AN INVESTIGATION OF THE PAMPEAN FLAT SLAB AND ITS ROLE IN ANDEAN TECTONICS THROUGH THE ANALYSIS OF ENHANCED EARTHQUAKE CATALOGUES

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ABSTRACT

The Andean margin of South America is an archetypal example of a convergent tectonic environment that has played a fundamental role in the creation, evolution, and consumption of both continental and oceanic lithosphere. An enigmatic aspect of this margin is the existence of several regions of flat slab subduction, the causes and consequences of which remain poorly understood despite years of intensive investigation. The research discussed in this dissertation focuses on the Pampean flat slab beneath central Chile and Argentina, partly in response to the discovery of an unusual pattern of P wave arrival times from earthquakes beneath Argentina recorded in Chile, but also because recently developed techniques in automated phase onset estimation allow us to take significantly greater advantage of a substantial archive of seismic data that has been collected in this region over the past several decades. After constructing enhanced catalogues of seismic activity and analyzing them with various techniques of seismic tomography, the improved images of the elastic wavespeed structure of this region allow insights into the state of the subducted Nazca plate, the effects of subducting anomalous bathymetry such as ridges and seamounts shortening on the interaction between the Nazca plate and the overriding South American plate, and the consequences of lithospheric shortening of the South American plate on the geometry of the Nazca plate. We find that dense clusters of seismicity associated with the Pampean flat slab are most likely a consequence of the subduction of ridges and seamounts, and that the extensive alteration and devolatilization of the oceanic lithosphere by these features leads both to an increased buoyancy and to the release of volatiles that hydrate the South American mantle and fracture the lower crust. The existence of these features far from the trench lends support to the contention that subduction erosion is more episodic than continuous. We also find 2 low velocity anomalies, particularly in Vp, within and below the subducting slab. The western
low velocity anomaly correlates with the Juan Fernandez Ridge and is postulated to be due to an increase in silica content or the presence of supercritical fluids. The eastern low velocity anomaly seems to be a result of rising hot asthenosphere. Furthermore, it appears that the gradient in wavespeeds that generated the pattern of anomalous P wave residuals that originally motivated this research may be due more to the low wavespeeds in overthickened crust beneath the high Andes grading into the higher wavespeeds of the Nazca plate, rather than a pronounced lithospheric root beneath a normal crustal root. At the same time, much of the original signal remains, which suggests that the ultimate cause of the higher apparent wavespeeds may be due as much to anisotropy as opposed to lateral heterogeneity.
1. INTRODUCTION AND BACKGROUND

The term “Andean Margin” often is generically applied to any convergent plate boundary where oceanic lithosphere subducts beneath continental lithosphere. These types of margins are important in Earth Science because of the role they play not only in the generation and recycling of continental crust but also in the physical state of the subducted plate boundary and the associated stress variations and metamorphic reactions that give rise to volcanism and megathrust earthquakes. Decades of investigation of the archetypal Andean Margin, located along the western coast of South America, have shown that these margins contain a broad spectrum of local tectonic environments that reflect variations in, for example, oceanic bathymetry and buoyancy, erosion and convergence rates, and changes in the dip and orientation of the subducted slabs in space and time. Because of its size (roughly 14,000 km in length), age (subduction began in the late Paleozoic - Rebolledo & Charrier, 1994; Willner et al., 2008), and relative accessibility, the western margin of South America provides an ideal natural laboratory for examining these environments through targeted investigations.

This dissertation reviews an investigation of a part of the Andean margin in central Chile and Argentina between approximately 28°S and 35°S, often referred to as the “Pampean Flat Slab” region because of a well-documented change in the dip of the subducting Nazca plate from “normal” (~30°) to “flat” over a distance of a few 10s of km. This region has been the subject of numerous previous investigations, reviewed below, which, while providing a number of insights, pose nearly as many questions as they answer. Indeed, the particular scientific motivation for the study discussed here arose from an unusual and somewhat serendipitous observation from a previous investigation of this region, reviewed in the next section, that suggested that our current understanding of this type of environment is fundamentally lacking.
The study that makes up this dissertation is presented in four chapters. The first chapter provides the background and motivation for the work, along with a summary of the seismological analysis performed. The second describes an investigation of the region above the flat slab using Rayleigh waves generated by ambient noise. The third discusses the generation of an extensive dataset of body wave arrival times using an automated catalog generator. The fourth and final chapter describes the combined analyses of the datasets discussed in chapters 2 and 3, along with some additional data from analogous sources, to generate a large-scale elastic wavespeed model of the region of interest. The final part of the last chapter discusses this result in some detail and provides new inferences related to the geodynamics of the region.

1.1 Motivation and Scientific Objectives

The motivation for the study carried out in this dissertation was sparked by an unexpected result of an investigation into the aftershocks of the 2015 M8.3 Illapel Earthquake, recorded by the Chile ILLapel Aftershock eXperiment (CHILLAX) deployment (Comte et al., 2019; Figure 1.1). During the generation of a catalogue of the aftershocks, one of the authors (S. Roecker) noticed that a large number of earthquakes appeared to be occurring east of the network beneath Argentina, and expanded the search volume to see if additional observations could be obtained. Generally, one expects the hypocenters of events that occur outside the aperture of a network to appear scattered because of a greater sensitivity to noise, but the exact opposite happened – in this case the hypocenters outside the network were located in a well-defined deeper Wadati-Benioff zone continuous with that defined by shallower events within the network (Figure 1.2). While this pattern might be what one would prefer from a simple application of Occam’s razor, these hypocenters should reside in the (horizontal) Pampean flat slab and so are unequivocally, and systematically, wrong. To investigate what might cause a systematic mislocation, Roecker
compared these hypocenters to those of equivalent events derived from contemporaneous teleseismic locations reported in the IRIS-Interactive Earthquake Browser (IEB) catalogue, and the longer term Engdahl, van der Hilst and Buland (EHB) catalog based on 28 years of data from 1980 to 2008 (Figure 1.2). The locations from these networks match well from the near surface until ~100 km depth, and then diverge at that depth to the east. Arrival times from one of these divergent events, a M4.9 event on 2015/12/26 that was reported by both the IRIS-IEB and CHILLAX networks, were then calculated for CHILLAX stations from both the CHILLAX and the IRIS-IEB hypocenters using the 1D model used to create the CHILLAX catalog. Residuals at all the CHILLAX stations from the CHILLAX hypocenter were small (<0.1s) and marginal probability density functions (MPDF) of the hypocenter parameters showed no evidence of alternate minima. However, while P-wave residuals from the IRIS-IEB hypocenter were small for the easternmost stations, they systematically increased to as large as -2 seconds to the west (Figure 1.3). This gradient in observed arrival travel times reflects an increase in apparent wavespeed that could be caused by a westward gradient of increasing wavespeeds. A comparison of raypaths from these two hypocenters (Figure 1.2) suggests that the source of this wavespeed anomaly is located somewhere beneath the high Andes between the stations in Chile and the Pampean flat slab beneath Argentina. While offsets due to an incorrect origin time are always possible, to the extent that the absolute residuals are correct, one would conclude that this region has an abnormally high wavespeed.

A consequence of this body being located directly beneath the high Andes is that, because of its relative remoteness and the logistical difficulties of operating seismic networks simultaneously across an international border, the structure of this area is not well constrained by any prior subsurface imaging studies. As discussed in detail below, there have been a number of
both temporary and permanent seismic deployments in this region over the past several decades that operated largely independently of each other, but which overlap sufficiently in time and space to potentially fill in this poorly illuminated part of the high Andes. One of the main objectives of the study carried out in this dissertation is to improve the images of this and surrounding regions by enhancing the arrival time catalogues from one side of the Chile-Argentina border with known locations from the other side of the border.

![Figure 1.1: The CHILLAX deployment. Black triangles are the seismic stations. Circles are aftershock locations of the 2015 Illapel earthquake where the colors represent the depth of the hypocenter as indicated in the palette above the figure (Comte et al., 2019). The large star locates the mainshock. Slab contours in intervals of 10 km depth with thicker lines at intervals of 50km are from Slab 1.0 (Hayes et al., 2012).](image-url)
Figure 1.2: Cross-section of hypocenter locations at ~31.5°S between 73°-68°E. The x-axis is distance in km from 70.5°E, which is close to the eastern edge of the CHILLAX network (Figure 1.1). Crosses represent locations determined from the CHILLAX deployment, red circles are locations from the IRIS/IEB catalogue, and the blue circles are from the EBH catalogue. The green line is the Slab1.0 contour taken at 31.5°N. Red and blue lines are ray paths for the 26/12/2015 M4.9 event from the IRIS/IEB and CHILLAX hypocenters, respectively, to the stations in the CHILLAX network.

Figure 1.3: P residuals (absolute value – all residuals are negative) from the comparison of travel times from the IRIS/IEB and CHILLAX hypocenters at the CHILLAX stations for the event shown in Figure 1.2. The size and color of the circles correspond to the size of the residuals. The orange diamond locates the event epicenter. Slab contours in intervals of 10 km depth with thicker lines at intervals of 50 km are from Slab 1.0 (Hayes et al., 2012).
An initial evaluation of this concept was carried out using data from a subset of these networks (CHILLAX, CHARMSME, and CHARGE), primarily to find out how many more observations might be available from a reanalysis of these datasets. Hypocenters from the resulting catalogues (Figure 1.4) show that the slab is not actually flat, but deflected upward in the region of the travel time anomaly just under the high Andes until it shallows by ~25 km beneath Argentina, after which it re-subducts in the east back into the mantle. Similar features were noted in other studies (e.g., Anderson et al., 2007; Alvarado et al., 2009; Linkimer et al., 2020) although largely downplayed as a probable artifact of poorly constrained hypocenters. The bending of the slab back into the mantle has been hypothesized to be a result of either (1) buckling due to resistance to subduction from the east or (2) negative buoyancy of the slab underneath Argentina due to some anomaly in that region (e.g., the presence of a subducted Juan Fernandez Ridge, dehydration of the slab, or episodic subduction erosion of the South American crust). A preliminary local earthquake tomography image using the enhanced catalogues (Figure 1.5) provide some support for the negative buoyancy hypothesis in the form of lower Vp and Vs in the upper part of the seismic zone that could be caused by phase conversions or subduction-eroded South American crust (e.g., Comte et al., 2019; Stern, 2020). However, the ray coverage beneath the high Andes from these particular networks was insufficient to image the structure above the initial bend in the slab.

The fate of the South American lithosphere as it shortens in response to its convergence with the Nazca plate has been the subject of prolonged debate. Competing hypotheses include the formation of a lithospheric root that periodically detaches and sinks into the underlying mantle, versus a lithosphere that it is continuously consumed by attaching (“ablating”) onto the Nazca plate beyond the “S” or subduction point and descending, escalator-like, into the depths of the mantle.
Both of these scenarios are easily visualized in “normal” subduction environments which allow unencumbered consumption of the continental lithosphere but seem less plausible in a flat-slab environment due to its cul-de-sac nature – there simply is nowhere for the shortened or ablated lithosphere to go. We hypothesize that the initial concave upward bending of the Nazca plate is a response to deformation of a nominally flat slab caused by localized shortening of the South American lithosphere. Such a high-density lithospheric root beneath the high Andes would be prevented from detaching via a Raleigh-Taylor instability by the Pampean flat slab, and could explain the relatively fast P arrival times observed at the coast of Chile.

Figure 1.4: Hypocenter locations along latitude 31.5°S, from a preliminary analysis of augmenting catalogues from the CHILLAX, CHARSMEME, and CHARGE networks in Chile and Argentina.

Figure 1.5: Results from a local earthquake tomography using the data from the augmented catalogue of the CHILLAX, CHARSMEME, and CHARGE networks. (A) The Vp image and (B) The Vs image. Hypocenter locations are plotted as white circles on the bottom panels.
An obvious question to ask about this hypothesis would be: is it physically plausible for a high-density body to load a slab sufficiently to create the observed deflection? To address this question, we assumed three simple but relevant geodynamic frameworks (Figure 1.6) described in Turcotte & Schubert (2014); (1) an elastic plate under a line load, (2) a beam pinned at both ends under a uniform load, and (3) a beam pinned at both ends under a sinusoidal load. Based on the anticipated wavespeed anomaly, we considered a maximum increase in density of the lithosphere of 15% from a base of 3300 kg/m$^3$, a maximum length of 175 km, a maximum height of 60 km (assuming the entire lower continental lithosphere transitioned into this denser material) and a maximum deflection of 20 km. Contreras-Reyes & Osses (2010) suggest that the Nazca plate to the south of our study area is exceptionally weak, and we use their estimate of flexural rigidity for the plate that is some 90% less than that of normal oceanic lithosphere. Note that in the region of the flat slab, the Nazca plate may be even weaker because of the presence of the Juan Fernandez Ridge. We found the line load was unlikely to create the observed deflection, but that a 150 x 40 km load with an 11% (uniform) or 13.5% (sinusoidal) increase in density could produce a 15 km deflection. Of course, this result does not “prove” the existence of a high-density body beneath the Andes, but does show that it could be a plausible cause of the observed deflection and hence worth investigating further. Testing this hypothesis, along with determining the general physical state of the mantle in the vicinity of the Pampean flat slab, constitute the principle scientific objectives of the research summarized in this dissertation.
1.2 Geologic and Tectonic Background

1.2.1 Flat Slab Subduction

By definition, flat slab subduction involves the subducting plate moving horizontally beneath the overriding plate mantle for some extended distance prior to eventually descending into the mantle. Flat slabs are estimated to occur in about 10% of all subduction zones (Gutsher et al., 2000), although some (e.g., Manea et al., 2017) would argue that flat slabs really only occur in Mexico and along the Andean margin in Peru and Chile-Argentina (e.g., the Pampean flat slab) while the other regions are actually normal subduction zones with gently dipping slabs or various complex geometries. Nevertheless, most geologists would agree that flat slab subduction has played a major role in the evolution of western North America (e.g., Coney & Reynolds, 1977) and perhaps Tibet (e.g., Ding et al., 2003).

Along the Andean margin, one of the easily recognizable consequences of flat slab subduction at the surface is the disruption of volcanic activity (Figure 1.7), presumably due to the
removal of asthenosphere resulting in cooler temperatures above the slab that inhibit melting (e.g. Kay et al., 1988; Gutscher et al., 2000; Kay & Mpodozis, 2002; Ramos et al., 2002; Alvarado et al., 2007; Gerya et al., 2009). These regions therefore experience gaps in the magmatic record. They are also characterized by higher rates of seismicity and stronger interplate coupling, larger amounts of crustal shortening and deformation, the creation and reactivation of fault systems and migration of deformation inland resulting in backarc uplifts (e.g., Barazanghi & Isacks, 1976; Gutscher et al., 2000; Espurt et al., 2008; Ramos & Folguera, 2009; Richardson et al., 2013; Martínez et al., 2016).

A commonly proposed mechanism to explain flat slab subduction is an increase in positive buoyancy due to altered lithosphere resulting from the subduction of ridges, seamounts, and oceanic plateaus (e.g., Gutscher, 2002; Martinod et al., 2010). For example, the Juan Fernandez ridge (JFR) is spatially correlated with the Pampean flat slab (e.g., Ramos et al., 2002; Yañez et al., 2001; 2002 Hu et al., 2016) and the Inca Plateau and the Nazca Ridge collocate with the Peruvian flat slab (e.g., Gutscher et al., 1999; Skinner & Clayton, 2013; Hu et al., 2016). However, there seem to be no signs of similar features in the vicinity of the Mexican flat slab and there is normal subduction along the Andean margin in the regions of the Carnegie and Iquique ridges (Espurt et al., 2008; Rosenbaum & Mo 2011; Skinner & Clayton, 2013; Manea et al., 2017). One could argue that, in the case of the Mexican flat slab, buoyant bodies were present in the past but have now been fully subducted so that there is no longer any evidence of their existence. In the case of the Carnegie and Iquique ridges, which started subducting ~1 Ma (Lonsdale & Klitgord, 1978) and ~2 Ma (Rosenbaum et al., 2005), respectively, it has been postulated that there is a delay between the subduction of buoyant features and the onset of flat slab subduction (Espurt et al., 2008; Hu et al., 2016). Therefore, more time may be needed for the slab to flatten in these regions.
In any event, studies have shown that the sizes of these buoyant features are not sufficient to initiate slab flattening to the extent observed but may contribute to the positive buoyancy needed for the slab to remain flat (e.g., van Hunen et al., 2002; Martinod et al., 2005; Gerya et al., 2009; Antonijevic et al., 2015).

Since altered lithosphere alone cannot initiate slab flattening, other contributing mechanisms have been proposed. One class of mechanisms invokes increased buoyancy due to [a] lithospheric hydration, with bathymetric irregularities resulting in greater hydration (e.g., Kopp et al., 2004; Porter et al., 2012), [b] a delay in the basalt to (denser) eclogite transition, [c] younger slabs being less dense and less resistant to bending (e.g. Gutscher et al., 2000; van Hunen et al., 2002; Hu et al., 2016), and [d] slab breakoff where the dense tip of the normal subducting slab is removed, leaving behind less dense lithosphere (Haschke et al., 2002, Haschke et al., 2006). A second class appeals to suction forces in the mantle wedge (van Hunen et al., 2004; Ma & Clayton, 2015) where these forces are increased by the presence of a craton (Manea et al., 2012; Taramon et al., 2015). A third class involves trench rollback and overthrusting of the subducting plate (e.g., van Hunen et al., 2002; 2004; Manea et al., 2012; Martinod et al., 2013), and a fourth to slab folding where the descending slab folds rather than penetrates through the 660 km discontinuity (Gibert et al., 2012; Cerpa et al., 2014). At present, the current consensus is that no one mechanism dominates the development of flat slab subduction zones, and perhaps some combination of mechanisms operates in each specific case (e.g., Manea et al. 2017).

