary provinces from western Andean to older eastern cratonic sources (Horton, 2018b), and 2) the island arc affinity of the volcanic rocks of northern Chile, could indicate that an intraoceanic-collision process was involved in the genesis of the Central Andes during late Cretaceous. In other words, prior to the Paleocene, subduction might have been running in an intra-oceanic setting, in addition to or instead of under the margin of South America. If this is the case, then the presence of the tomographic slab in the mid-mantle (in fig. 5.3) is justifiable.

Collision of the hypothetical intra-oceanic arcs with proto-Andes might support the theory that the Andean building initiated around Late Cretaceous. Probably the compressional stress resulting from the collision transferred to the continent and provided the necessary force to shorten the lithosphere, which led to the initial thrusts and the growth of initial topography of the Andes. This early topography, which was almost coeval with the eastward subduction of Farallon, might have resulted in the development of a foreland basin during late Cretaceous-Paleocene. This argument might support the Late Cretaceous initiation of the Central Andes' uplift, as was inferred from sedimentary basins in the Andes (Arriagada et al., 2006).

5.2.3 Slab configuration and Andean history during Cenozoic

Figures 5.4 and 5.5 illustrate tomo-tectonic integrations at ~ 70 -30 Ma. Three main features can be inferred from these images, including variations in the curvature, the width, and the strike of the slab.

Figure 5.4 shows that the southern edge of Fa slab is bent at ~ 70 Ma, while the northern part of the slab is rather straight. The bending migrates northward with

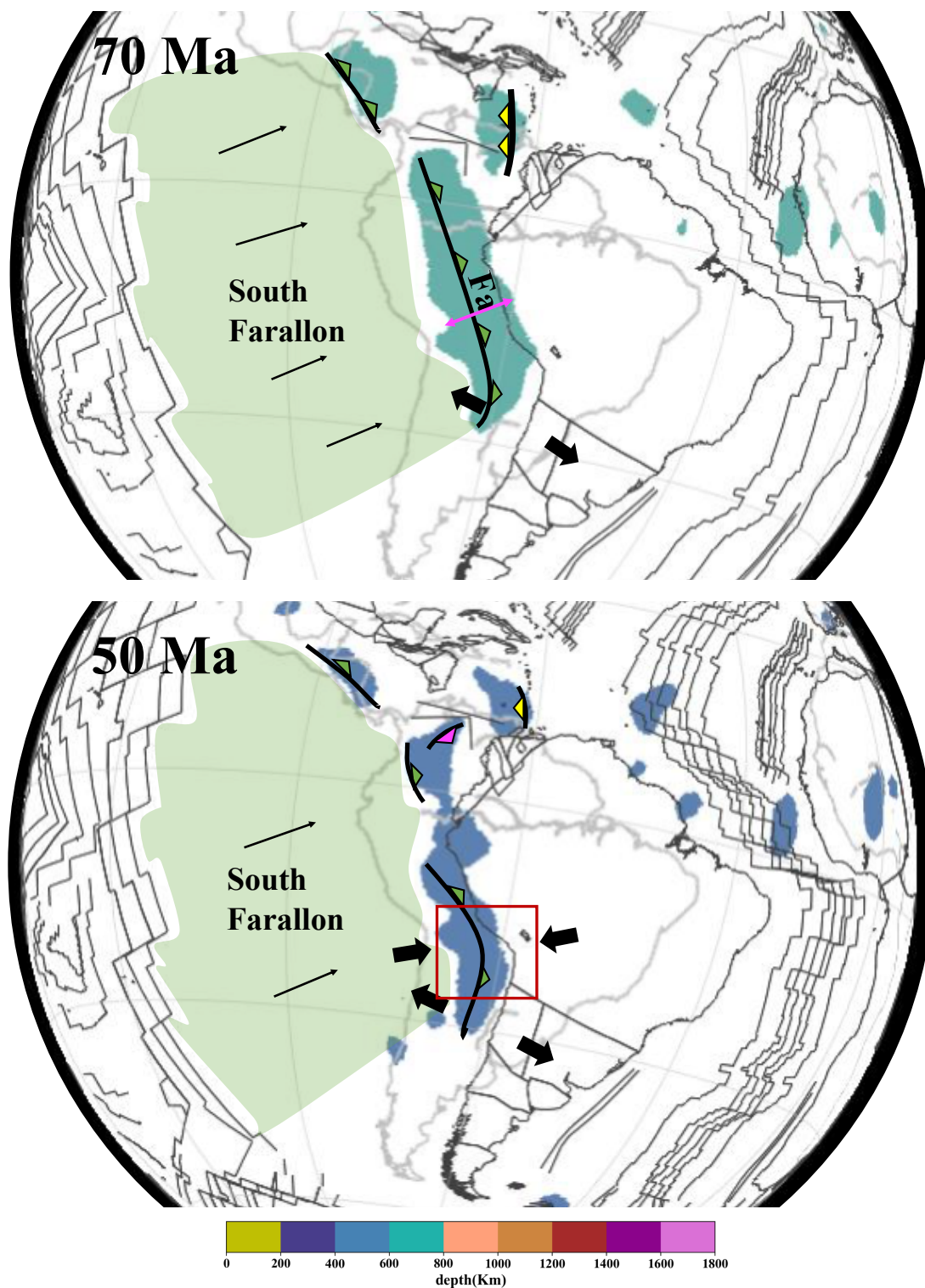


Figure 5.4: Tomo-tectonic reconstructions of the central Andes at 70 and 50 Ma, showing subduction of Farallon (Fa) under the western margin of South America. Black thick arrows indicate the stress regime resulting from Fa slab bending, while black thin arrows point in the direction of the subduction. The red box shows the central segment with higher lateral width, where the slab is curved.

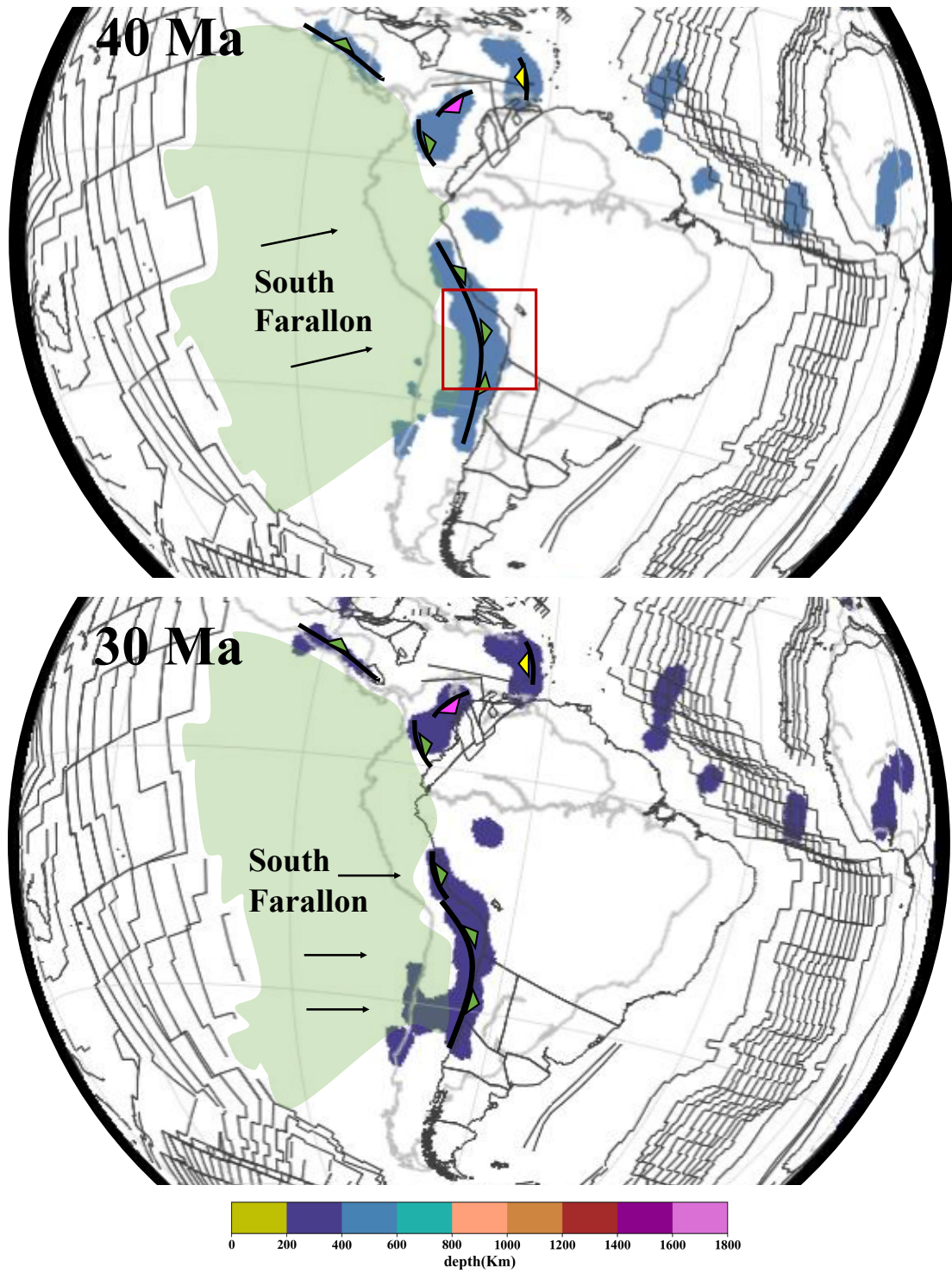


Figure 5.5: Tomo-tectonic reconstructions of the Andes between 40 and 30 Ma similar to fig. 5.4. The oblique subduction at ~ 50 Ma turns to normal at ~ 30 Ma.

the South American plate until ~ 50 Ma, when Fa slab roughly conforms to the shape of the westward-concave margin of Bolivia, which is always shown in its present-day coastlines. From 30-50 Ma, the curvature of the slab no longer changes much, and specifically it does not increase. At 50 Ma the bending of the central part of the slab and the rotation of the northern limb happened simultaneously. A concavity can also be observed along the western edge of the Fa slab.