1.2.2 Lithospheric Root Formation and Removal

We define a lithospheric root as any anomalously high-density body at the base of the lithosphere that evolved in situ as a result of local tectonic processes. Lithospheric root formation can occur through magmatic differentiation where partial melting at the base of the lithosphere
leads to differentiation of basalt with felsic melt rising to form continental crust and leaving behind a garnet pyroxenitic residue that acts as the root (e.g., Ducea, 2002; Lee & Anderson, 2015; Currie et al., 2015). In a flat slab subduction zone, this process is less likely to occur due to the cooler temperatures above the slab. More plausible processes for root formation in the flat slab environment include (1) a metamorphic phase transition of the lower crust to eclogite as a result of crustal shortening and increased pressure (e.g., Kay & Kay, 1993), and (2) thickening of continental lithosphere until it reaches a critical thickness of material that is colder and denser than that in the surrounding asthenosphere (e.g., Houseman et al., 1981).

In a subduction zone, material can be removed from the base of the continental lithosphere through ablation, in which subduction-induced mantle flow erodes and carries material to the subducting slab which subsequently drags it down into the mantle (e.g., Currie et al., 2015). In the case of dense material at the base of the lithosphere, two additional mechanisms can result in removal; (1) delamination, as proposed by Bird (1979) and (2) a Rayleigh-Taylor instability drip as proposed by Houseman et al. (1981). Delamination is a decoupling mechanism where the mantle part of the lithosphere “peels” away from the crustal part and sinks into the asthenosphere. The Rayleigh-Taylor instability drip involves the downwelling of the lithosphere due to the gravitational instability caused by a dense fluid overlying a less dense fluid. One imagines a “drip” as a blob of cold, dense lithosphere driven by gravity into the asthenosphere. Unlike delamination, the Rayleigh-Taylor instability may remove only a fraction of the lithosphere (e.g., Houseman & Molnar, 1997; Molnar et al., 1998; Molnar & Houseman, 2004). In a normal subduction zone, the Rayleigh-Taylor instability drip can be removed by shear entrainment as the root is pushed into the mantle wedge corner by subduction-induced mantle flow (Currie et al., 2015). However, in a
flat slab subduction zone there is little to no asthenospheric flow that would remove the root in this manner. In this case it would simply continue to thicken.

1.2.3 The Pampean Flat Slab Region

The Pampean flat slab is located beneath the South American plate between 28°S and 33°S (Figure 1.7) where the ~40 Ma Nazca plate converges with the South American plate at an oblique angle and a convergence rate of ~7 cm/yr (Manea et al., 2017). In this region, the Nazca plate subducts at a 30° angle to a depth of ~100 km, beyond which it extends more or less horizontally for ~300 km (from Chile to Argentina) before resuming its steep descent into the mantle (e.g., Cahill & Isacks, 1992; Anderson et al., 2007; Hayes et al., 2012; Manea et al., 2017; Linkimer et al., 2020). Bello-González et al. (2018) suggest that in addition to the Juan Fernandez ridge, the younger and hence more buoyant Copiapo and Taltal Ridges contributed to the flattening of the slab. The flattening of the slab began in the Miocene ~5 Ma (Manea et al., 2017) and is associated with an eastward migration and eventual cessation of volcanic activity in the area (Kay & Mpodozis, 2002; Ramos et al., 2002).

The flat slab also appears to be responsible for the eastward migration of deformation inland to the Precordillera and the Sierra Pampeanas (e.g., Jordan et al., 1983; Ramos et al. 1996; Ramos et al., 2002; Anderson et al., 2007). The thick-skinned Sierra Pampeanas is characterized by basement-cored uplifts of mafic-ultramafic metamorphic rocks of the Grenvillian Cuyania Terrane to the west (the Sierra Pie de Palo), and toward the east, Neoproterozoic – Early Paleozoic felsic metamorphic rocks of the Pampia Terrane, with sedimentary basins located between the uplifts (e.g. Ramos et al., 2002; Vujovich et al., 2004; Alvarado et al., 2007; 2009; Pfiffner, 2017; Linkimer et al., 2020). The Precordillera is a thin-skinned fold and thrust belt composed of Paleozoic sedimentary rocks which is underlain by the Cuyania Terrane (e.g., Ramos et al., 1996;
Ramos et al., 2002; Levina et al., 2014; Pfiffner, 2017; Linkimer et al., 2020). West of the Precordillera is the Iglesia basin (North) and Calingasta basin (South) followed by the high Andes (The Principal Cordillera and The Frontal Cordillera). The Principal Cordillera is composed of Mesozoic to Cenozoic sedimentary and volcanic rocks and the Frontal Cordillera is composed of Paleozoic to Mesozoic volcanic rocks and the western extent of the Cuyania Terrane, along with the Chilienia Terrane which consists of metamorphic rocks of Grenville age. Both are characterized by thick-skinned and thin-skinned thrust belts (e.g., Ramos et al., 1996; Ramos et al., 2002; Martinez et al., 2015; Pfiffner, 2017; Capaldi et al., 2020). Further east of the Sierra Pampeanas is the Rio de la Plata Craton which is composed of Precambrian to Early Paleozoic metamorphic rocks (Pfiffner, 2017).

The crust above the flat slab decreases in thickness from west to east. It can be as thick as ~70 km beneath the high Andes, ~50 km beneath the western Sierras Pampeanas, and thins to ~35 km beneath the eastern Sierra Pampeanas (Gilbert et al., 2006; Gans et al., 2011; Porter et al., 2012; Ammirati et al., 2015; Linkimer et al., 2020). The root of the Rio de la Plata craton has also been suggested to reach deep enough below the surface to have inhibited the eastern advancement of the flat slab (Booker et al., 2004). This evidence of overthickened crust in conjunction with evidence of eclogitization of the lower crust in the Cuyania Terrane (Alvarado et al., 2007; 2009; Marot et al., 2014; Ammirati et al., 2015; Pfiffner, 2017, Linkinmer et al., 2020) provide support for lithospheric root formation in the region of the flat slab. Several studies also support the hypothesis of episodic lithospheric removal north of the Pampean flat slab (e.g., Kay & Kay, 1993; Beck & Zandt, 2002; Bianchi et al., 2013; Ducea et al., 2013; Wang et al., 2015; Beck et al., 2015) which multiple investigators have attributed to Rayleigh-Taylor instability dripping (e.g., DeCelles et al., 2015; Schoenbohm et al., 2015; Murray et al., 2015). These inferences corroborate the
plausibility of the hypothesis of a dense lithospheric root in the region of the high Andes thickening onto the Pampean flat slab and causing a deflection in the slab.

Figure 1.7: Map showing area of study with geologic provinces (green lines) and terranes (yellow dashed lines) based on Marot et al. (2014) and Linkimer et al. (2020). Volcanoes (red triangles) and convergent plate boundary (yellow line with triangles) between the Nazca Plate and South American plate are from Gómez et al. (2019). Slab Contours (black lines) are from Hayes (2018). SP - Sierra Pie de Palo and SF - Sierra de Valle Fértil. Blue box locates the region shown in Figure 1.8. Elevation is indicated in the palette at the bottom of the figure.

1.3 Data

The present investigation of the Pampean flat slab region involves a joint inversion of P and S arrival times from local earthquakes and phase delays from surface wave dispersion measurements from ambient noise and earthquakes to generate a 3D image of elastic wavespeeds. Arrival time data should provide key evidence as the discovery of a substantial travel time anomaly provided the initial motivation for this dissertation. Hypocenter locations can give additional
insights into brittle failure in the region. Surface wave dispersion measurements are complementary to the arrival time data and help resolve the shallow structure of the subsurface. We originally planned to incorporate teleseismic arrival times in this study, but a preliminary attempt at building a dataset based on waveform correlation failed due to the complexity of the P waveforms we had available. Our main concern was the potential for contamination by cycle skips that would degrade our images; hence we decided to focus on the more robust local earthquake datasets.

Most seismological studies done in this region have been carried out in either Chile or Argentina, leaving the target area under the high Andes under-sampled. Our plan has been to exploit existing recordings from prior seismic networks to create better coverage of the area as well as obtain a larger volume of data than previous studies. Much of the data we analyze were available from archives of both permanent networks and temporary IRIS Portable Array Seismic Studies of the Continental Lithosphere (PASSCAL) deployments in both Chile and Argentina (Figure 1.8) over the past several decades. The permanent networks include the Chilean National Seismic Network (CSN) that has been in operation since 1991, and the Instituto Nacional de Prevencion Sismica (INPRES) that has been in operation in Argentina since 1972. Continuous waveform data for CSN has been archived since 2010; prior to that time only manually generated catalogues with corresponding windowed data was archived for 33 stations. The data were made available to us from our collaborators at the University of Chile, Santiago. Continuous waveform data recorded by 21 stations from INPRES for the time period from September 2015 to June 2017 were made available to us, from our collaborators at the University of San Juan, Argentina. The shorter term PASSCAL deployments include continuous waveform data from four deployments: the Chile ILLapel Aftershock eXperiment (CHILLAX) network that operated from November
2015 to November 2016 with 20 broadband stations and 18 short period sensors, the CHile-ARgentina Geophysical Experiment (CHARGE) network that operated from December 2000 to May 2002 with 23 broadband stations, the Eastern Sierra Pampeanas (ESP) network which operated from August 2008 to August 2010 with 12 broadband stations (Gilbert, 2008), and the Sierras Pampeanas Experiment using a Multicomponent BRoadband Array (SIEMBRA) network which operated from December 2007 to November 2009 with 44 broadband stations (Beck & Zandt, 2007). In addition to the PASSCAL data, our Chilean colleagues provided us continuous recordings from the CHile ARgentina Seismological Measurement Experiment (CHARSME) network that operated from November 2002 to February 2003 with 29 broadband stations. Finally, we obtained a manually picked catalogue for the Ovalle 1999 (OVA99) network that was in operation from November 1999 until January 2000 with 31 short period sensors. We also reviewed data from an analogue telemetered network that operated in the Pampeanas in the early 1980’s (PANDA) but determined it would not be useful for this study.

Figure 1.8: Map showing the locations of the seismic stations used in this study, color coded by their associated network. Elevation is indicated in the palette at the bottom of the figure.
1.4 Organization of the Dissertation

The research reviewed in this dissertation involves the application of passive seismic source techniques to an augmented dataset generated from existing seismic data that has been underutilized and often confined to either side of the border in Chile or Argentina. These techniques include ambient noise tomography to obtain dispersion measurements (Chapter 2), the automatic generation of a catalogue of P and S arrival times and hypocenter locations (Chapter 3) and joint inversion tomography of the arrival time data and dispersion measurements (Chapter 4). By imaging the geometry of the Pampean flat slab and the structure of the surrounding region, our objective is to contribute to the current understanding of the dynamics of the flat slab environment including but not limited to the possible constraints on root formation and removal in flat slab environments, and the ongoing debate concerning the driving mechanisms for flat slab subduction. Thus, this study will not only help in understanding processes that are presently occurring but also what might have happened in the past, and what could eventually happen in the future. For example, flat slab subduction has been suggested as a mechanism for the formation of the Laramide orogeny where the Andes has been used as a present-day analogue (Jordan & Allmendinger 1986; English & Johnston, 2004). There are also possible developing flat slabs along the Andean margin in the regions of the Carnegie ridge, Iquique Ridge, Tehuantepec ridge and Guanacos fold and thrust belt as a result of proposed cyclicity of normal and flat slab subduction along the margin (Espurt et al. 2008; Ramos & Folguera 2009). As a whole, therefore, the results of this study should allow us to gain insights into the general mechanics that govern subduction zones.
2. SURFACE WAVE DISPERSION MEASUREMENTS FROM AMBIENT NOISE

2.1 Abstract

Ambient noise tomography is a useful technique for resolving the shallow structure of the Earth without the need of earthquakes or active seismic sources. We use ambient noise tomography to generate Rayleigh wave phase velocity maps for the SIEMBRA network in Argentina. Dispersion curves were created using two techniques (1) a narrow bandpass filtering approach and (2) a zero\textsuperscript{th} order Bessel function zero value approach. We found these techniques gave very similar results, although, unlike the narrow bandpass filtering approach, the Bessel function approach theoretically can yield more data as it is less restricted by the relationship between wavelength and interstation distance. Hence, we used the Bessel function measurements for our analysis. At the same time, we determined some sensitivity to wavelength and interstation distance in several instances, specifically at longer periods and for station pairs with a short interstation distance, which is reminiscent of the original wavelength/distance constraint. Consequently, we created a standardized dataset that discards outliers by using a dynamic weighting approach which considers both absolute and percentage time residuals. This standardized dataset was used to create phase velocity maps that correlate well with the regional geology. In particular, high velocity anomalies are associated with the Frontal Cordillera, the Precordillera and the Sierra de Pie de Palo and low velocity anomalies are associated with the Bermejo Basin, Tulum Valley and the Jocoli Basin.

2.2 Introduction

Seismic waves may be categorized as body waves and surface waves. Body waves consist of P and S waves which can interfere at a free surface and generate surface waves. There are two
basic types of surface waves, Love waves and Rayleigh waves. Love waves are transverse waves polarized in the horizontal plane and are created by the interference of SH waves. Rayleigh waves are polarized in the vertical plane with an elliptical motion in the direction of propagation and are a result of interference of P and SV waves. Surface waves are dispersive, meaning different frequencies propagate at different velocities. Love waves are intrinsically dispersive, while Rayleigh waves are dispersive in heterogeneous media. Typically, longer period Rayleigh waves penetrate deeper into the Earth and hence travel faster than shorter period waves. As the wave packet propagates, two velocities are observed: (1) the group velocity, which is the velocity of the wave packet and (2) the phase velocity, which is the velocity of the individual frequency waves in the wave packet. Because longer period waves generally are sensitive to deeper structures and shorter period waves are more sensitive to shallower structures, we can use the dispersive property of surface waves to recover velocity variations within the Earth.

Seismic waves from earthquakes or active sources have long been used to provide information about the properties of the Earth’s subsurface. Traditionally, ambient noise (vibrations from anthropogenic sources and natural sources such as ocean waves or wind) was suppressed to enhance coherent signals. However, several studies have shown that we can use ambient noise to extract the impulse response, or Green’s Function (GF), of a medium between two points on the Earth’s surface through cross-correlation (e.g., Aki, 1957; Claerbout, 1968; Lobkis & Weaver, 2001; Campillo & Paul, 2003; Derode et al., 2003; Snieder, 2004; Weaver & Lobkis, 2004; Wapenaar, 2004; Shapiro & Campillo, 2004; van-Manen et al., 2005; Roux et al., 2005; Sabra et al., 2005a,b; Shapiro et al., 2005; Larose et al., 2005; Wapenaar & Fokkema, 2006; Lin et al., 2008; Moschetti et al., 2010; Ritzwoller et al., 2011; Lin et al., 2013; Wang et al., 2017). Extracting the GF is useful because, once it is known, the response (output) of a linear, time-invariant system
to any input can be calculated. Under the assumption of a diffuse wavefield or equipartitioning, the cross-correlation of ambient noise recorded at two receivers retrieves the GF between the receivers by essentially turning one of the receivers into a virtual source where surface waves are “generated” and then recorded at the other receiver (Wapenaar et al., 2010; Snieder & Wapenaar, 2010; Schuster, 2016). This procedure is referred to as Seismic Interferometry (see Appendix A) and is the basis of Ambient Noise Tomography (ANT). We typically extract surface waves from ambient noise as they are of larger amplitude and duration (e.g., Shapiro et al., 2005) and use these to measure dispersion from which we obtain group and phase velocities.

Here, we focus on recovering the phase velocity from Rayleigh waves and we can then use that information to recover shear wave velocity, which is generally about 5% higher than the phase velocity (e.g., Ammon et al., 2020). We choose to analyze phase velocity measurements over group velocity measurements because the former allow us to obtain constraints at greater depth (their sensitivity kernels usually extend to greater depths), they result in smaller uncertainties, and they can be used effectively at smaller interstation distances (e.g., Lin et al., 2008; Crosbie et al., 2019). We extract the GF of the fundamental mode of Rayleigh waves rather than Love waves or higher modes because the measurements are generally more robust.

2.3 Technique

2.3.1 Cross-correlation

For the most part, our analysis follows the procedure outlined in Bensen et al. (2007) to extract the Empirical Green’s Function (EGF) of Rayleigh Waves from the cross-correlation of ambient noise.