The second noticeable characteristic of the slab geometry is rough along-strike variations in its width. At 700 km depth (~ 70 Ma), when Farallon began subducting under South America, it appears as a laterally wide slab. But the width of the slab seems to be reduced with time. At 50 Ma the central segment of the slab is quite wide, while the width of the slab is reduced towards the northern and southern edges.

The third characteristic of the slabs in figs. 5.4 and 5.5 is a clear variation in the strike of the Fa slab, which gradually changes from the straight at ~ 70 Ma to the concaved slab at and after ~ 50 Ma and rather NS striking at ~ 30 Ma. Based on the above observations, I connect the curved shape of the Fa slab to the Bolivian bending and divide my interpretation in the following arguments:

1. A robust conclusion is that increased slab curvature from deep to shallow reflects increased coastline curvature from older to more recent times.
2. If my depth-to-age conversion is correct, then the absolute time connected to orocline formation is interpreted to be ~ 50 Ma. Although the curvature appears first around 70 Ma in the southern segment of the slab, the central segment of the slab is curved at ~ 50 Ma.
3. The variation in the strike of the Fa slab might be associated to the rotation history of the Andes.

5.2.3.1 Bolivian orocline

One of the significant features in the Central Andes is Bolivian orocline, a noticeable seaward bending in the central segment of the plateau between $\sim 13^\circ\text{S}$ and 29°S

latitude. From the spatial correlation between the forearc shape and the present-day geometry of Nazca, Isacks (1988) argued that the Bolivian curvature formed during Neogene. However, later paleomagnetic data proposed an Eocene-Oligocene time span for the formation of the bending (Roperch et al., 2000). The onset of the formation of the curved slab, in my timing, encourages the idea that the Bolivian curvature formed ~ 50 Ma, which is close to what paleomagnetic data suggested. The contemporaneous occurrence of the convexity and widening of the central part of the Fa slab might mark the link between Bolivian orocline and the surface deformation. In my time-to-depth conversion, the widening is concurrent with the Incaic phase of intense deformation in the Central Andes. James and Sacks (1999) observed a similar widening in the seismicity of the region and connected it to the occurrence of a flat subduction. They suggested that the extensive crustal deformation of the inner arc during late Eocene-early Oligocene time resulted from the transition from normal to flat subduction. Inner arc was defined as a region between eastern Altiplano and Eastern Cordillera, which is located under the widened part of the Fa slab.

Martinod et al. (2010), connected the formation of Bolivian curvature to the occurrence of this hypothetical flat subduction and proposed the flat slab may be caused by the subduction of a buoyant volcanic plateau late Eocene. This suggestion seem intriguing, because of the presence of a concavity along the western edge of the Fa slab, which might suggest the presence of a heterogeneity on Farallon as a possible driver of the curvature. Structural and stratigraphic study by McQuarrie et al. (2005) also linked the cessation of the Western Cordillera arc volcanism in southern Peru at ~ 50 Ma and in northern Chile at ~ 38 Ma to the initiation of the flat slab during Eocene. This timing is rather close to the slab widening in my model.

However, there is a problem with occurrence of a flat subduction during Early Eocene, when the slab is widened in my model. Present-day flat subduction zones in Peru and Chile are coincident with the gaps in volcanism. This means that we should expect a region of no magmatism during this widening period (~ 50 -30 Ma).

On the other hand, magmatic analyses showed that the volcanism in the central segment of the Andes was active during Upper Cretaceous to Late Eocene (Mamani et al., 2010). However, the Isotopic ageing of magmatic samples by Pepper et al. (2016) (e.g. Fig.3, or Fig.13 therein), seems to indicate a reduction at ~ 50 to ~ 30 Ma. Although the writers proposed magmatism at this period, my personal inference is that the diagrams at this period reach a minimum, which might reflect the low or lack of magmatism during this time slot. If this is the case, then we might be able to link this widening to a flat subduction. However, I would not emphasise on this hypothesis.

The proximity of the site of the Bolivian indentation and that of the high plateau may strengthen the idea that they would have occurred at the same time. Two constraints for this argument are the shortening and rotation history (Isacks, 1988). As it was pointed out earlier, paleomagnetic analyses by Roperch et al. (2000) suggested that the curved shape of the margin formed during early to mid-Tertiary period.