In a preprocessing stage, any gaps in the continuously recorded seismic signal are identified and, if not more than a few samples, filled by interpolation. The instrument response is removed,
as are the mean and any linear trend. High sample rate data is decimated to 1 Hz and then culled into individual day-length records. We then apply a trapezoidal bandpass filter for periods between 2s and 150s. To mitigate the effects of large amplitude and coherent signals such as earthquakes, the resulting time series is normalized using time-domain, running absolute mean normalization. Finally, we whiten the spectrum to enhance the noise.

The vertical components of the preprocessed day volumes are then cross-correlated in the frequency domain for each station pair and stacked (summed) in the time domain. We note that as the period of time used in correlation and stacking increases, the signal to noise (SNR) ratio increases and the observational uncertainty decreases.

The main contribution to ambient noise is often from primary and secondary oceanic microseisms which themselves are due to the interaction of ocean gravity waves and the seafloor (Yang & Ritzwoller, 2008). Hence, the proximity of stations to the source of the ambient noise (oceans) can result in asymmetric signals (Figure 2.2B) which in turn could cause a biased result. While some studies suggest that the results from ambient noise tomography in these cases are still useful and relate closely to what is derived from earthquake signals and to the geology of a region, we can account for the asymmetry in the signals by processing the “symmetric component”, i.e., the average of the causal and acausal signals (e.g., Lin et al., 2007; Zheng et al., 2008; Yang & Ritzwoller, 2008; Li et al., 2010; Nicolson et al., 2012; Buffoni et al., 2018). This averaging also tends to enhance the SNR. We can also apply a correction for azimuthal bias in the noise distribution to phase velocity estimates as described in Yao & van der Hilst (2009). However, a study by Roecker et al. (2017) found little difference in the phase velocity maps generated with and without this correction, which they attributed to the smoothing process employed in the creation of the phase velocity maps. We employ the same process here.
2.3.2 Phase Velocity Dispersion Curves

The first step in recovering velocity information from surface waves is the generation of dispersion curves. We obtain dispersion curves by employing two methods: (1) the narrow bandpass filtering approach of Yao et al. (2006) and Yao & van der Hilst (2009) and (2) the zero\textsuperscript{th} order Bessel function zero-value approach of Ekstrom et al. (2009) and Ekstrom (2014). As discussed below, the motivation for trying these two methods is that they are radically different in their approaches, and each offers their own particular advantages.

2.3.2.1 A Summary of the Narrow Bandpass Filtering Approach

Following Dahlen & Tromp (1998), the GF for a far-field fundamental mode surface wave travelling between a source at A and a receiver at B at an angular frequency $\omega$ and time $t$ may be theoretically approximated as:

$$G_{AB}(\omega, t) = A \cos \left( k_{AB} \Delta_{AB} - \omega t + \frac{\pi}{4} \right)$$ (2.1)

where $A$ is the amplitude, $\Delta_{AB}$ is the distance between A and B along a great circle path, and $k_{AB}$ is the average wavenumber along the path. The phase travel time, $t$, occurs when:

$$k_{AB} \Delta_{AB} - \omega t + \frac{\pi}{4} = 0$$ (2.2)

Here

$$k_{AB} = \frac{\omega}{c_{AB}}$$ (2.3)

where $c_{AB}$ is the phase velocity. Also

$$\omega = \frac{2\pi}{T}$$ (2.4)

where $T$ is the period. Hence

$$\frac{2\pi \Delta_{AB}}{T c_{AB}} - \frac{2\pi t}{T} + \frac{\pi}{4} = 0$$ (2.5)

and
\[
c_{AB} = \frac{\Delta_{AB}}{t - \frac{T}{8}}
\]  

(2.6)

The EGF is obtained from the derivative of the cross-correlation function (see Appendix A):

\[
\frac{dC_{AB}(\omega, t)}{dt} = -G_{AB}(\omega, t) + G_{BA}(\omega, -t)
\]  

(2.7)

Integrating the theoretical GF with respect to time gives the cross-correlation function as:

\[
C_{AB}(\omega, t) = H \sin \left( k_{AB} \Delta_{AB} - \omega t + \frac{\pi}{4} \right)
\]  

(2.8)

where \( H = \frac{A}{\omega} \). The EGF is thus recovered from the cross-correlation function by applying a \( + \frac{\pi}{2} \) phase shift to \( C_{AB} \): 

\[
G_{AB}(\omega, t) = A \sin \left( k_{AB} \Delta_{AB} - \omega t + \frac{\pi}{4} + \frac{\pi}{2} \right) = A \cos \left( k_{AB} \Delta_{AB} - \omega t + \frac{\pi}{4} \right)
\]  

(2.9)

Following Yao et al. (2006) and Yao & van der Hilst (2009), the resulting EGFs are passed through a series of narrow bandpass filters with central periods at 1s intervals and a passband width of \( \pm 0.2s \). A transit time vs central period plot is created which is then transformed into a velocity vs period plot and smoothed using spline interpolation. The phase velocity dispersion curve is then determined by following the maximum energy along the most physically reasonable trace. However, the potential ambiguity in determining what is “reasonable” can result in cycle skips. To mitigate this ambiguity, we guide the picking by requiring that it adhere to a polynomial “pilot function” generated by manually reviewing all the original dispersion curves for which there appear to be no cycle skips. Dispersion curves are then generated by choosing the cycles that are closest to the pilot function.

Since we are using the far-field approximation to surface wave generation, we also apply a criterion which states that the interstation distance must be at least 3 wavelengths to be considered reliable:
\[ c_{AB} T = \lambda \leq \frac{\Delta_{AB}}{3} \]  

(2.10)

The far-field approximation requires that there is enough distance between the source and receiver for surface waves to form through the interference of P and S waves. Close to the source, there can be inhomogeneous waves that are not surface waves which, if interpreted as surface waves, will result in errors in finding the dispersion curves.

2.3.2.2 A Summary of the Zero\(^{th}\) Order Bessel Function Zero-Value Approach

This Bessel function approach is based on a study by Aki (1957) which postulates that the real part of the Fourier Transform (FT) of the cross-correlation function \( C_{AB}(t) \) can be described by the zero\(^{th}\) order Bessel function, \( J_0 \) for a frequency \( \omega_0 \):

\[
\text{Re} \left\{ \text{FT} \left( C_{AB}(t) \right) \right\} = J_0 \left( \frac{\omega_0 \Delta_{AB}}{c_0} \right)
\]

(2.11)

where \( c_0 \) is the phase velocity at \( \omega_0 \). Following Ekstrom et al. (2009) and Ekstrom (2014), we can use the zero crossings, i.e., values of the argument where \( J_0(z) = 0 \), to approximate a solution as they are independent of amplitude scaling and their values may be robustly determined. In this case, the n\(^{th}\) zero crossing \( z_n \) is:

\[
z_n = \frac{\omega_n \Delta_{AB}}{c_n}
\]

(2.12)

so

\[
c_n = \frac{\omega_n \Delta_{AB}}{z_n}
\]

(2.13)

where \( \omega_n \) is the corresponding crossing frequency, and \( c_n \) is the phase velocity at \( \omega_n \). There are instances where the zero crossing may be ambiguous, leading to it being linked to the wrong zero. To compensate for this limitation, we use the equation:
\[ c_m = \frac{\omega_n \Delta_{AB}}{z_{n+2m}} \]  

(2.14)

where \( m = 0, \pm 1, \pm 2, \ldots \) representing zeros to the left and right of the original pick. Hence, multiple zero values of the Bessel function are assigned to the same zero crossing and we use the same pilot function method described above to choose the appropriate dispersion curve. One advantage of the Bessel function approach is that theoretically it does not depend on a far-field approximation and so it is potentially less sensitive to the 3-wavelength criterion, allowing for measurements at shorter interstation distances.

### 2.3.3 Phase Velocity Maps

To estimate the spatial distribution on phase velocities within a network, we use the interstation dispersion curves to generate two-dimensional phase velocity maps for each period. Since we have multiple station pairs with rays crossing each other, we determine this spatial distribution through a least-squares solution to an inverse problem (see Appendix B). In this inversion, the calculated phase delay times, \( T_{\text{cal}} \), are related to observed phase delay times, \( T_{\text{obs}} \), by a first order Taylor expansion:

\[ T_{\text{obs}} - T_{\text{cal}} = \sum \frac{\partial T_{\text{cal}}}{\partial s} \Delta s \]  

(2.15)

where \( s \) is the phase slowness (the reciprocal of the phase velocity), \( \Delta s \) represents the slowness perturbations to the model, and the sum is taken over all points in the model.

For each period, we start by estimating a constant velocity model based on the average of the observed phase delays. The phase delay time \( T \), which is the time taken for the wave to travel between the station pairs along the great circle path, \( l \), is:

\[ T = \int_l s \, dl \]  

(2.16)
We discretize this equation to calculate the delay time in the model by first finding the midpoint of a section of the path within 4 grid points and using bilinear interpolation to find the slowness. We define the fractional distance from a point \((x_0, y_0)\) to the mid point \((x_{mid}, y_{mid})\) as:

\[
\begin{align*}
    f_x &= \frac{x_{mid} - x_1}{h} \\
    f_y &= \frac{y_{mid} - y_1}{h}
\end{align*}
\]

where \(h\) is the grid spacing. The interpolation coefficients are:

\[
\begin{align*}
    c_1 &= (1 - f_x)(1 - f_y) \\
    c_2 &= f_x(1 - f_y) \\
    c_3 &= (1 - f_x)f_y \\
    c_4 &= f_x f_y f_z
\end{align*}
\]

So

\[
s_{mid} = s_1 c_1 + s_2 c_2 + s_3 c_3 + s_4 c_4
\]

which can be written as

\[
s_{mid} = \sum_{i=1}^{4} s_i c_i
\]

The delay time between the two stations can then be calculated by multiplying the slowness of each path section by their corresponding lengths, \(l_i\), and summing the products. Note that the partial derivatives are calculated by:

\[
\frac{\partial T}{\partial s_i} = c_i l_i
\]

The map is perturbed for several iterations until the observed and calculated times are sufficiently well matched. At each iteration the map is smoothed by passing a moving average window over the perturbations. Perturbations are calculated using the LSQR algorithm (Paige &
Saunders 1982a, b), which is an iterative approach to find an approximate solution for the damped least squares problem:

\[
\Delta m = (G^T C_d^{-1} G + C_m^{-1})^{-1} G^T C_d^{-1} \Delta d
\]  

(2.26)

where \(\Delta m\) is the vector of perturbations, \(G\) is the sensitivity matrix (the partial derivatives), \(C_d\) is the data covariance matrix, \(C_m\) is the model covariance matrix, and \(\Delta d\) is the vector of residuals. Instead of solving the above equation directly, LSQR (see Appendix B) uses Lanczos Bidiagonalization and QR factorization to solve:

\[
\begin{bmatrix}
\tilde{G} \\
\lambda I
\end{bmatrix} \Delta m = \begin{bmatrix}
\Delta \tilde{d} \\
0
\end{bmatrix}
\]  

(2.27)

where

\[
\tilde{G} = C_d^{-\frac{1}{2}} G
\]

(2.28)

\[
\Delta \tilde{d} = C_d^{-\frac{1}{2}} \Delta d
\]

(2.29)

\(\lambda\) is a damper that prevents overstepping, and \(C_d^{-\frac{1}{2}} = w I\), where \(w\) is the weight given by the inverse of the variance. Dispersion measurements (predictions) calculated from the resulting maps can then be used as the input in surface wave tomography.

2.4 SIEMBRA Network Results and Discussion

We performed ambient noise tomography on the SIEMBRA network using 932 contemporaneous station pairs. Cross-correlations were calculated and stacked for a total of 713 days from December 14th 2007 to November 25th 2009 (Figure 2.1).

As a quality check of the data and process, we tested the temporal repeatability of the cross-correlations (e.g., Bensen et al., 2007). We created 3-month cross-correlation stacks for the Argentinean Spring, Summer, Autumn and Winter and a stack of 12 months for the corresponding year to observe how the signal changed over these time periods (Figure 2.2B). We concluded that
there was no significant time dependent variation in the results, so we proceeded to use them to generate dispersion curves.

Dispersion curves were first generated using the Bessel function approach. After experimenting with 3rd, 4th, and 5th order polynomials we found that the following 4th order polynomial (Figure 2.3A) gave the best results (i.e., there were only a few instances of cycle jumps at higher frequencies, which would be expected because higher frequency measurements are harder to pick):

\[
c(T) = 3.884691 - 25.3454647T + 312.33197T^2 - 1755.36804T^3 + 3398.61719T^4
\]  

(2.30)

We tested the temporal repeatability of these measurements using the 3-month stacks and the 12-month stack to create dispersion curves. As we found with the cross-correlations, these dispersion curves showed very good correspondence (Figure 2.2C).

Figure 2.1: Stacked cross-correlations (yellow) from the SIEMBRA network, in order of increasing interstation distance in degrees. Red line denotes zero time delay. As the distance between the stations increases, the time taken for the signal to travel between the virtual sources receivers increases.
Figure 2.2: Results for the station pair ABRA-CASP for temporal repeatability tests. (A) Map view of the locations of the stations. (B) Stacked cross-correlations for the seasons of Argentina and corresponding year. Each cross-correlation shows an asymmetric signal with more energy on the acausal side of the response compared to the causal. (C) Dispersion curves for the seasons and corresponding year using the Bessel function approach. All the curves follow the same trend with little to no difference seen between ~0.1Hz and 0.23Hz.
Figure 2.3: Comparison of dispersion curves (phase velocity vs frequency or period) for the station pair ABRA-CASP (Figure 2.2A) from the Bessel function and Bandpass filtering approaches using the 4th order pilot polynomial. (A) The Bessel function approach. Crosses represent other possible dispersion curves as a result of multiple zero values from the zeroth order Bessel function to compensate for unclear zero crossings. (B) The narrow bandpass filtering approach. Blue and red areas represent peaks and troughs, respectively, of the contoured spectral amplitude. (C) Comparison of the results from the Bessel function approach (white squares) from (A) and narrow bandpass filtering approach (red circles) from (B). The two approaches generate essentially the same values with the Bessel function approach resulting in more observations.
We next employed the narrow bandpass filtering approach and again, found that a 4th order polynomial produced the best results (Figure 2.3B). A comparison of the dispersion curves generated from these two approaches gave very similar results, with the exception that, because of the relaxation of the 3-wavelength restriction, the Bessel function approach yielded more data points. For example, for the station pair ABRA and CASP (Figure 2.3C), the narrow band pass approach allowed estimates from 6s (0.167 Hz) to 27s (0.037 Hz) while the Bessel function approach allowed estimates from 4s (0.25 Hz) to 36s (0.028 Hz). For this reason, we favored the Bessel function approach and used those measurements in the creation of phase velocity maps for periods from 5s to 40s. In the generation of these maps, we applied a range of dampers (15, 20 and 40), and different moving window averages (7, 11, 15, 17 and 21 grid points) for several iterations. We found the optimal damping-smoothing pair to be 40 and 21, and obtained suitable results after 6 iterations.

As an extra measure of quality control, predicted dispersion curves generated from the phase velocity maps were compared to the observed dispersion curves. In some cases, we found significant discrepancies between the two curves. We tested to see if this could be attributed to an azimuthal feature but some station pairs along the same azimuth showed good correlation between the two curves while others did not. After further inspection, we noticed that the large discrepancies were caused by observations at longer periods for the station pairs that were close together, suggesting a residual sensitivity to wavelength and interstation distance. We therefore applied the 3-wavelength criteria to the Bessel function measurements by finding the maximum period cut off, $T_{max}$, for a station pair:

$$T_{max} = \frac{\Delta_{AB}}{c_0 \times 3}$$  \hspace{1cm} (2.31)
where $\Delta_{AB}$ is the interstation distance and $c_0$ is the characteristic velocity from the pilot function. Unfortunately, this restriction resulted in a lot of observations being discarded (from 24,189 to 8,776 observations) to the point where phase velocity maps for the longer periods were no longer realistic. For example, there were only 2 remaining observations at the longest period in the dataset (38s). Moreover, test inversions with this reduced dataset suggest that this criterion is too strict. Specifically, we found that phase velocity maps generated with the reduced dataset did not fit the data as well as the original maps (Figure 2.4), suggesting that many observations that fit the original model well had been discarded and that the inversion was not able to determine the global minima.

![Figure 2.4: A comparison of residuals from the phase velocity map for 28 s period for the dataset filtered with 3-wavelegth criteria (red) and the original dataset without the restriction of the criteria (blue).](image)

For this reason, we adopted a weighting approach that would focus on identifying outliers in the observations. We applied residual thresholds in terms of percentage time and absolute time and defined outliers as observations with travel time residuals greater 1 second and more than 5
percent of the total delay time. Dynamic application of this weighting scheme proved unstable, and so we created a standardized dataset of 23,272 observations by applying the outlier criteria after the first iteration. We also found that several periods (8, 18, 19, 20 and 21s) required increased damping to allow convergence (400 for 8s, 100 for 18, 19, 20 and 21s). The model at each period was obtained after 20 iterations, with a significant reduction in variance compared to that obtained from the initial dataset.

We tested the sensitivity of our results to choices of damping and smoothing by observing how different the delay times were from one map \((t_A)\) to another \((t_B)\). In general, we found that the percent changes:

\[
\%_{\text{change}} = \left| \frac{t_A - t_B}{t_A} \right| \times 100
\]

were all less than 2.5%, which is on the order of the uncertainty in the observation. In sum, the optimal value for smoothing was 21 grid points for all periods and for damping were 100 for periods of 18, 19, 20 and 21s, 400 for 8s, and 40 for the remaining periods.

![Figure 2.5: Sensitivity plot for the periods 5s to 40s at 5s intervals. As the period increases, peak sensitivity decreases, and the depth of maximum sensitivity increases. Sensitivity is calculated using the Gomberg & Masters (1988) locked-mode method.](image-url)
To determine which parts of the crust are controlling our observations, we created sensitivity plots for a range of periods using a representative 1D model (Figure 2.5). Sensitivity is defined as the change in phase velocity with respect to shear wave velocity at that depth. The shorter periods have a maximum sensitivity at approximately 5km. Their respective phase velocity maps (Figure 2.6 – 5s to 8s) show 4 prominent anomalies: (1) a high velocity anomaly to the west of 69.25°W along the latitudinal length of the map, (2) a high velocity anomaly between 68.5°W to 67.25°W and 31°S to 31.75°S, (3) a low velocity anomaly between 69°W to 67.5°W and 30°S to 31°S, and (4) a low velocity anomaly south of 31.5°S between 68.5°W and 67°W. Comparing these phase velocity maps to the regional geology of the area (Figure 2.7) we see that the high velocity anomalies correspond with the mountainous regions of the Frontal Cordillera and the Precordillera (anomaly 1) and Sierra de Pie de Palo (anomaly 2) and the low velocity anomalies correspond with two sedimentary basins, the Bermejo Basin (anomaly 3) and the Jocolí Basin (anomaly 4). The Frontal Cordillera is uplifted basement made up of volcanic rocks and is adjacent to the Precordillera which is made of sedimentary rocks. The Sierra Pie de Palo consists of crystalline metamorphic basement rock and the Bermejo and Jocolí basins are filled with Neogene & Quaternary sediment (e.g., Levina et al., 2014; Gallastegui et al., 2014; Pfiffner, 2017; Riesner et al., 2018). Thus, it is expected that the uplifted terranes will result in an increase in velocity and the more unconsolidated sedimentary basins will result in a decrease in velocity. The sedimentary nature of the Precordillera could explain the lesser percentage change in the western region of anomaly 1. The two low velocity anomalies show a connection which we interpret to be the Tulum basin. The Iglesia and Calingasta basins located between the Frontal Cordillera and the Precordillera do not show up as low velocity anomalies. The sediment in the Iglesia and Calingasta basins seems to be very shallow, that is, not reaching depths below mean sea level, whereas the
Bermejo and Jocoli basins extend to at least 5km below the surface (e.g., Suvires et al., 2012; Lossada et al., 2018). Thus, we postulate that because the sediment in the Iglesia and Calingasta basins are not as thick as in the Bermejo and Jocoli basins, they do not show up on the phase velocity maps.

At the 20s period there is a transition to slow velocity anomalies to the west and high velocity anomalies to the east, and, as the period increases, higher velocity anomalies appear between the slow anomalies in the west (Figure 2.6 – 25 s to 40 s). In this region, the deeper crust consists of the Cuyania formation (Figure 1.7). This is the same type of material as the Sierra de Pie de Palo so we would expect similar velocity anomalies throughout. However, there are large thrust faults and ramp detachment structures in the region (e.g., Siame, 2002; Pfiffner, 2017) that could be responsible for the complicated structure we see. There is also evidence of eclogized crust (e.g., Alvarado et al., 2007, Linkimer et al., 2020) in this area that could be responsible for the higher velocity anomalies. We also have to take into consideration that although we are sampling the deeper crustal structure at longer periods, we lose sensitivity with depth as well as the number of observations and spatial resolution decreases. For this reason, we will use this model as a starting point as we attempt to resolve deeper features with body waves in the joint inversion (Chapter 4).

Porter et al. (2012) also created phase velocity maps for this region using earthquake-sourced and ambient noise dispersion measurements. At their shorter periods they observed higher velocities for the Andes and Sierra de Pie de Palo and lower velocities for the Bermejo and Jocoli Basins in agreement with our results. They also observed an eastward increase in velocity from the periods of 20s to 30s which they attributed to the decrease in crustal thickness from west to east, observed by other studies (e.g. Gilbert et al., 2006), such that in the west the crust is being sampled and in the east the mantle is sampled. This interpretation could provide an explanation for
our results at deeper periods, that is, the low velocity anomaly in the west could be due to the presence of the thicker crust while to the east, the high velocity anomaly could be due to the mantle. Since Porter et al. (2012) used both ambient noise and earthquake-sourced dispersion measurements their results are more sensitive to deeper structures. The correspondence of our analysis to theirs gives us more confidence in our results.

Figure 2.6: Phase Velocity Maps as a function of period as shown in the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 5%, respectively. The black box shows the extent of the regional map in Figure 2.7. The numbers represent the anomalies described in the text.
Figure 2.7: Map showing geologic provinces based on Voldman et al. (2010). Blue represents uplifted units that correspond to high velocity anomalies (1 and 2 in Figure 2.6) and red represents basins that correspond to low velocity anomalies (3 and 4 in Figure 2.6).
3. AUTOMATICALLY GENERATING AND AUGMENTING SEISMIC CATALOGUES USING A REGRESSIVE ESTIMATION ALGORITHM

3.1 Abstract

Body wave arrival times depend on the properties of the medium through which a wave travels from a source to a receiver, making them a useful source of information about the interior of the Earth. Manual estimation of arrival times becomes intractable when large volumes of data are involved, as is the case for this study. Hence, we employed an automated regressive estimation algorithm (REST) to determine arrival times for P and S waves and generate catalogues of earthquake hypocenters in the Pampean flat slab region using archived data from multiple seismic networks. Using this approach, we were able to obtain catalogues with several orders of magnitude more arrival times and hypocenter locations than had been previously available for the region. Specifically, we were able to determine (1) 146,647 hypocenters with 3,459,029 P and 2,653,563 S arrivals for the contemporaneous SIEMBRA, ESP and CSN networks (2) 14,672 events with 140,663 P arrivals and 115,910 S arrivals for the CHARGE network; (3) 8,419 events with 108,771 P arrivals and 95,142 S arrivals for the contemporaneous CHARM and CSN networks and (4) 38,387 events with 508,465 P arrivals and 435,538 S arrivals for the contemporaneous CHILLAX and INPRES networks. An application of strict location quality criteria to these catalogues resulted in (1) 32,206 events with 1,081,399 P arrivals and 1,009,853 S arrivals for the SIEMBRA and ESP deployments, (2) 7,313 events with 76,185 P arrivals and 69,055 S arrivals for the CHARGE deployment; (3) 3,176 events with 48,658 P arrivals and 45,285 S arrivals for the CHARM and CSN deployments and (4) 25,323 events with 361,237 P arrivals and 323,106 S arrivals for the CHILLAX and INPRES deployments. Some of the results obtained from these new
catalogues corroborate those from previous studies using these networks, such as, a high level of seismicity in the subducted Nazca plate associated with the Juan Fernandez Ridge (JFR), the shallowing of the slab underneath Argentina and the presence of a double seismic zone where the slab initially descends steeply into the mantle. They also revealed some patterns in the seismicity that were not previously detected, in particular several parallel lineations trending NNE-SSW (20° azimuth) that intersect the subducted JFR, and an apparent coupling between the deeper mantle seismicity and the shallower crustal seismicity directly above it. These lineations are reflected in minor ridges found in the bathymetry of the Nazca plate. Based on this correlation, in conjunction with the relationship between the deep and shallow seismicity, we infer that subducted bathymetric highs are releasing volatiles, through dehydration or decarbonation, which are infiltrating and fracking the overlying crust.

### 3.2 Introduction

Seismic body waves traverse the interior of the Earth and are classified as either P waves or S waves. P (primary or pressure waves) waves are compressional waves with particle displacements in the direction of propagation, while S (shear or secondary waves) waves are transverse waves with particle displacements perpendicular to the direction of propagation. Material properties, specifically density and the bulk and shear moduli, determine the speed of these waves as they propagate through the Earth. Hence, we can use the arrival times of P and S waves at seismic stations to gain insights into the lithology of the subsurface. Historically, P and S wave arrival times were manually picked, but as our data collection methods improved the amount of data available has become overwhelming and the task of manual picking is in many cases impractical. Moreover, manual picking can introduce individual bias and uncertainties that are difficult to quantify. As a result, a number of algorithms have been proposed to take advantage
of present-day computing power to automatically pick P and S wave arrivals. Early examples include short- and long-term average ratio (e.g., Allen, 1978; Baer & Kradolfer, 1987), autoregressive modelling (e.g., Sleeman & van Eck, 1999), higher order statistics (e.g., Saragiotis et al., 2002), and shallow neural networks (e.g., Gentili & Michelini, 2006). However, there have been concerns with the computational time needed to run these algorithms and their accuracy with complex waveforms when compared to manual picking, so researchers have more recently shifted focus to deep neural networks for phase picking (e.g., Zhu & Beroza, 2019; Wang et al., 2019; Yu & Ma, 2021).

In this study we employ the Regressive Estimation (REST) algorithm written by S. Roecker. REST is an iterative approach that makes arrival time picks based on an autoregressive approach that uses the statistical properties of an evolving time series (Pisarenko et al., 1987; Kushnir et al., 1990), and locates earthquakes using Bayesian inverse theory (Tarantola & Valette, 1982a). REST works by being initially very inclusive to maximize the number of potential events, and sequentially removing false positives. REST has been successfully applied to datasets from Chile (Comte et al., 2019; 2023), New Zealand (Lanza et al., 2019; Yarce et al., 2023), Canada (Merrill et al., 2022) and Alaska (Littel et al., 2023). Here, we use REST to automatically generate catalogues of P and S wave arrival times and hypocenter locations using data from previously deployed seismic networks in the Pampean flat slab region; specifically CHARGE, CHARME, CHILLAX, ESP, SIEMBRA, CSN and INPRES. We also use catalogues from networks in either Chile or Argentina to search for arrivals in the other country to further enhance the dataset and improve sampling in our region of interest beneath the high Andes. Previous studies of the Pampean flat slab region have produced catalogues of events that are largely manually picked (e.g., Wagner et al., 2005; Anderson et al., 2007; Marot et al., 2014). More recently, Linkimer et al.
(2020) used short- and long-term average ratio to generate 69,691 detections and create a catalogue of 1,092 manually picked events for the SIEMBRA and ESP networks. We were able to produce a catalogue of more than 2 orders of magnitude more events for these two networks using REST.

### 3.3 Technique

The REST algorithm has two fundamental stages: a “detection” stage and an “onset estimation” stage. The purpose of the detection stage is to find any part of a time series, regardless of origin, that differs in some statistical way from the ambient background. The onset estimation stage then uses these detections as a starting point to decide if they could be related to a coherent point source (such as an earthquake or explosion) and, in the process, estimate the arrival times of P and S waves generated by that source.

Following Pisarenko et al. (1986) and Kushnir et al. (1990), detections in a time series are made by assuming that the noise and signal are zero mean Gaussian random processes. The spectral density of the noise is estimated over a long time interval while the spectral density of the signal remains unknown. Two hypotheses are formed for testing for signals within a data window; \( H_0 \), the hypothesis that there is purely noise and \( H_1 \), the hypothesis that there is noise and a signal. The likelihood ratio of \( H_0 \) and \( H_1 \) is used as a criterion for detection:

\[
L_t \left( \frac{x_t}{\theta_N}, \theta_S \right) = \frac{w \left( \frac{x_t}{\theta_S} \right)}{w \left( \frac{x_t}{\theta_N} \right)}
\]

(3.1)

where \( x_t \) is the windowed time series, \( \theta_N \) and \( \theta_S \) are the noise and signal spectrums respectively and \( w \left( \frac{x_t}{\theta_S} \right) \) and \( w \left( \frac{x_t}{\theta_N} \right) \) are their corresponding probability density functions. A detection is registered if the detection function above exceeds a user-specified threshold.
Once detections are made for each station in a network, they are sorted chronologically and classified as potential events based on the simple criterion that an earthquake will be recorded at several stations within a certain window (typically a few seconds) of each other depending on interstation distance. This simple criterion eliminates a significant percentage of detections but still may include many false positives. In this case the user can raise the threshold or require that some number of included detections have an exceptionally high threshold. Based on the first and last detections associated with a potential event, directories containing windows of seismograms are created for the “estimation” stage of analysis.

As described in Pisarenko et al. (1986) and Kushnir et al. (1990), onset estimations are made based on a maximum likelihood function for the arrival of the signal at some unknown time, $\tau$. For every $\tau$ within $(t_1, t_2)$, autoregressive models of the observations are calculated between $(t_1, \tau)$ and $(\tau, t_2)$ using the Levinson-Durbin procedure. The variances of the autoregressive model residuals, $\sigma_i^2$, are then used to calculate the likelihood function:

$$L(\tau) = [\tau \ln \sigma_1(\tau) - (N - \tau) \ln \sigma_2(\tau)]$$

where $N$ is the length of the time interval. Essentially, this function provides a quantitative estimate of how points earlier than the sampled point are different from the points after it. The onset is defined as the peak of this function, that is, where there is the greatest difference before and after the point in question.

A strength and weakness of this approach is that it makes no assumptions about the waveform morphology, i.e., it does not try to ascertain if the signal “looks seismic”. The advantage is a lack of potential bias – it does not need to know what the signal should look like. The disadvantage is that such indifference can lead to false arrivals. We mitigate this lack of background information by specifying some additional criteria based on simple characteristics of
a seismic waveform and the estimation function, specifically: (1) the shape of the onset function should be asymmetric (gentle ascent with a sharp descent), (2) the ratio of the amplitude before and after the estimate should be larger than some threshold, and (3) the ratio of the maximum/minimum values of the onset function should be greater than 1. The third criterion is related to the quality of the function itself, while the first two help ensure causality.

An essential part of the REST algorithm is the specification of a suitable window in which the onset is estimated, which in turn depends on the computation of a reasonable phase arrival time. The windowing procedures for these onset estimations used by REST are based for the most part on techniques used by Rawles & Thurber (2015).

Arrival time computation requires specifying a wavespeed model, source and recording locations (i.e., a hypocenter and station), and a means to calculate the travel time between the two. Station locations and wavespeeds are presumed to be known. This is typically not an issue for station locations, and as we only need to be sure that an arrival occurs within a window of several seconds duration, a simple 1D wavespeed model often is adequate. If REST is used in “augmentation” mode, then hypocenters are provided by an existing catalogue and are presumed to be correct. In the more general “raw” mode, hypocenters are determined as part of an iterative process. Both applications require the computation of a travel time from source to receiver. The algorithm we use for this purpose is the program SPHFD (S.Roecker), that can calculate travel times for P and S waves in an arbitrary 3D wavespeed model. SPHFD uses the finite difference solution to the spherical eikonal equation (Zhiwei et al., 2009; Zhang et al., 2012) based on techniques proposed by Vidale (1988,1990) and Hole and Zelt (1995). The spherical eikonal equation is:

\[
\left( \frac{\partial t}{\partial r} \right)^2 + \frac{1}{r^2} \left( \frac{\partial t}{\partial \theta} \right)^2 + \frac{1}{r^2 \sin^2 \phi} \left( \frac{\partial t}{\partial z} \right)^2 = s(r, \theta, \phi)^2
\]  

(3.3)
where $r$ is the radius from the center of the earth, $\theta$ is the colatitude, $\phi$ is the longitude, and $s$ is slowness (reciprocal of velocity). Solving the eikonal equation (derived in Appendix C) numerically is more accurate than ray tracing methods in regions where there are sharp changes in gradient (i.e., strong lateral heterogeneity) (Roecker et al., 2006). We use a spherical coordinate system rather than a cartesian system to more accurately represent the curvature of the Earth over large distances (e.g., Zhang et al., 2012).

The subsurface model is specified by assigning slownesses to a regularly spaced grid of nodes. The slowness between the nodes is calculated using trilinear interpolation of 8 neighboring grid nodes weighted by the fractional distances from the target point and summed. The fractional distance for a point at $(r, \theta, \phi)$ from the point $(r_1, \theta_1, \phi_1)$ where the other seven points are positive distances is:

$$f_3 = \frac{x - x_1}{r \sin(\theta) d\phi}$$

$$f_4 = \frac{y - y_1}{rd\theta}$$

$$f_7 = \frac{z - z_1}{h}$$

where $h$ is the radial grid spacing. The interpolated slowness is:

$$s = s_1 (1 - f_x)(1 - f_y)(1 - f_z) + s_2 f_x(1 - f_y)(1 - f_z) + s_3(1 - f_x) f_y (1 - f_z) + s_4(1 - f_x)(1 - f_y) f_z$$

$$+ s_5 f_x(1 - f_y)f_z + s_6(1 - f_x)f_y f_z + s_7 f_x f_y (1 - f_z) + s_8 f_x f_y f_z$$

or

$$s = \sum_{i=1}^{8} s_i c_i$$
where \( s_i \) is the slowness at grid point, \( i \) and \( c_i \) is the corresponding interpolation coefficient. Travel times to nodes in the immediate vicinity of the starting point are calculated assuming a straight ray path and integrating the slowness. The algorithm then expands to the other points using differencing schemes described in Zhang et al. (2012), where the choice of scheme depends on the points to be determined. If needed, ray paths can be calculated from the travel timetables using Fermat’s Principle of least time.