As was explained in section 5.1, stratigraphic and structural studies indicated that the late Cretaceous shortening advanced eastward during Eocene (McQuarrie et al., 2005; Decelles et al., 2011). This inboard shift was accompanied by the migration of magmatic arc to the Western Cordillera , and a regional broadening in the foreland flexural basins, which was interpreted as the expansion of the orogen. By Eocene-Oligocene, much of the orogen including Altiplano-Puna and Eastern Cordillera were suggested to be affected by the regional shortening (Horton, 2018a). Therefore, I conclude that the formation of the Bolivian orocline were simultaneous to the second phase of the shortening during early Paleogene and the slab widening in my model might be related to the eastward expansion of the orogen.

5.2.3.2 Slab rotation and variations of stress regimes

I proposed that the change of the slab's strike might indicate the rotation history of the upper plate. The current model proposes that the southern limb of the slab was rotated first at ~ 70 Ma, after which, at ~ 50 Ma, the northern and the central

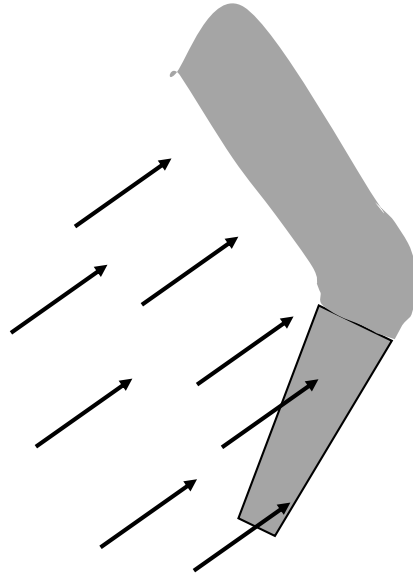


Figure 5.6: Block rotation in the central Andes during Eocene. Black arrows show the subduction orientation. Gray blocks represent the Farallon segments. The oblique subduction in the southern block (below $\sim 19^\circ$ S) transfer weaker stress to the upper plate, causing a weaker plate coupling.

segments of the slab were curved simultaneously. This concurrency might imply that the curvature is partly a consequence of the rotation of the northern limb of the Andes, as was suggested by magnetic anomaly analyses of Roperch et al. (2000).

The implication of this slab rotation is to investigate the possible paleo-stress settings along the convergent boundary. Considering the oblique subduction during most of the Eocene (Pardo-Casas and Molnar, 1987; Beck, 1998), it is inferred that the rotated southern limb of the slab regulated a weaker plate coupling, hence an extensional mode. By contrast, the strong coupling in the northern and central part of the Andes transmitted a compressive stress.

This contrasting regime has also been recorded in the surface data. While the extensive shortening was documented for the Cenozoic history of central block of the Andes (McQuarrie et al., 2005), an extension history has been reported by different studies for the northern Chile (Horton, 2018a; Charrier et al., 2002). Magnetic anomaly analyses by Pardo-Casas and Molnar (1987) also suggested a

rapid, dextral-transform interaction, possibly with some extension, for the latest Cretaceous to Early Eocene history of northern Chile.

5.3 Central Andes synthesis

Tomo-tectonic results relating the subduction history is interpreted in terms of a new model for the tectonic development of the Central Andes. The model is based on the geometry of subducted slabs and their inferred sinking rates, and is discussed in light of complementary observations of deformation history and paleo-magnetism. It includes the following stages for Cretaceous to Oligocene evolution of the Andes:

1. During Lower Cretaceous times, no subduction under Central Andean continental margin is observed in tomo-tectonic reconstructions. However, intra-oceanic subductions is likely to have operated to the west of the margin during this period. This interpretation was inferred from the slab configuration deeper than ~ 800 km.
2. Hypothetical island arcs formed in stage 1 collided with the western margin of the continent during Upper Cretaceous times. The collision transmitted significant compressive stress to the upper plate, which resulted in the inception of the initial topography in the Andes ~ 70 Ma. The initiation of the Andes was concurrent with an east-dipping subduction of Farallon, which can be observed in the current tomography in shallower depths $< \sim 800$ km.
3. A curvature in the subducted Farallon slab can be observed at and above ~ 600 km, which was proposed to be indicative of the formation of the Bolivian orocline during the Eocene to mid-Oligocene (~ 50 -30 Ma).
4. The rotation of the southern segment of the slab ($\sim 19^\circ$ to 30° S) might show that the southern block of the Andes was rotated earlier at ~ 70 Ma, while that of the northern segment (above $\sim 20^\circ$ S) might have commenced at the same time as the curvature of the Bolivian orocline.

5. This localised rotation of the slab is interpreted to show a mixed tectonic regime on the upper plate. While strong plate coupling between the convex part of the slab and upper plate led to a compressive regime in the central and northern segments, the southern limb of the Andes was under extension.