In “augmentation” mode, a window is specified after an arrival time is calculated and an onset estimated using the method described above. Typically, the vertical component is used for P arrivals and both horizontals are used simultaneously for the S arrival. The same is done in “raw” mode, but because the hypocenter is unknown (or even if the potential event is in fact an earthquake), additional processing needs to be done. Essentially, REST applies a criterion that earthquake related phases must, to a certain tolerance, appear to have emanated from the same location in space at the same time (i.e., from a hypocenter). It applies this criterion iteratively, gradually reducing thresholds for acceptable uncertainties in arrival times and locations. Estimations of P wave arrival times are first made on the vertical (Z) component, and if an acceptable hypocenter can be found, the process is repeated for S arrivals on the horizontal components. Typically, four iterations are performed: two each for P and S.

Hypocenter locations are estimated through a grid search procedure based on theory from Tarantola & Valette (1982a) in which a Gaussian distribution of uncertainties is presumed in a Bayesian approach. We define the a posteriori probability density function (PDF) function \( \sigma(X,Y,Z) \) as:

\[
\sigma(X,Y,Z) = K \rho(X,Y,Z) e^{-\frac{1}{2} \left[ \bar{\kappa}(X,Y,Z) \right]^T \left[ \bar{\kappa}(X,Y,Z) \right] + \bar{\kappa}(X,Y,Z) \}
\]

(3.9)
where $\rho(X, Y, Z)$ is the a priori density function (generally assumed to be a constant for $Z > 0$), $\tilde{\tau}_0$ are the demeaned observed arrival times (the difference between the observed arrival times and the weighted mean of those times), $\tilde{h}(X, Y, Z)$ is the demeaned calculated travel times (difference between the calculated travel times and their weighted mean), $P$ is the inverse data covariance matrix (usually a diagonal matrix under the assumption the data uncertainties are not correlated) with elements that are weights $p_i$ (inverse of the variance), and $K$ is the sum of the weights $p_i$ over all the phases for an event. The hypocenter is defined as the point $(X, Y, Z)$ where the PDF is a maximum. The location determined by grid search can be refined by a subgrid search where travel times are calculated by trilinear interpolation in the vicinity of the primary grid point. The algorithm also determines Marginal PDFs (MPDFs) for $\sigma(X, Y)$ and $\sigma(Z)$ by integration:

$$\sigma(X, Y) = \int \sigma(X, Y, Z) \, dZ$$  \hspace{1cm} (3.10)

$$\sigma(Z) = \iint \sigma(X, Y, Z) \, dXdY$$ \hspace{1cm} (3.11)

These MPDFs are used to estimate location uncertainty by calculating 66% and 95% of the area under the curve surrounding the maximum likelihood point.

Origin time is determined by the weighted residuals of the arrivals at all stations. At a station $j$, the relationship between the arrival time, $\tau_j$, the origin time, $OT$, and the travel time, $T_j$ is:

$$\tau_j - OT = T_j$$  \hspace{1cm} (3.12)

$$\sum \tau_j - NOT = \sum T_j$$  \hspace{1cm} (3.13)

$$OT = \frac{\sum T_j - \sum \tau_j}{N}$$  \hspace{1cm} (3.14)

where $N$ is the number of stations. When using weighted residuals, $w_j$ the equation becomes:
In addition to hypocenters, REST also makes estimates of magnitude and P wave polarity. The magnitudes $M$ of the earthquakes are determined by fitting the maximum amplitudes, $A_{\text{max}}$, and the S-P arrival time difference $\Delta t$ to a version of the Richter formula, e.g.:

$$M = \log A_{\text{max}} + 3 \log 8 \Delta t - 2.92$$

(3.16)

The polarity of the P wave arrival is determined by finding the zero crossings of the time series within a short window of the time series following the onset estimation. These polarities can then be used in another routine, such as the FOCMECH routine of Snoke (2003) to calculate focal mechanisms.

### 3.4 Results

Table 3.1: Summary of catalogues generated for various networks with the REST algorithm.

<table>
<thead>
<tr>
<th>Catalogues</th>
<th>Events</th>
<th>P Arrivals</th>
<th>S Arrivals</th>
</tr>
</thead>
<tbody>
<tr>
<td>SIEMBRA-ESP-CSN</td>
<td>146,647</td>
<td>3,459,029</td>
<td>2,653,563</td>
</tr>
<tr>
<td>CHARGE</td>
<td>14,672</td>
<td>140,663</td>
<td>115,910</td>
</tr>
<tr>
<td>CHARMSME-CSN</td>
<td>8,419</td>
<td>108,771</td>
<td>95,142</td>
</tr>
<tr>
<td>CHILLAX-INPRES</td>
<td>38,387</td>
<td>508,465</td>
<td>435,538</td>
</tr>
</tbody>
</table>

We generated catalogues by applying REST to detect and locate events for the seismic networks deployed in Chile and Argentina and then further augmented all but one of these catalogues (CHARGE) by searching for arrivals from contemporaneous networks. A 1D model based on the results of Marot et al. (2014) for this region was assigned to a grid with a spacing of 5km. Intragrid spacings of 10 were assigned for hypocenter location, meaning that subgrid searches were made at a 5m spacing. We obtained enhanced catalogues (Table 3.1) for: (1) the
SIEMBRA deployment and its overlap with the ESP deployment and CSN (2) the CHARGE deployment, (3) the CHARSME deployment and its overlap with CSN, and (4) the CHILLAX deployment and its overlap with INPRES. In this section we discuss the details of these new catalogues.

### 3.4.1 SIEMBRA-ESP-CSN

The SIEMBRA network is of particular importance to this investigation because it is located directly above the region where the “flat” slab shallows. Analysis of the SIEMBRA continuous data with REST resulted in 108,776 events, with 2,315,424 P-wave arrivals and 2,427,996 S-wave arrivals. We combined this dataset with arrivals from the ESP network during the time it overlapped with the SIEMBRA network. Merging the two catalogues resulted in 143,716 events. Many of these hypocenters will have large uncertainties or otherwise be poorly constrained, so we applied multiple criteria to remove the less reliable ones. These criteria included requiring a minimum number of 12 phases with at least 3 S-wave arrivals, a maximum residual standard deviation of 0.5s, and a maximum total location uncertainty of 10km. Application of these criteria resulted in 32,206 events with 1,081,399 P arrivals and 1,009,853 S arrivals. Interestingly, applying the same criteria to the SIEMBRA catalogue by itself resulted in about 3,000 more events, which suggests that the standard deviation of residuals increased for some events using ESP stations, probably because of unmodeled lateral heterogeneity between the two networks. As a further quality check on this catalogue, we compared subsets of locations from 2008 and 2009 and obtained similar results, suggesting that these locations are representative of ambient seismicity in the region and not contaminated by transients.

Plots of the well-constrained events from the SIEMBRA-ESP catalogue in cross-section (Figures 3.1 & 3.2) show that these events are concentrated in two zones; one within the upper 50
km, and hence within the South American crust, and a second at depths greater than 80 km, and hence in the mantle, most likely within the subducted Nazca plate. The deep events in the N-S cross-section (Figure 3.1) show a convex up slab geometry where the Pampean flat slab, located at the shallowest part of the zone, is bordered on each side by normal subduction. W-E cross-sections (Figure 3.2) show that the “flat” slab shallows by ~25km before it re-subducts into the mantle. Similar geometries were noted by previous investigators (e.g., Anderson et al., 2007; Alvarado et al., 2009; Linkimer et al., 2020).

**Figure 3.1:** N-S cross-sections of the well constrained hypocenters from the SIEMBRA-ESP catalogue. The longitude of the section is shown at the lower left corner of each panel. Events are included if they are within 10 km of a vertical plane passing though that longitude.

**Figure 3.2:** W-E cross-sections of the well constrained hypocenters from the SIEMBRA-ESP catalogue. The latitude of the section is shown at the lower left corner of each panel. Events are included if they are within 10 km of a vertical plane passing though that latitude.
Figure 3.3: Map views of the well constrained hypocenter locations for (A) shallow events (< 70 km depth) and (B) deeper events (> 70 km depth). Hypocenters are circles with colors corresponding to depth as indicated in the palette above each figure. Yellow triangles are the stations of the SIEMBRA-ESP network.

Figure 3.4: (A) Histogram of shallow events from 0 to 70 km depth (B) Histogram of shallow events in the inverted U shape.
Map views of this seismicity (Figure 3.3), show that the deeper zone (Figure 3.3B), in addition to the high concentration of events between 31°-32°S that trends ~70°, contains four distinct and parallel NNE-SSW (~20°) linear trends that are separated by ~50 km. These linear trends coincide with similar concentrations seen in the shallow events between 31°-32°S (Figure 3.3A). In particular, there is a sharp decrease in seismicity in both the shallow and deep events both north of 31°S and to the east of ~67.5°W between the SIEMBRA and ESP networks. There is also a distinct inverted U-shaped feature in the deep events that is mirrored in the shallow events (Figure 3.4) but shifted slightly to the east.

Of the 12,172 focal mechanisms obtained from the joint catalogue (Figures 3.5 & 3.6), most (10,819) were from the deeper events. Of these, 2,686 of the mechanisms from deeper events and 156 mechanisms from the shallower events (Figures 3.5 & 3.6), were constrained by at least 18 impulsive polarities. The deeper events show predominantly normal/extensional faulting while the shallower events show mostly thrust faulting with E-W compression. These trends are similar to those from other focal mechanism studies done in the region (e.g., Anderson et al., 2007; Alvarado & Ramos 2011; Ammirati et al., 2015) although with many more examples.

![Figure 3.5](image)

Figure 3.5: Map view of the lower hemisphere focal mechanisms for the (A) shallow hypocenters and (B) deep hypocenters. Normal faulting is represented in red, thrust faulting in blue and strike-slip faulting in green.
Figure 3.6: W-E cross-section of back-hemisphere focal mechanisms for the (A) shallow hypocenters and (B) deep hypocenters. Normal faulting is represented in red, thrust faulting in blue and strike-slip faulting in green.

To extend coverage from Argentina to Chile and sample the region under the high Andes, we augmented the SIEMBRA-ESP catalogue with data from the CSN network. The CSN data for this time period consists of a manually picked catalogue of 3,069 events with corresponding windowed seismograms. Since we did not have continuous CSN recordings we could only search for arrivals in the SIEMBRA and ESP stations using the CSN catalogue and not vice versa. To do so, we created a catalogue with predicted arrival times at the SIEMBRA, ESP and CSN stations from the hypocenter locations in the CSN catalogue. REST then used these predicted arrival times to create windows where the P and S arrival times should be and searched for viable phases in these windows. This exercise resulted in 157,927 P arrivals and 105,962 S arrivals for the 3,069 events. The combined catalogue shows the same steeply dipping trend of the hypocenter locations where the flat slab is expected (Figure 3.7) as found with the CHILLAX locations discussed in Chapter 1 (Note that what appear to be lines of events at certain depths, most notably at 150 km,
are artifacts generated by gradients in the 1D model). A comparison of this catalogue with the SIEMBRA-ESP only catalogue showed the standard deviation of residuals was significantly larger when the CSN stations were added, which, as before, is most likely due to the inadequacy of the 1D model used in our application of REST. We should note that the hypocenters in this figure were not subjected to the same stringent criteria for location quality because we suspect that the relatively poor fit could contain valuable information about the media. This catalogue will be discussed in more detail in Chapter 4.

Figure 3.7: CSN hypocenters after relocation with data from the SIEMBRA, ESP and CSN stations. (A) Map view of the hypocenter locations (circles with colors corresponding to depth as indicated in the palette at the top of the figure) and stations (blue triangles) within the bounds of the map. (B) W-E cross-section of all the hypocenter locations in the dataset.
Figure 3.8: CHARGE hypocenter locations. (A) Map view of the hypocenter locations (circles with colors corresponding to depth as indicated in the palette at the top of the figure) and stations (blue triangles) within the bounds of the map. (B) W-E cross-section of all the hypocenter locations in the dataset.

3.4.2 CHARGE

Application of REST to data from the CHARGE network (Figure 1.8 in Chapter 1) resulted in a catalogue of 14,672 events with 140,663 P arrivals and 115,910 S arrivals. The CHARGE network did not overlap with any other network for which we could obtain data and so no further augmentation was possible. Because this dataset is considerably smaller than that from SIEMBRA-ESP, we applied a less restrictive criteria in filtering the hypocenters. Specifically, an accepted
event was required to have a minimum of 10 phases with at least 3 S arrivals, a maximum residual standard deviation of 0.8s, a maximum total uncertainty of 25km. Application of this criteria resulted in 7,313 events with 76,185 P arrivals and 69,055 S arrivals. Plots of these events (Figure 3.8) show trends similar to those seen in the SIEMBRA-ESP catalogue in the deep seismicity; specifically, the high concentration of earthquakes that trends ~70° between 31°-32°S and the 4 parallel NNE-SSW trending lineations, the concentration of earthquakes in the “flat” Nazca slab and the ~25km shallowing of the slab (Figure 3.8B). Additionally, the westernmost band of the linear trend can be seen extending further south (Figure 3.8A), mostly because of the larger aperture of this network.

3.4.3 CHARSMCE-CSN

Application of REST to data from the joint CHARSMCE and CSN networks allowed the creation of a catalogue of an additional 8,419 events 108,771 P arrivals and 95,142 S arrivals. We were able to detect events with the continuous CHARSMCE network and augment the resulting catalogue with manually picked arrivals from about 500 events in the CSN catalogue. This catalogue was then filtered with the same criteria applied to the CHARGE dataset, resulting in 3,176 events with 48,658 P arrivals and 45,285 S arrivals. Plots of the filtered locations (Figure 3.9) show the same ~70° trend in the earthquakes seen in the SIEMBRA-ESP and CHARGE catalogues (Figure 3.9A) and clearly show a bend in the slab (Figure 3.9B). There is also some indication of a double seismic zone between 72°W and 71°W. A similar feature was reported by Marot et al. (2013) using a manually picked catalogue of events from these networks (CHARSMCE and CSN) and the OVA99 network (Figures 3.9B & 3.10). Note that while we have access to the OVA99 catalogue we could not do any additional processing as the waveform data were unavailable.
Figure 3.9: CHARM-CSN hypocenter locations. (A) Map view of the hypocenter locations (circles with colors corresponding to depth as indicated in the palette at the top of the figure) and stations (blue triangles). (B) W-E cross-section of all the hypocenter locations.

Figure 3.10: W-E cross-section of the CHARM-CSN hypocenter locations between 31.2°S and 32.2°S where Marot et al. (2013) observed the double seismic zone clearly.
Figure 3.11: CHILLAX-INPRES hypocenter locations. (A) Map view of the hypocenter locations (circles with colors corresponding to depth as indicated in the palette at the top of the figure) and stations (blue triangles) within the bounds of the map (top). (B) W-E cross-section of all the hypocenter locations.

Figure 3.12: W-E cross-section of the CHILLAX-INPRES hypocenter locations between 31.2°S and 32.2°S where Marot et al. (2013) observed the double seismic zone clearly.
3.4.4 CHILLAX-INPRES

The CHILLAX network was previously processed with REST (Comte et al. 2019; this was the initial application of REST and the results of that study provided the initial motivation for this dissertation). We used their catalogue of 38,388 events to search for arrivals recorded by the INPRES network and found 11,885 of them recorded by both networks. Filtering with the less restrictive criteria left 7,836 events with 142,190 P arrivals and 115,207 S arrivals for the joint CHILLAX-INPRES catalogue and an additional 17,487 events with 219,047 P arrivals and
207,899 S arrivals for the CHILLAX only catalogue. Note that the steep dip seen in the deep events in the CHILLAX only catalogue (Figure 3.13B) discussed in Chapter 1 disappears when the INPRES data is included, leaving only the horizontal trend of the slab (Figure 3.11B). This confirms that, as suspected, the steepening effect is a consequence of out-of-aperture locations. In the relocated catalogue we also see the concentration of earthquakes trending ~ 70° between 31°-32°S (Figure 3.10A) and the limbs of the double seismic zone (Figure 3.12)

### 3.5 Discussion

Several studies have remarked on the correlation between the JFR and the clustering of events we see trending at ~70° (e.g., Smalley & Isacks, 1987; Yañez et al., 2002), strongly suggesting the subduction of this feature as the cause of this dense cluster of activity. The JFR is composed of seamounts, guyots, and islands that trend roughly East-West (~ 80°) extending ~30 km in width and ~ 800 km in length offshore (von Huene et al., 1997; Rodrigo & Lara, 2014; Lara et al., 2018). It presently intersects the trench at ~33.4°S after migrating southward over time (Yanez et al., 2001; Bello-Gonzales et al., 2018; Lara et al., 2018). This relationship between the JFR and the concentration of events we see in the Nazca slab (Figure 3.14) led us to look for similar features that might explain the NNE-SSW (20°) trends in the seismicity. An inspection of Nazca plate bathymetry (Figure 3.15) shows that the linear trends parallel the Nazca and Iquique Ridges, although these features are far to the north of the JFR. During the late Oligocene-early Miocene, the JFR apparently was parallel to the present-day trend of the Nazca ridge and Iquique Ridge (Bello-González et al., 2018; Lara et al., 2018) but this would not explain the more recent 20° linear trends. Looking more closely at the Nazca plate bathymetry in our region of study (Figure 3.16), one can see several less prominent ridges that parallel the linear trends in seismicity. One of
these minor ridges that intersects the JFR at the Isla Alejandro Selkirk (Figure 3.17) shows a remarkably strong correlation with the westernmost lineation.

Figure 3.14: Map of our region of study (Figure 1.7) extended west to 80°W to include the projection of the JFR (orange line) based on Bello- González et al. (2018). The black circles represent the deep events from the well constrained SIEMBRA-ESP catalogue. Elevation is indicated in the palette at the bottom of the figure.

Figure 3.15: Bathymetry of the Nazca plate, in meters, as indicated in the elevation palette at the top of the figure. Boxes show where Figures 3.16 and 3.17 are located. The thin and thick contour intervals are 1000m and 8000m, respectively. The black circles in the black box represent the well constrained SIEMBRA-ESP events in our region of study.
Figure 3.16: (A) Bathymetry north of the JFR, with depth in meters as indicated in the palette at the top of the figure. The thin and thick contour intervals are 500 m and 5000 m, respectively. Note the NNE-SSW trending ridges that appear as bright green features near 2800 m depth. (B) Gradient analysis based on edge detection where red represents negative gradient and blue positive gradient as indicated in the palette at the top of the figure. Note the same “green” features in (A) generate larger gradient signals than the surrounding regions.

Figure 3.17: Bathymetry in the region of Isla Alejandro Selkirk with depth in meters as indicated in the palette at the top of the figure. A NNE-SSW trending minor ridge north of Isla Alejandro Selkirk correlates with the westernmost trend in the deep seismicity.
This potential causal relationship between subducted bathymetric highs and clustering of seismicity far from the trench suggests that the effects of these features persist in a subduction regime long after they are consumed. Presently, the extent to which these features remain as coherent entities after they are subducted is a controversial topic. Different investigators have argued that when a seamount arrives at the margin, it may be decapitated and accreted onto the forearc, forming part of the accretionary prism (e.g., Yang et al., 2022), or it can remain intact, denting and uplifting the margin (Dominguez et al. 1998; 2000; Rosenbaum & Mo, 2011; Ruh et al., 2016), and possibly be decapitated further into the subduction channel. Watts et al. (2010) speculate that the timing of decapitation is determined by the thickness of the subduction channel in relation to the height of the feature, the strength of the overriding plate relative to the strength of the feature, the internal structure of the feature (e.g., presence of volcanic cores) and the buoyancy of the feature, i.e., locally vs regionally compensated. Wang & Bilek (2011) suggest that seamounts are more likely to have small pieces break off as they are dragged against the upper plate rather than being fully decapitated in the subduction channel. Our results suggest that bathymetric highs are likely to remain largely intact, persisting at least some 200km beyond their initial encounter with the trench. If this is the case, these intact minor ridges can add positive buoyancy to the slab, which would contribute to slab flattening. The subduction of bathymetric highs has also been linked to higher rates of subduction erosion (e.g., Hampel et al., 2004; Navarro-Arânguiz et al., 2022). Erosion can occur in several ways: (1) as the feature collides with the accretionary prism and drags material into the subduction channel; (2) in the wake of the feature as it faults the overriding plate, allowing sediments to slide into the subduction channel, through abrasion as the feature is dragged beneath the upper plate; and (3) hydrofracturing as overpressured fluids migrate to the subduction interface (e.g. Dominguez et al., 1998;2000; Ranero & von Huene,
In the long term, basal erosion can lead to thinning of the upper plate and thickening of the subduction channel in the region of the feature (e.g., Marcaillou et al., 2016) which will in turn affect the timing of decapitation of the seamount (Watts et al., 2010). The increase in subduction erosion due to bathymetric highs could add extra buoyant material on top of the slab, which in turn would alter its geometry. The 50km spacing between limbs would also seem to corroborate the inference of Goss et al. (2013) regarding the episodic nature of subduction erosion.

Bathymetric highs can also affect seismicity and interplate coupling, acting as an asperity or barrier to large earthquake rupture and propagation, although the manner in which they do so is controversial. As asperities, they could act to locally increase stress and friction and thus increase interplate coupling and generate large earthquakes (e.g., Watts et al., 2010; Rosenbaum & Mo, 2011; Wang & Lin, 2022). Conversely, as a barrier, bathymetric highs can increase normal stresses to the point where friction and the yield shear stress are too high to allow slip and rupture (Watts et al., 2010; Contreras-Reyes & Carrizo; 2011). Studies have also shown that seamounts tend to undergo ductile deformation and creep rather than brittle deformation, so stress is not accumulated but rather released aseismically as numerous small earthquakes (e.g., Watts et al., 2010; Rosenbaum & Mo, 2011; Bonnet et al., 2019). Wang and Bilek (2011) postulate that crosscutting fracture networks that develop as the seamount is subducted is the mechanism by which seamounts creep aseismically, resulting in a heterogeneous stress regime where fractures locally fail at different times randomly. They also suggest that in the rare occasion that large earthquakes are generated, it is due primarily to source and/or rupture complexity. Contreras-Reyes & Carrizo (2011) also propose that the thickness of the subduction channel above seamounts partially controls whether seamounts act as barriers or asperities, where a thicker channel smooths the
subduction interface and enhances lateral rupture propagation. Fluid-rich sediment from seamounts could also reduce interplate coupling (e.g., Mochizuki et al., 2008) thus contributing to aseismic creep. Alternatively, fluids could favor rupture propagation by increasing fluid pore pressure (Contreras-Reyes & Carrizo, 2011). In short, while there is no definitive conclusion as to whether seamounts act as asperities or barriers to earthquake rupture and propagation, the preservation of seamounts in the subduction channel are likely to play some role.

The coupling suggested by the spatial correlation between the seismicity in the subducting Nazca plate and the crust directly above at first seems unlikely because differences in temperature and lithology in the mantle and the crust would lead to different deformation regimes. At the same time, the geology of the region does not show any clear correlation with these trends in the crustal seismicity, implying a greater influence of deeper dynamics. While a local increase in stress could be associated with the bathymetric highs (Scholz & Small, 1997; Contreras-Reyes & Carrizo, 2011), such stress would tend to be concentrated near the feature and diffuse over long distances, and hence is unlikely to be a causative factor. The near-vertical nature of this correlation suggests that gravity (which defines vertical) in the form of negative buoyancy, plays a fundamental role. A viable inference is that this near-vertical correspondence between the deep and shallow seismicity is as a result of subducted bathymetric highs releasing volatiles that travel up into the crust, increasing the pore pressure and fracking the crust, which in turn activates the lower crustal seismicity (Figure 3.18).

The Nazca slab has been postulated to be hydrated due to extensive faulting in the outer rise region associated with the JFR and the bending of the plate (Kopp et al., 2004; Ranero et al., 2005; Fromm et al., 2006). With depth, pressure and temperature (PT) conditions are not favorable, for brittle deformation and faulting (e.g., Meade and Jeanloz, 1991). However, seismicity at
intermediate depths in flat slabs has been related to dehydration reactions (phase changes) and an increase in fluid pressure which results in apparent brittle deformation and reactivation of faults (e.g., Ammirati et al. 2015; Zheng et al., 2016). Porter et al. (2012) suggested that the PT conditions in the mantle of the Pampean flat slab cause dehydration of serpentinite and seismicity related to dehydration embrittlement in the oceanic mantle, in agreement with inferences from other studies (Gans et al., 2011; Marot et al., 2014; Linkimer et al., 2020). Dehydration of serpentinite has also been linked to the lower plane of double seismic zones (Hacker et al., 2003b). At the same time, Ammirati et al. (2015) postulate that slab seismicity is related to dehydration of the oceanic crust rather than the mantle. In either case, serpentinite is stable at temperatures less than 680°C and pressures less than 6.5 GPa (Zheng et al., 2016). Temperatures in the Pampean flat slab are likely less than 600°C, which promotes retention of water to intermediate depths (Marot et al., 2014; Manea et al., 2017). As pressure and temperature increases, the stability of serpentinite decreases, which promotes dehydration (Zheng et al., 2016). There is evidence of serpentinized mantle in the region above the flat slab as seen in low seismic velocities and high Vp/Vs ratios (Porter et al., 2012; Marot et al., 2014; Linkimer et al., 2020) and at the base of the South American plate where the Nazca Plate subducts under Chile as seen in a strong anisotropic signal (Nikulin et al., 2019). Progressive dehydration of the flat slab from west to east has been suggested by multiple studies (Wagner et al., 2006; 2008; Porter et al., 2012; Marot et al., 2014; Ammirati et al., 2015; Linkimer et al., 2020) and has been postulated to be related to the permeability of the slab changing with the transition from normal to flat subduction (Porter et al., 2012) or to migration of water deeper into the slab which is released later on (Linkimer et al., 2020). The trends observed in this study are specifically related to the bathymetric highs of the plate and not to the entire subducting oceanic lithosphere, suggesting that there is some characteristic of these features that promotes
devolatilization reactions more than “normal” oceanic crust or lithosphere. A possible explanation is that seamounts have been associated with transporting large amounts of water into the subduction channel compared to smoother oceanic lithosphere, increasing dehydration reactions and fluid release within the subduction channel (e.g., Ellis et al., 2015; Pommier & Evans, 2017; Chesley et al., 2021). Additionally, we note that the extra fluid introduced into the slab by minor ridges could add extra buoyancy which would contribute to slab flattening.

An alternative explanation that may more easily account for these trends in seismicity is the release of carbon dioxide due to decarbonation (e.g., Miller et al., 2004; Famin et al., 2008; Gunatilake & Miller, 2022) as bathymetric highs spend more time above the carbonate compensation depth (CCD) prior to subduction and hence would be capable of accumulating more biogenic carbonate than the surrounding sea floor. The seamounts of the JFR are between heights of 1,000m above sea level to 500m below sea level with a common base at 3,900m depth (Rodrigo & Lara, 2014; Lara et al., 2018) and the CCD is ~4,500m (Hebbeln et al., 2000). Additionally, the minor ridges in our region of study are well above 4,500m depth (Figure 3.15A) and studies in this region have provided evidence of carbonates on the ridges of the Nazca Plate (e.g., Hebbeln et al., 2000; Paul et al., 2019; Devey et al., 2021). Perkins et al. (2006) show that carbonates can persist up to depths of 100-200km in a subduction zone, supporting the inference that subducted carbonates can be a source for decarbonation in the region. We also note that carbonated melts have been found in the JFR (Devey et al., 2000) and in other regions of the Nazca Plate (Villagómez et al., 2014). Carbonated melts transport carbon from the mantle to the crust over a wide range of temperatures (Jones et al., 2013). More support for decarbonation as the mechanism of devolatilization is the predominance of extensional focal mechanisms in the Nazca slab as a relatively larger space is required for the release of carbon dioxide compared to water.
Decarbonation has also been suggested to be responsible for the aftershock sequence of the 2014 Iquique earthquake in Chile (Gunatilake & Miller, 2022). Furthermore, a magnetotelluric study done by Burd et al. (2013) showed regions of low resistivity in the South American plate which correlate with hypocenter locations we observe at shallower depths. There is also a region of low resistivity beneath the Sierras de Córdoba (Booker et al., 2004) that coincides with a low velocity zone which Porter et al. (2012) suggest occurs due to the release of fluids from the slab as it resumes normal subduction. These low resistivity zones could be due to the presence of water, but the connectivity required to lower resistivity is uncertain. Low resistivity can also be attributed to graphite, which even in small amounts is very conductive and has been postulated to form in the subducting and overriding plate through decarbonation of subducted carbonates (Galvez et al., 2013) and carbonated melts (Selway, 2014) in reducing conditions.

An important question to address is: How would volatiles travel from the subducting slab to the upper crust? Several studies have inferred the presence of structures that could act as conduits for fluid migration from the subducting plate to the overlying crust. For example, Farias et al. (2010) postulated a westward-dipping ramp detachment structure from the upper South American crust to the flat slab at 60km depth which Marot et al. (2014) associate with fluids migrating from the plate interface to the continental crust resulting in locally higher Vp/Vs ratios in the forearc crust. Marot et al. (2014) also suggest that detachment faults like this may be occurring throughout the continental mantle. Ammirati et al. (2016) infer a westward dipping thrust fault between the Chilenia and Cuyania terranes down to 40km depth that accommodates crustal deformation which may be linked to an east dipping shear zone that extends down to the Moho. Linkimer et al. (2020) connect this shear zone to an eastward-dipping paleosuture of a Gondwana subduction zone, which may also allow for hydration of the upper mantle. The presence
and reactivation of such paleosutures and faults in the region (e.g., Ramos et al., 2002; Alvarado et al., 2005) can act as zones of weakness and potential conduits of relatively high permeability. Finally, bathymetric highs can also play a role in creating and reactivating faults and fracture networks as they move through the subduction channel (Dominguez et al., 1998, 2000; Wang & Bilek, 2011; Rosenbaum & Mo, 2011; Marcaillou et al., 2016, Ruh et al., 2016).

Figure 3.18: A conceptual diagram of volatiles being released from seamounts on the subducting oceanic slab and creating seismicity in the slab, which ascend into the continental crust and create seismicity through increased pore pressures (i.e., “fracking”).
4. TOMOGRAPHIC IMAGING OF THE PAMPEAN FLAT SLAB USING AN AUGMENTED DATASET

4.1 Abstract

The main scientific objectives of this dissertation center on improving our understanding of the tectonics of subduction along the Andean margin, and in particular the phenomena related to flat slab subduction. We focus on the Pampean flat slab beneath central Chile and Argentina, in part because of an unusual pattern in body wave arrival times observed from that region, but also because the abundance of seismic data available from decades of recording on both sides of the border offer an outstanding opportunity to apply techniques of seismic tomography to probe the structure of the subsurface. In this chapter we discuss an application of these techniques to arrival times from local body waves and dispersion curves of surface waves derived from ambient noise and earthquakes to create 3D models of Vp, Vs, and Vp/Vs in the Pampean flat slab region. Because our dataset is extensive, we first analyze smaller datasets to obtain insights into how best to generate images with the entire dataset. We quantify the quality of our results by using different starting models as well as running sensitivity, checkerboard, and reconstruction tests. A comparison of results reveals that several features are independent of the starting model and are also resolvable as confirmed by the sensitivity tests. Among these robust features, we image a low velocity root-like structure that extends from beneath the Andes to the deflection in the flat slab. We interpret this feature to be an overthickened Andean crustal root that could be of either a felsic or mafic-ultramafic composition, not eclogitized and slightly hydrated. We also confirm results from other studies of a high velocity, high Vp/Vs region associated with a less hydrated and depleted South American mantle. Additionally, our results show that the subducting Nazca plate is associated with an anomalously low Vp/Vs ratio. There are two enigmatic low velocity zones
within and below the flat slab, more prominently seen in \( V_p \) than \( V_s \), which have not been previously reported by other studies. These low velocity zones are separated by one of high velocity which we postulate is due to orthopyroxene enrichment of the oceanic lithosphere. The eastern low velocity zone we suggest that while this zone could be a result of a rising hot asthenosphere, a low \( V_p/V_s \) ratio and an insufficiently lowered \( V_s \) make the presence of any melting unlikely. The western low velocity zone is much more evident in \( V_p \) than \( V_s \) and spatially correlates with the JFR. We offer two potential explanations for this anomaly: (1) increased silica content in the JFR and (2) the presence of supercritical fluid located within the fracture and fault networks associated with the JFR. While the features observed are intriguing, they fail to provide an adequate explanation for the anomalous travel times that originally motivated this analysis. This could be a result of the anomaly being too small for the resolving power of the current analysis. Another possibility is that our assumption that the signal is caused by heterogeneity may be invalid, and that it is caused instead by anisotropy. Azimuthal anisotropy from SKS shear wave splitting suggest a fast East-West direction in the flat slab region, which would be consistent with this explanation, but the effects on local body waves are difficult to evaluate at this stage and must await future investigation.

### 4.2 Introduction

The velocity of seismic waves is sensitive to variations in certain material properties such as density, bulk modulus, and shear modulus, which in turn are functions of temperature, pressure, and lithology. We can therefore make inferences about the interior of the Earth by estimating seismic velocities from variations in travel times and phase dispersion through seismic tomography. Seismic tomography is a class of inverse problem (Appendix B) that is (1) non-linear, (2) often ill-posed because of uneven distributions of observation, and (3) sometimes ill-
conditioned (small eigenvalues) that can lead to high sensitivity to noise. Typically, we seek to optimize an objective function in a least squares sense by iteratively solving a linearized version of it, at each stage comparing predicted to observed data as we update the model. In the application discussed here, we combine local earthquake tomography (LET) with surface wave dispersion modeling to create images of the subsurface in the Pampean flat slab region using seismic data recorded by the CHILLAX, CHARMSME, CHARGE, SIEMBRA, ESP, OVA99, CSN and INPRES networks (Figure 1.8). We originally attempted inclusion of a third type of data, teleseismic arrival times, but found that the complexity in waveforms across the networks we analyzed made the creation of a dataset by cross-correlation problematic due to the potential for cycle skips.