This model shows that the orogenesis in the Central Andes involve a combination of neutral, extension and compression histories, which I argued to provide a valid match to geological observations. I attempted to answer three main questions in the central Andes. First I highlighted the role of a late Cretaceous collision in the formation of the early topography in the Central Andes. Second I inferred from deep slab observations in the mantle when the Bolivian orocline formed. Finally, I described how the higher magnitude of plate coupling might have led to the contrasting deformation regimes along the Central Andes during Eocene-early Oligocene time.

6

Summary and Conclusions

In this thesis two major topics were addressed. First in chapters 2 and 3, I discussed the theory and the results of the multi-frequency tomography under South America. The sharper geometry of the results of the new tomography compared with the previous models, enabled me to have a more accurate inference of the subduction history under the continent. In the second part of the thesis, I integrated the new model with kinematic models, to make a new interpretation of the mantle structures. Inspired by the segmented geometry of the slab configuration, it turned out there is a noticeable correlation between slab distributions and the evolution of the Andean mountain belt. I investigated this correlation in chapters 4 and 5.

P-wave tomographic model

An important aspect of this study is the calculation of finite-frequency kernels in multiple frequency bands. Source Time Functions, which are necessary to estimate more realistic synthetic seismograms, were prepared for 2000 earthquakes. In addition to multi-frequency kernels, hand-picked data from ISC-EHB catalogue was included to improve the data coverage in the regions of the continent, where broadband seismograms are rare. Most of the recovered structures underlay the western side of the continent, which is well sampled. A main problem in this

study is that poorly sampled regions in the east remain unsolved by direct P-wave tomography. This is a serious problem in the case of studying early Cretaceous history of the Andes. Therefore, incorporation of other phases might improve the sampling, hence the inversion results, in these regions. This improvement can be observed in the previous models which used later arrivals. I employed the results of other studies in these regions in my interpretations.

I demonstrated the new model and compared it with previous P-wave models in chapter 3. I explained that four slab units can be recognised from the high velocity anomalies in the mantle down to 1800 km depths. These slab segments were classified according to their latitudes, depth and strikes. Two flat subduction zones, which are important features in south America, are recovered well in my model. The presence of both the segmentation and flat subduction settings were compared with those in previous models in chapter 3.

Paleo-reconstructions and evolution of the Andes

The slab configuration, which was explained in chapter 3, was combined with plate tectonics model to make a sufficient interpretation. To integrate plate motion and mantle structures, I assumed slabs sink vertically. This assumption, which was inspired by wall-like volumes of slabs in the tomography, was tested in chapters 4 and 5 through geological records. The sinking rate was a second null hypothesis, which was examined by geological observations in the region. It turned out the sinking rate of $\sim 10 \text{ mm yr}^{-1}$ in the mantle results in the best correlation. Under this reconstruction I provided a reasonable explanation for the contradictory evolution of the Andean mountain belt since early Cretaceous, which is summarised below.

A double-sided subduction zone and accretion of Romeral-Peltetec terrane

Two intra-oceanic subductions (Farallon and Mezcalera oceans) were initiated far to the west of the western margin of South America during Upper Jurassic-

Early Cretaceous. The double-sided subduction zone may have been terminated by the opening of the Caribbean ocean plateau at ~ 120 Ma. The inception of Caribbean plateau was suggested to be accompanied by the collision of the island arcs, formed in the double-sided zone, with the western margin of South America. This collision might have led to the accretion of the Romeral and Peltetec terranes on the Central Andes.

Western Cordillera of the North Andes

Inspired by the correlation of the site and time of the migrated double slabs with those of the Caribbean plateau's initiation, I proposed that the related island arcs were erupted over the Caribbean plateau since ~ 120 Ma. Late-Cretaceous was marked by the collision of these arcs with the north-western margin of South America, which led to the amalgamation of the Western Cordillera during the Late Campanian. This was inferred from the shape of the island arcs, which conforms to the north-western margin of South America.

Coastal Cordillera of the North Andes

A noticeable slab gap was recovered at ~ 600 - 300 km depths in the tomographic model under Peruvian flat subduction. The gap, which appears under Ecuador during Eocene in the current tomo-tectonic reconstructions, was interpreted as the collision of the oceanic Piñon plateau with Ecuador, forming coastal Cordillera of Ecuador during Eocene.

The Piñon gap broke Farallon into two segments. The northern part of the Farallon subduction formed an oceanic arc along the edge of the plateau. The island arc was proposed to be collided with western Colombia during Oligocene, forming the coastal terrane of Serranía de Baudó.

Evolution of the Central Andes

Under my time-depth conversion, subduction of Farallon plate under proto-Central Andes initiated ~ 70 Ma, prior to which the current P-wave tomography does not show any subduction. However, some of the previous studies indicated a mid-mantle slab under the south-eastern margin of South America. The mid-mantle slab, was interpreted as an intra-oceanic arc by previous studies. In my reconstructions, the intra-oceanic arc might have collided with South America before the subduction of Farallon. I proposed that this collision might show the initiation of the mountain building in the Central Andes. This collision may explain why the initial shortening, which is documented by geological studies, commenced during late Cretaceous in the Andes. The Late Cretaceous collision of the intra-oceanic arcs to the western margin might have created necessary compressional force for the growth of the shortening.