The motivation for using body waves in conjunction with surface waves is the improvement in resolution of shallower structure provided by surface waves. We note that a number of LET studies have already been done in this region with data from these networks, but with far fewer observations. For example, Wagner et al. (2005) used 142 events with 2,600 P arrivals and 1,600 S arrivals from CHARGE and the ISC and a grid spacing of 40 x 40 x 20 km. Marot et al. (2014) used 3,770 events with 55,128 P arrivals and 54,889 S arrivals from OVA99, CHARGE, CHARMSME, CHASE, and CSN with a grid spacing of 40 x 40 x 10 km. Here we use an order of magnitude more observations with about an 80% reduction in grid spacing.

4.3 Methodology

The data analysis techniques employed here follow the procedures of Roecker et al. (2004; 2006; 2017), Greenfield et al. (2016), and Comte et al. (2019). The subsurface is parameterized by assigning a slowness (inverse of the velocity) to a regularly spaced mesh of nodes (or “grid”); the slowness at any point between the nodes is found by trilinear interpolation:
\[ s = \sum_{i=1}^{8} s_i c_i \]  

(4.1)

where \( s_i \) is the slowness at grid point \( i \), and \( c_i \) is the corresponding interpolation coefficient.

### 4.3.1 The Body Wave Inverse Problem

The forward problem for body waves involves the calculation of travel times from each station to each grid node in the model using the 3D spherical eikonal solver (SPHFD; see discussion in Chapter 3). The inverse problem is solved by a linear approximation of the Taylor expansion of the observed travel time, \( T_{obs} \), which is a function of both the slowness, \( s_i \), and hypocenter location (latitude, longitude, depth and origin time), \( h_k \):

\[ T_{obs} - T_{cal} = \sum_{k=1}^{4} \frac{\partial T_{cal}}{\partial h_k} \Delta h_k + \sum_{i=1}^{m} \frac{\partial T_{cal}}{\partial s_i} \Delta s_i \]  

(4.2)

where \( T_{cal} \) is the calculated travel time and \( m \) is the number of model elements. The partial derivatives are a measure of the sensitivity of the travel times to changes in the model parameters.

To calculate the partial derivatives with respect to slowness we first trace a ray from the source to the station. The travel time \( T \) of a ray along its path \( l \), is given by:

\[ T = \int_l s \, dl \]  

(4.3)

Hence, slowness, \( s \), is the gradient in time along the path:

\[ s = \frac{\partial t}{\partial l} \]  

(4.4)

Since a ray is perpendicular to the wavefront (constant travel time surface) we can use the travel time tables determined by solving the eikonal equation to determine the ray path by following the maximum time gradient (i.e., the “slowest” direction) from any point in the model to the station. The time required to traverse a ray segment \( l \) is approximated as:
\[ T = s_m l \]  

where \( s_m \) is the slowness at the midpoint of a ray segment. Hence:

\[ \frac{\partial T}{\partial s_i} = \frac{\partial s_m}{\partial s_i} l \]  

Using the trilinear interpolation formula:

\[ T = \sum_{i=1}^{8} T_i c_i \]  

we can calculate the partial derivatives for slowness as:

\[ \frac{\partial T}{\partial s_i} = c_i l \]

Derivatives with respect to hypocenters are calculated in a similar fashion. For example, in the \( x \) direction:

\[ \frac{\partial T}{\partial h_1} = \frac{\partial T}{\partial x} = \sum_{i=1}^{8} T_i \frac{\partial c_i}{\partial x} \]

Note that the partial derivative of travel time with respect to origin time is simply 1. Hypocenter perturbations are solved for as part of the inverse problem, but they are not used to update the hypocenter coordinates because the location problem itself is highly nonlinear. Instead, we relocate the hypocenter in the new model at each iteration. We include the calculation of the hypocenter perturbations in a joint inversion because otherwise the inversion would be biased toward a model that keeps the hypocenter at its original location (a full discussion of this issue can be found in Roecker et al., 2006).

We solve for perturbations in \( V_p \) and the \( \frac{V_p}{V_s} \) ratio and calculate for \( V_s \) by dividing \( V_p \) by \( \frac{V_p}{V_s} \).

We do this in part because the \( \frac{V_p}{V_s} \) ratio is often useful in making inferences about lithology, but
also because the $V_p$ is often better resolved than $V_s$ as a result of the typically larger quantity and smaller uncertainties in P arrival times. This disproportionate resolution between $V_p$ and $V_s$ generally results in large errors in estimating the $\frac{V_p}{V_s}$ ratio by simple division. Defining the ratio at a grid point $i$ as $r_i = \frac{V_{pi}}{V_{si}} = \frac{s_{pi}}{s_{si}}$, a perturbation to the S wave slowness can be written as:

$$\Delta s_{si} = \Delta (r_i s_{pi}) = s_{pi} \Delta r_i + r_i \Delta s_{pi} \quad (4.10)$$

The difference between observed and calculated S wave travel times, $T_{sobs}$ and $T_{scal}$, is then:

$$T_{sobs} - T_{scal} = \sum_{k=1}^{4} \frac{\partial T_{scal}}{\partial h_k} \Delta h_k + \sum_{i=1}^{m} \frac{\partial T_{scal}}{\partial s_{si}} (s_{pi} \Delta r_i + r_i \Delta s_{pi}) \quad (4.11)$$

### 4.3.2 The Surface Wave Inverse Problem

The forward problem for surface wave dispersion involves calculating the phase delay time in the model as a function of period. To do this we first convert the shear wavespeed in the model to phase velocity, $c(w)$, using the locked-mode method of Gomberg & Masters (1988) and assume that each point in the 3D model can be represented as a 1D model that changes with depth (Montagner, 1986). Phase delays are then calculated as in Chapter 2 by integrating the phase slowness between two points along the great circle path. The inversion problem typically involves perturbations only to shear wave slowness, but we can take advantage of a minor sensitivity to $s_{pi}$ by using the same Taylor expansion of the observed S-wave travel time as above:

$$T_{sobs} - T_{scal} = \sum_{i=1}^{m} \frac{\partial T_{scal}}{\partial s_{si}} (s_{pi} \Delta r_i + r_i \Delta s_{pi}) \quad (4.12)$$

We then relate the surface wave phase slowness (the reciprocal of the phase velocity), $s_{ph}(w)$, and shear wave slowness using the chain rule:
\[ T_{\text{obs}}(w) - T_{\text{cal}}(w) = \sum_{i,j} \frac{\partial T(w)}{\partial s_{\text{phi}}(w)} \sum_k \frac{\partial s_{\text{phi}}(w)}{\partial s_{\text{ijk}}} \left( s_{\text{pijk}} \Delta r_{ijk} + r_{ijk} \Delta s_{\text{pijk}} \right) \]  

(4.13)

where \( i, j, k \) sum over latitude, longitude, and depth, respectively, and \( w \) is frequency. \( \frac{\partial T(w)}{\partial s_{\text{phi}}(w)} \) is calculated as described in Chapter 2, and

\[ \frac{\partial s_{\text{ph}}}{\partial s_s} = \frac{\partial (\frac{1}{c})}{\partial (\frac{1}{V_s})} = -\frac{\partial c}{c^2} \times -\frac{V_s^2}{\partial V_s} = \frac{\partial c}{\partial V_s} \times \frac{V_s^2}{c^2} \]  

(4.14)

where the partial derivative \( \frac{\partial c}{\partial V_s} \) is calculated using the Gomberg & Masters (1988) locked-mode method.

### 4.3.3 Tomography

We can solve for body wave and surface wave perturbations individually or jointly (simultaneously) by collecting all of the travel and/or delay time residuals in the vector \( \Delta d \), the sensitivities (the partial derivatives) in the matrix \( G \) and then estimate model perturbations, vector \( \Delta m \), by the some form of:

\[ \Delta m = G^{-1} \Delta d \]  

(4.15)

\( G \) is a large but typically sparse matrix because a given raypath encounters only a small subset of variables along its path. Hence, we save significant memory by storing only the non-zero values and their position in the matrix. Each row of the system of equations is weighted by the inverse of the square of the estimated data uncertainty (1/variance). Since particular variables (hypocenters, velocities, and ratios of velocity) have different units, their sensitivities are scaled relative to one another by applying columnar weights to the partial derivatives. Typically, we weight \( \frac{V_p}{V_s} \) and surface waves by a factor of 10 and hypocenters by a factor of 100 relative to \( V_p \). We can also scale
(or reweight) different types of observations to compensate for their relative abundance, and typically we multiply rows of surface wave observations by a factor of 10.

We also use a parallel computing adaptation of LSQR (Paige & Saunders, 1982 a,b - Appendix B) called PLSQR3 (Lee et al., 2013). PLSQR3 iteratively solves for $\Delta m$ until some condition (e.g. a tolerance level for fit or maximum number of iterations) is met. The resulting perturbations are smoothed using a moving window average and then added to the existing model. Hypocenters are relocated in the new model and the inversion process is repeated. This is repeated for several iterations until the residuals are sufficiently minimized. We also reevaluate the dataset for outliers at each iteration by applying quality criteria. Specifically, we require that observations above an absolute (0.5s) and percentage residual threshold (5%), and a maximum standard deviation of the residuals (0.8s), are not used in the subsequent iteration. Note that as the model improves, all residuals are reevaluated and, as a result, more observations tend to be included at each iteration.

The specification of size of the damper and moving average window (used for smoothing the perturbations) is somewhat arbitrary, so in order to find an optimal choice we test a range of dampers and windows and examine the trade-off curve between model roughness and residual variance. We estimate model roughness (model complexity) by calculating the root mean square of the second spatial derivative of velocity (Greenfield et al., 2016) through finite differences, e.g.:

$$\frac{d^2 v}{dx^2} = \frac{d([v(x + 1) - v(x)] - [v(x) - v(x - 1)])}{dx^2}$$

$$= \frac{d^2}{dx^2} [v(x + 1) - 2v(x) - v(x - 1)]$$  (4.16)
Note that the second derivative measures curvature (the change in gradient), so the smaller the curvature the “rougner” the model. Generally, we want to find the smoothest (least “rough”) model that can adequately explain our observations.

### 4.4 Application

Our approach to the analysis of the large joint dataset was to first perform preliminary inversions with subsets involving individual or combinations of a few networks. In this way we can understand how to properly set various parameters and also investigate structure in key areas at a smaller scale. We decided to begin with the SIEMBRA, ESP, and CSN networks largely because their temporal overlap allows us to image a part of the Andes critical to our science objectives. [N.B.: Numerous models were generated as part of the analysis reviewed in this chapter. We generated figures for all of them for documentation purposes and to allow readers to draw their own conclusions. For brevity and clarity, only representative figures are included in the body of this dissertation. The remainder can be found in Appendix D, and referred to here as needed as “Figure D.number”.]

#### 4.4.1 Inversion of Surface and Body Waves with the SIEMBRA Dataset

A shear wavespeed model was created by the analysis of 23,272 phase delays of Rayleigh waves between a period of 5s and 40s generated from ambient noise (Chapter 2) and used as a starting model in a joint inversion of surface and body wave data. This approach offers two advantages: (1) it can resolve long wavelength shallow structure and mitigate short wavelength artifacts (e.g., Roecker et al. 2017) and (2) ambient noise measurements provide absolute transit times while body wave arrival times depend on the unknown origin time and location of the event which can trade off with wave speed.
We parameterize the subsurface using nodes spaced ~5km in all directions (latitude, longitude and depth). The initial 3D Vs model generated from ANT surface wave observations alone started with a 1D model derived from a previous tomographic result for this area by Marot et al. (2014). A trial 3D inversion indicated that these wavespeeds were too high and a new 1D model was created using 1 month of ANT data (Figure 4.1) with a damping of 1000 and smoothing window of 21 grid points for 20 iterations. The resulting 1D model was then used as a starting model in a 3D Vs inversion using the same damping and a smoothing window of 11 grid points for 10 iterations (Figure 4.2). To generate a starting model for the joint inversion with body waves, a corresponding Vp model was created by scaling the Vs model using a Vp/Vs ratio of 1.72. A body wave dataset of 35,224 filtered events with 1,032,581P arrivals and 1,112,906 S arrivals was then jointly inverted with the surface wave phase delays for 10 iterations with a damping of 5000 and smoothing window of 7 grid points for latitude and longitude and 5 for depth (a shorter smoothing window for depth is used because there we expect more heterogeneity vertically than laterally).

Figure 4.1: Comparison of the 1D models used to generate the 3D surface wave model for the SIEMBRA network. The model by Marot et al. (2014) proved to be too slow (blue line) and the model generated with 1 month of data (dispersion measurements from the phase velocity maps) showed a dip in wavespeed from 100 to 200km that was determined to be poorly constrained. The preferred model (green line) was created with all of the data and is faster than the Marot et al (2014) model and the 1-month data model between 100 to 200 km.
Figure 4.2: Plot of variance (s²) vs roughness (“Rough” in units of km⁻¹s⁻¹) for different smoothing and damping pairs tested for 10 iterations in the creation of the 3D surface wave model for SIEMBRA. The optimal model was chosen as the 10th iteration model with a smoothing of 11 grid points and a damping of 1000 (blue circle).

Figure 4.3: The residuals from the joint inversion of body waves and surface waves for SIEMBRA after 10 iterations using the 3D surface wave model generated from a Vp/Vs ratio of 1.72 and the “1 month of data” 1D model (Figure 4.1). There is a positive offset in the S wave residuals (red) while the P wave residuals are more centered (blue). The negative offset is seen in arrivals from deep events and not from shallow events.
While joint inversions of this sort can seem straightforward in theory, in practice they can be quite challenging due to the nonlinear nature of the problem. The following summary of the many steps taken to arrive at a final model provide an illustration. An examination of the residuals from the initial joint inversion revealed a positive bias in the S residuals (Figure 4.3), which suggests that the shear wavespeed model is too fast. We examined the residuals from deep (>50km) and shallow (<= 50km) events and found that the shift originated primarily from the deeper (and more numerous) events. Because this type of offset could be a symptom of an inappropriate choice of Vp/Vs, we compared Wadati plots of (1) all the events, (2) the shallower events and (3) the deeper events and determined that the Vp/Vs ratio increased with depth from 1.72 to 1.74. However, rescaling the Vs model by 1.74 only partially solved the problem. Recalculating the surface wave model with different constraints and the entire ANT dataset showed that an original decrease in Vs between 100 and 200km depth (Figure 4.1) was not required, and the revised deeper structure was scaled by a revised Vp/Vs ratio (1.73) to obtain a new 3D surface wave model using the same damping-smoothing pair as before for 10 iterations (Figure D.1). While this new model reduced the offset in the residual mean for the deeper events, it generated the opposite offset for residuals from shallower events. While this was an improvement, it seemed clear that more emphasis needed to be given to the body wave arrival times. Hence, we changed the scaling of the surface waves and inverted all the data for 10 iterations followed by just the shallow body wave data for 15 iterations, and 5 more iterations with all of the data. This resulted in a negative offset in the deeper events, so we focused on the deeper body wave data for 10 iterations followed by 5 more iterations with all of the data. Our final model was achieved after 55 iterations, with an associated reduction in variance 0.6681 to 0.2374 s² (64%) for surface waves (relative to the 1D starting model) 0.0886 to 0.0418 s² (53%) for body waves (relative to the starting surface wave
While this procedure eventually led to a satisfactory result, one can appreciate that a joint inversion of this sort can require a significant amount of user intervention.

Figure 4.4: W-E cross-sections of the Vp model generated for SIEMBRA from the joint inversion of surface and body waves after 55 iterations using the starting model in (Figure D.1). The latitude of the section is shown at the lower left corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.1 km/s and 1 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude.
Figure 4.5: Same as Figure 4.4 for Vs. The thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively.
Figure 4.6: Same as Figure 4.4 for Vp/Vs. The thin and thick contour intervals are 0.01 and 0.05, respectively. The numbers correspond to features described in the text.

4.4.1.1 Results

The final model obtained from the joint inversion (Figures 4.4–4.6—a map-view and cross-sections without hypocenters are shown in Figure D.2) show a general trend of seismic velocities increasing with depth with velocities in the flat slab reaching 8.6 km/s for Vp and 5 km/s
for Vs. An additional zone of high velocity is found above the slab to the east where it subducts steeply into the mantle. The upper crust is highly heterogeneous between 0 and 40 km depth with low Vp/Vs (~1.65) appearing in several regions (labeled 1 in Figure 4.6). A sharp gradient above a high Vp/Vs zone appears where a lot of the shallower seismicity is concentrated. A prominent high Vp/Vs zone above the flat slab seismicity exists between 40 km and 90 km (labeled 2 in Figure 4.6). This anomaly has a maximum length of about 350 km with an interruption in its continuity at 30.4 °S, 30.6 °S and 30.8 °S. Finally, there are three regions of low Vp/Vs (~1.71) below the high Vp/Vs region above and within the slab (labeled 3 in Figure 4.6).