South America and the southern Caribbean boundary

The presence of a NE slab under Venezuela was interpreted as the subduction of the Caribbean plate under Eastern Cordillera of the Northern Andes since early Cenozoic. This observation was also made by previous tomographic and seismicity studies.

Heterogeneities on the Farallon slab: Bolivian Orocline

In the current reconstructions, Cenozoic-to-Present subduction of Farallon is characterised by a significant bending of the slab. The curvature, which appeared first at ~ 800 km depth (equivalent to ~ 70 Ma) in the southern verge of Farallon, migrate to the central segment of the slab at ~ 600 km depth (~ 50 Ma). Conforming to the present-day margin of South America, the curvature was inferred to be observed until ~ 30 Ma. This feature was linked to the formation of the Bolivian Orocline ~ 50 - 30 Ma. The widened Farallon slab at this time was connected to a period of shortening, which was inferred from the sedimentary and structural

studies. Appearance of the bending in the central segment of Farallon, concurrent to the rotation of the southern limb of the slab, was proposed to indicate a period of contrasting regime in the Central Andes during Eocene to early Oligocene time, when shortening and extension occurred rather simultaneously.

Appendices

A

The uneven data coverage in South America increases the non-uniqueness of the tomographic inversion. The effects of ill-posedness can be reduced by regularisation. Yet marginal changes in the regularisation (order of magnitude 10^{-3}) results in a significant variations of the chi-squared values (order of magnitude 10^{-1}) in my model. However, the main structures along the western regions, which are sampled well by the kernels, do not show a noticeable variance. The main changes in the solutions are mostly caused by poorly sampled areas, i.e. under the eastern regions of the continent. In fact, solutions which are less penalised do not fit the data anymore. as there is not enough data in the poorly sampled areas. Thus, artefacts are created in these regions and the roughness of the problem is increased. The model, which was showed in Chapter 3 (Fig.3.1-3.5), is selected for interpretation in the following chapters. However, I recovered another solution (with $\chi^2 = 1.108$), which is almost similar to the selected model (Chapter 3, figs. 3.1 to 3.5) in major structures. This less-regularised model displays higher amplitude, although in price of creating some less-reliable structures in the low-sampled regions. The model is illustrated in figs. A.1 and A.2 at every 100 km. Cross-section of the Chilean flat slab in this model, is also presented in fig. A.3.

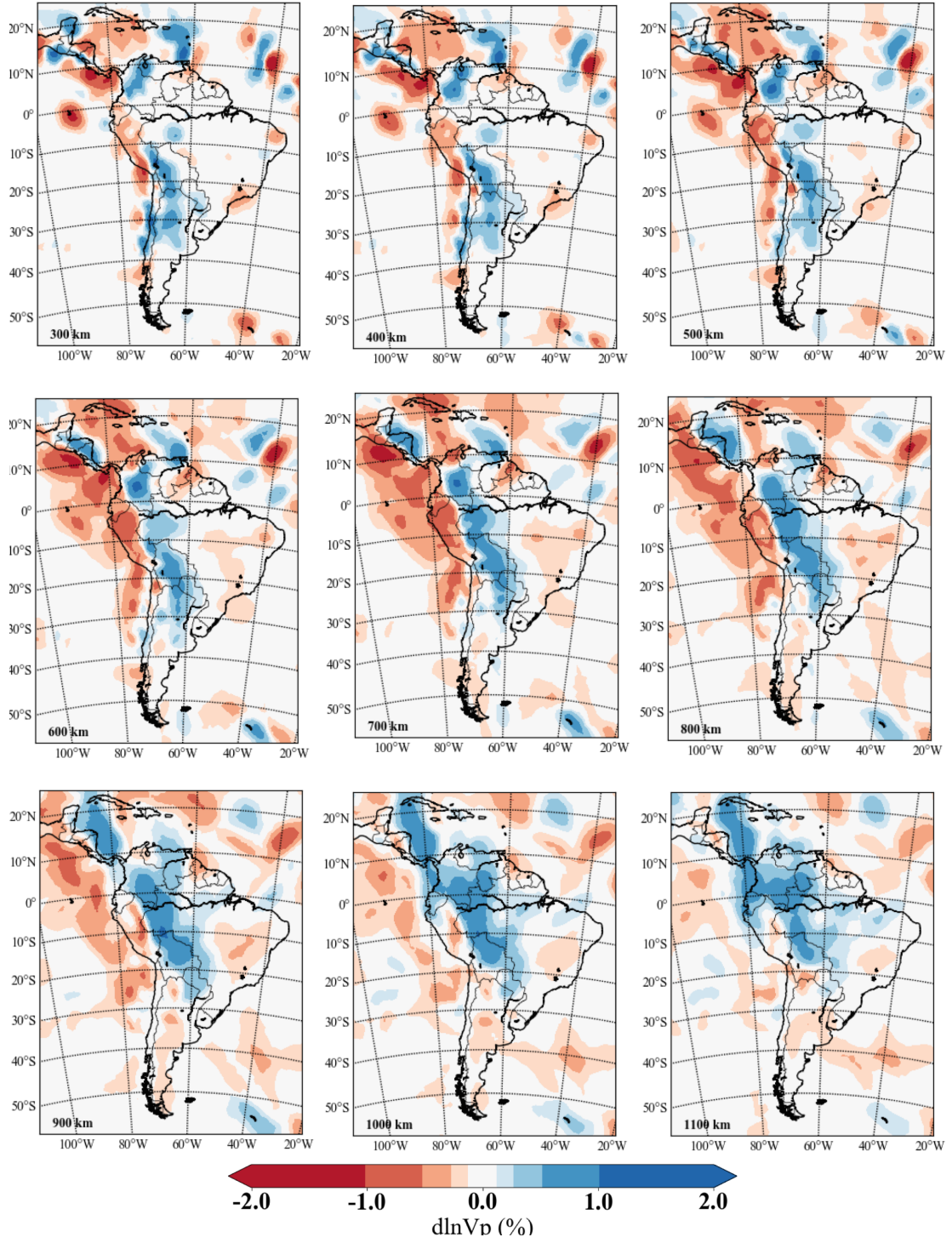


Figure A.1: Map view of V_p anomaly at selected depths between 300 and 1100 km, recovered from the inversion with $\chi^2 = 1.108$.

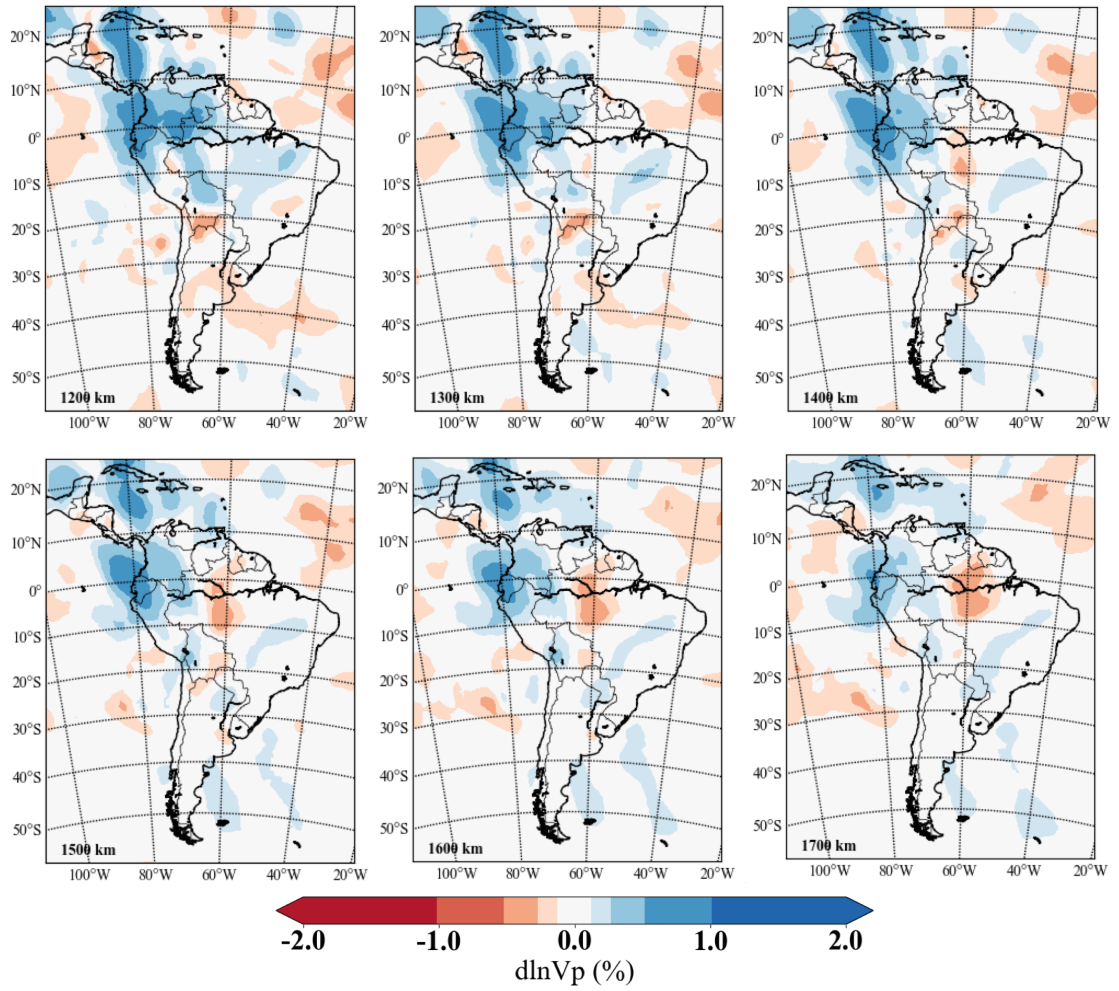


Figure A.2: Map view of V_p anomaly at selected depths between 1200 and 1700 km, recovered from the inversion with $\chi^2 = 1.108$.

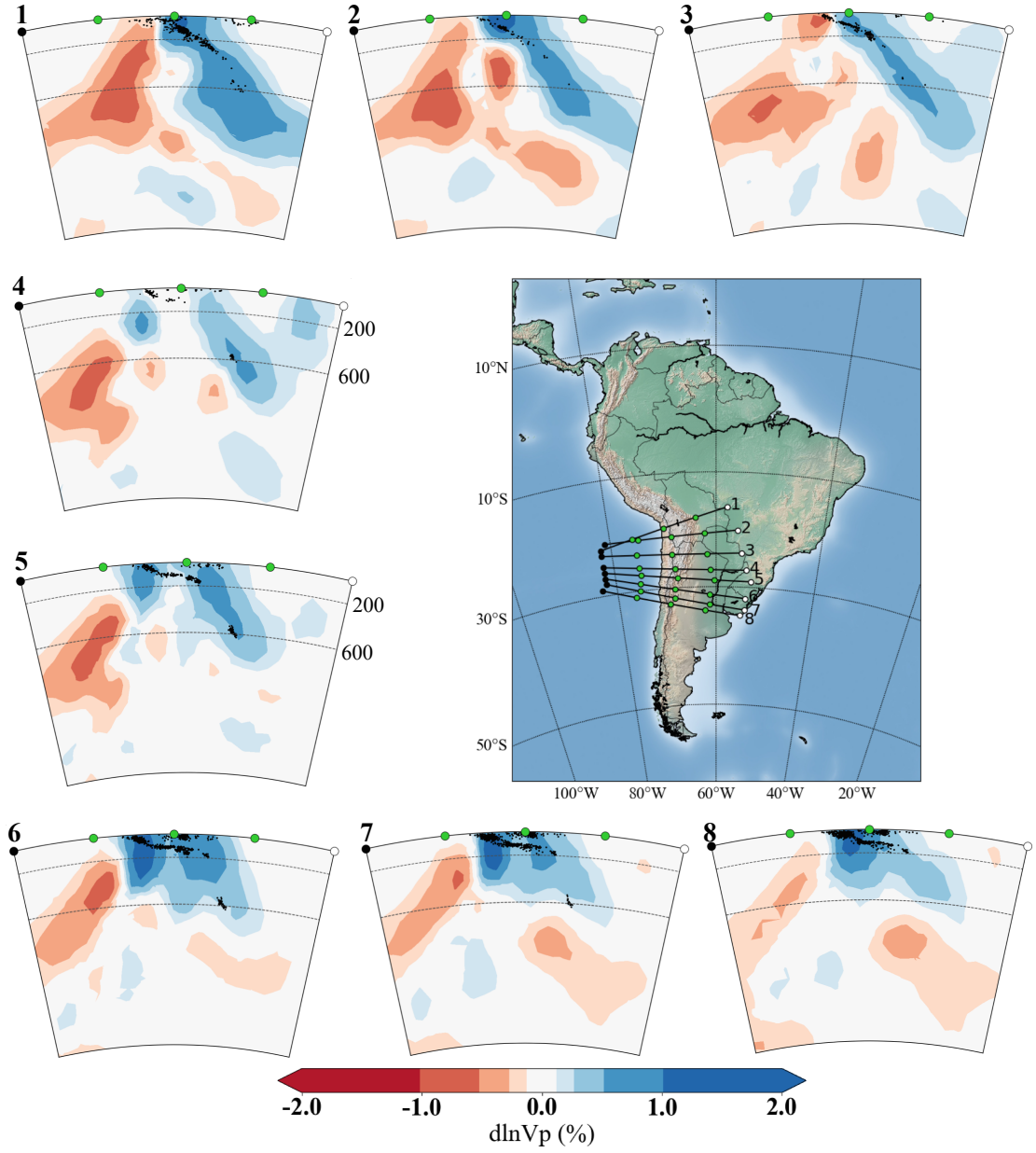


Figure A.3: Map view of northern-Chile region, showing location of 8 cross-sections from the inversion with $\chi^2 = 1.108$. The dashed lines in all cross-sections indicates 200 (shallower line), and 600 km depths. Black circles represent location of seismic sources.

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