4.4.2 Local Earthquake Tomography with the SIEMBRA-ESP-CSN Dataset

As mentioned in Chapter 3, when augmenting the SIEMBRA-ESP catalog with the CSN arrival times, we found that the resulting standard deviations of residuals for many events were so large that they could be considered outliers. These large residuals are most likely due to the inadequacy of the 1D model used in the location process. In particular, this dataset includes many rays from events beneath SIEMBRA-ESP recorded by CSN stations that pass through a suspected wavespeed anomaly. As one of the primary scientific objectives of this dissertation is to understand the nature of this anomaly, these “outlier” events are potentially critical to our analysis and so we need to understand how best to include what at first would be classified as “bad data”. Simply throwing out large residuals at the start could generate a “self-fulfilling” image with little to no anomaly at all. We therefore investigated two approaches to creating a 3D velocity model with this combined dataset: (1) a “brute force” approach where we simply invert all the data and see if the correct structure can be recovered through iterative reevaluation of outliers, and (2) a “gentle” approach which tries to fit the arrivals at the CSN stations alone using fixed SIEMBRA-ESP hypocenters before the SIEMBRA-ESP arrival times are added.
4.4.2.1 “Brute Force” Approach

In applying the “brute force” approach, we merged the CSN dataset (a catalogue of 3,069 events located using CSN stations augmented with SIEMBRA-ESP arrival times) with arrival times from the 32,206 well constrained SIEMBRA-ESP events. We inverted this data (35,101 events with 1,232,441 P arrivals and 1,109,205 S arrivals) testing different choices for damping and smoothing (Figure 4.7). An optimal inversion was achieved with a damping of 500 and smoothing window of 7 grid points. After 10 iterations the variance of the residuals decreased from 0.0645 to 0.0383 s² (41%).

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Figure 4.7: Plot of P-wave variance (s²) vs P-wave Roughness (“Rough” for Vp in km/s) for different smoothing and damping pairs tested for 10 iterations in the creation of the “brute force” model for the SIEMBRA-ESP-CSN catalogue. The optimal model was chosen as the 10th iteration model with a smoothing of 7 grid points and a damping of 500 (blue circle).
Figure 4.8: W-E cross-sections of the $V_p$ model generated for the “brute force” approach to handling the SIEMBRA-ESP-CSN catalogue. The latitude of the section is shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.1 km/s and 1 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing though latitude. The numbers correspond to features described in the text.
Figure 4.9: Same as Figure 4.7 for Vs. The thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively.
Figure 4.10: Same as Figure 4.7 for Vp/Vs. The thin and thick contour intervals are 0.01 and 0.05, respectively.

The results show a clear improvement in the CSN hypocenters (Figures 4.8 to 4.10 – a map-view and cross-sections without hypocenters are shown in Figure D.3) locations follow the horizontal trend of the slab and correlate with a high wavespeed region (labeled 4 in Figure 4.8) with a Vp of 8.2 - 8.5 km/s and Vs of 4.7 – 4.9 km/s and a low Vp/Vs of 1.70 – 1.75. This high wavespeed region extends into the mantle above the slab (labeled 2 in Figure 4.8) and has a high
Vp/Vs ratio (1.75 - 1.82). There are regions of low velocity below the slab with two (labeled 3 in Figure 4.8) being prominent in the Vp model (~8km/s) and present but less prominent in the Vs model (~4.6 km/s). Near 69.5°W, Vp and Vs gradients steepen downward as a convex downward feature that extends from the crust to just above the seismic zone (labeled 1 in the Figure 4.8). There is a similar feature to the east (labeled 1 in Figure 4.8) which is more prominent at the higher latitudes. There is also a high Vp/Vs region above the flat slab that is inclined to the north-east and parallels a trend of low Vp/Vs beneath it (Figure 4.10).

### 4.4.2.2 “Gentle” Approach

In applying the “gentle” approach, we first associated the CSN catalog events to the well constrained SIEMBRA-ESP events as we have greater confidence in the latter and found 131 events sufficiently close in time and space. Similar to the canonical event discussed in Chapter 1, travel time residuals to the CSN stations from the SIEMBRA-ESP hypocenters show a trend of positive residuals in eastern Chile which become increasingly more negative to the west. These 131 events were then used to create a 3D model using a damping of 50 and a smoothing window of 7 nodes for 10 iterations (Figure D.4). The resulting model reduces the variance by more than 50% (from 0.1523 to 0.0670 s²). The image shows a high velocity body to the west and a convex downward low velocity feature extending from the crust to above the seismic zone. This model was then used as the starting model for an inversion using the entire CSN dataset (3,069 events) with the same damping and smoothing as before. The variance decreased only slightly after 10 iterations, suggesting that the starting model was already fitting the data reasonably well. The residuals showed a negative shift in the S-wave mean indicating that the Vs still may be too slow. Nevertheless, as with the brute force approach, the locations improved significantly: instead of
steepening in the region of the flat slab the relocated CSN hypocenters follow the expected horizontal trend (Figure 4.11).

![Figure 4.11: Comparison of the hypocenter locations from the original CSN dataset (red) and after an initial inversion with the dataset. Note that the apparent horizontal concentrations in hypocenters in the original locations, in particular the one at 150 km depth, are artifacts generated by gradients in the 1D model used in locating the hypocenters.](image)

We attempted to fix the shift in the S wave residual mean by running an inversion using only S wave arrivals as travel times (fixed hypocenters). We inverted the S wave dataset with the same damping and smoothing as above for Vs alone but found that fixing the hypocenters resulted in very slow convergence. We next tried relocating the hypocenters with both P and S arrivals but only inverting for Vs, and this step allowed a consistent solution with the S-wave residual mean being close to zero. We then ran the inversion on the complete dataset, solving for Vp and Vp/Vs as before with an increased damper of 500. The resulting model (Figure D.5) was then used as a starting point to invert the entire merged dataset (including the arrivals from 32,206 SIEMBRA-ESP events resulting in 34,474 events with 1,199,146 P arrivals and 1,087,051 S arrivals) to determine a final model to compare with the “brute force” model (Figures 4.12-4.14; a map-view and cross-sections without hypocenters are shown in Figure D.6).
Figure 4.12: W-E cross-sections of the Vp model generated for the “gentle” approach to handling the SIEMBRA-ESP-CSN catalogue. The latitude of the section is shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.1 km/s and 1 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude.
Figure 4.13: Same as figure 4.12 for Vs. The thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively. The numbers correspond to features described in the text.
Figure 4.14: Same as figure 4.12 for Vp/Vs. The thin and thick contour intervals are 0.01 and 0.05, respectively.

4.4.2.3 Comparison of “Gentle” and “Brute Force” Models

Both the final “gentle” and “brute force” models show a high Vp zone associated with much of the slab (labeled 4 in Figure 4.8) but this becomes somewhat incoherent at lower latitudes in the eastern region above the flat slab in the “gentle” model (Figure 4.12). These areas do not have enough data to allow reasonable resolution. The low Vp zones (labeled 3 in Figure 4.8) beneath the slab are also seen in both models but the western low velocity zone seen in the brute
force model is now seen only at lower latitudes and the eastern one at higher latitudes for the gentle Vp model (Figure 4.12). Both models show high Vs associated with the seismic zone and the low velocity region to the east is much clearer in the gentle model (Figure 4.13). At the same time, the low velocity zones to the west in the brute force model are barely seen in the gentle model. There is also a high velocity zone to the east at 150 km depth and a high velocity region at 50 km depth in the gentle model (labeled 5 & 6 respectively in Figure 4.13) that are not seen in the brute force model. We also observe the convex downward feature (labeled 1 in Figure 4.8) but see a very low velocity in the upper section (Figures 4.12 & 4.13). The high Vp/Vs region above the slab clearly seen in the brute force model (Figure 4.10) is segmented in the gentle model by regions of low Vp/Vs regions (Figure 4.14). There are also regions of high Vp/Vs in the gentle model below the seismic zone that do not appear in the brute force model.

We suspected that the starting model could be causing some of these discrepancies, so we ran the inversion using the merged catalogue of 34,474 events with the same 1D starting model used in the brute force and found little to no difference between this result and the brute force model. This suggests that the starting model used in the gentle analysis introduced artifacts early on that remained in the model. These artifacts are possibly due to the volume and spatial distribution of the data used in the creation of the starting model, i.e., the 131 events were not sufficient to generate a good starting model. A lesson we learn from this exercise for the larger scale regional inversion is to be cognizant of the effects of our starting model on the results.

4.4.2.4 Resolution Tests

General indicators of resolution and robustness, such as hit maps and checkerboard tests, were carried out on the brute force model as a means of quality control. Hit maps show the ray density or the number of times a given parameter was “sampled” or “hit” by the observations
In general, most regions above the slab are “well hit” (defined as more than 50 hits), while those regions below and to the east of the slab where we see the elongated and inclined Vp/Vs anomalies are much less well sampled and hence are likely to be less well constrained.

We also ran a checkerboard test with “cubic” checkers of 5 grid points of ~ 25 km linear dimension. Not surprisingly, results of this test largely confirm our conclusions from the hit maps concerning resolution where sampling is high, although the checkerboards suggest resolution is reasonable below the slab as well (Figure 4.16; a N-S cross-section is shown in Figure D.7).

Figure 4.15: W-E cross-sections of hit maps showing ray density as the number of times individual nodes are sampled. The latitude of the section is shown at the lower right corner of each panel. Colors are the number of hits as indicated in the palette at the top of the figure. Solid blue regions have more than 50 hits (i.e. what we consider to be relatively well constrained). White circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude.
Figure 4.16: Map-view of the result of the checkerboard resolution test for the SIEMBRA-ESP-CSN “brute force” model. The depth of the section is shown at the lower right corner of each panel. Colors are deviations from the background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.5% and 2%, respectively.
4.4.3 Regional Inversion

Having successfully created an image with a limited subset of data, we next seek to generate a “regional” image using the entire dataset over a larger volume. As discussed in Chapter 3, we generated augmented catalogues for CHARME-CSN, CHILLAX-INPRES, CHARGE, and SIEMBRA-ESP-CSN networks (for the inversion we used the relocated SIEMBRA-ESP-CSN events from the brute force model). The CHILLAX catalogue was originally generated by REST, and some additional observations from CSN and the OVA99 (4,730 events) catalogue exists only as a manually picked catalogue. The OVA99 events were relocated in a 1D model (averaged from the SIEMBRA joint inversion model) to use in the inversion. The events from each of these catalogues, that is, the CHARME-CSN, CHILLAX-INPRES, CHARGE, relocated SIEMBRA-ESP-CSN and relocated OVA 99 catalogues, were filtered using the following criteria: an accepted event must have a minimum of 10 phases with at least 3 S arrivals, a maximum residual standard deviation of 0.8s, a maximum total uncertainty of 25 km, a maximum network azimuthal gap of 180 degrees, must be deeper than 0 km and must not be near the edge of the model. Note that these thresholds are slightly lower than those imposed on the original SIEMBRA-ESP catalogue in order to allow more data from comparatively less seismically active regions. The elevation constraint is applied because, while many stations in the Andes are several kms above sea level, events with “negative depth” would be of marginal use in imaging the subsurface. The azimuthal gap restriction ensures that all hypocenters used are within the aperture of the recording stations. The final dataset is composed of 49,406 earthquakes with 1,332,982 P arrivals and 1,213,722 S arrivals (Figure 4.17). We also have available a set of ambient noise derived Raleigh wave phase delays from the CHILLAX network from Comte et al. (2019) in addition to those generated for the SIEMBRA model (Chapter 2).
A lesson learned from the SIEMBRA-ESP-CSN (SEC) inversion is that an image generated by these datasets can be strongly influenced by the starting model. One way to mitigate this influence is to construct a model based on analyses by previous investigators. These analyses also often extend outside of the current region of interest and hence provide a larger scale context in which to interpret results. However, while this approach could have the advantage of starting the inversion closer to a global minimum, the veracity of such models is often difficult to quantify, and we risk introducing extraneous artifacts into our own models. Our approach in the present analysis is to attempt inversions with both “minimalist” 1D models and more complicated 3D models and note which features appear to be robust.

While several models of this region generated by previous investigators exist, many of whom used data from the same networks, stitching together smaller scale models into a larger regional model can be problematic as requisite assumptions about smoothing and interpolation invariably generate artifacts and artificial gradients. Hence, we focused on larger scale models that are either global in nature or at least encompass a significant part of the Andes outside of our region of interest. We began by reviewing regional models of the Andes available at the IRIS Earth
Model Collaboration (EMC\(^1\)), and a global surface wave model generated by Priestley et al. (2018). The IRIS EMC currently contains eight models for South America; a review of them revealed that most show short wavelength heterogeneity of an unknown origin or robustness (or explanation) in our region of interest. Such heterogeneity is likely to remain in any inversion going forward and their influence would be difficult to explain in any subsequently developed model. Fortunately, the Priestley model and an ANT based model developed by Ward et al. (2013; 2014) are sufficiently smooth to avoid this problem. Rather conveniently, the Ward model is restricted to the upper 50 km (i.e. Andean crust) while the Priestley model is defined between 40-700 km depth. Hence, combining the two is relatively straightforward. We interpolated both models onto our regional grid and then combined them by using the Ward model from the surface to the 46 km depth level and the Priestly model at depths of 56 km and greater (Figure D.8).

Incorporating our SEC model and the additional CHILLAX-SIEMBRA ANT observations pose the same issues of introducing artifacts caused by incorporating a smaller scale model into a larger one. We decided to first generate a Vs model that would be compatible with all sets of surface wave observations by jointly inverting our CHILLAX and SIEMBRA ANT observations with a dataset consistent with the Priestley/Ward model. We do not have the data used to create these models, and, even if we did, most of the surface wave paths used in those studies would originate well outside of our region. We therefore generated a virtual data set for the interpolated Priestley/Ward model by setting up virtual sources and receivers at grid points located near the edges of the model and calculating phase delays between them. Delays were calculated at periods in increments of 5s from 5-50s, and in increments of 20s from 60-140s (this spans the spectrum from the short period ANT observations of Ward et al. (2014) to the longest period used by

\(^1\) https://ds.iris.edu/ds/products/emc/
Priestley et al. (2018). Virtual data uncertainties (weights) were assigned as a percentage of the phase delay and scaled to be compatible with those determined for the local CHILLAX-SIEMBRA ANT observations. By inverting these virtual observations simultaneously with the local ANT observations, we seek to generate a model compatible with all of them. A starting Vs model was created by inverting all of the available observations for three iterations with the same damping and a smoothing used in the SEC inversion but with a larger grid spacing of 8 km as we are imaging a significantly larger volume (Figure D.9). One important result of this joint inversion is a reduction of Vs in the upper mantle beneath the South American plate, similar to a feature observed in the “brute force” model.

We attempted to create a compatible Vp model at this stage by allowing Vp sensitivities to be used in the generation of the local/regional Vs model from surface wave inversion (in this case we solve for Vp and Vp/Vs instead of Vs alone), but the Vp sensitivities were (perhaps not surprisingly) too small to have any meaningful effect. Hence, for all intents and purposes we have generated a starting Vs model, and as before we need to estimate a reasonable Vp model to accompany it.

As noted above, the image generated can be sensitive to the initial choice of Vp/Vs, and so several options were tried: (1) a depth dependent Vp/Vs based on a standard Earth model (IASP91), (2) a constant Vp/Vs (1.74) based on a Wadati plot for the entire combined dataset (as opposed to the that used for the SIEMBRA data set alone; Figure 4.18), and (3) a Vp/Vs variation based on a 1D average of the SIEMBRA model described above.

An initial inversion of the body wave arrival time data set (Figure D.10) using the 3D surface wave model described above scaled by the IASP91 Vp/Vs estimates showed that, while Vs variations relative to the background are reasonable, IASP91 Vp/Vs values clearly are much
too high. This was not surprising as the global values are mostly higher than the average taken from the Wadati plots, but confirms that the body wave data set is, as one would hope, capable of resolving absolute values of Vp/Vs.

![Wadati plot of S-P times vs P times for the regional body wave data set. A total of 1,045,523 S-P times were used to determine the Vp/Vs ratio. The red line represents the slope which gave a best fit least squares estimate of 1.74 for Vp/Vs.](image)

A second trial inversion with the 3D Vs starting model scaled by a constant Vp/Vs of 1.74 was initially carried out as a joint inversion with surface wave delay times (Figures D.11). However, the variance of the surface wave data set did not change from its starting value, indicating that nearly all of the signal from the surface wave data set was exhausted. As the joint inversion represents a significant computational burden, we switched to including only the body wave data set. The final model from this approach (Figures 4.19 - 4.21; a map-view and cross-sections without hypocenters are shown in Figure D.12) was obtained after 8 iterations and resulted in a reduction in residual variance from 0.0971 to 0.0592 s² (39%), and there was no significant offset in the residual mean.
Figure 4.19: W-E cross-sections of the final Vp model generated from the inversion of body wave arrival times after the surface wave data was exhausted in the joint inversion (model shown in Figure D.11). The latitude of the section is shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.1 km/s and 1 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude. The numbers represent the observations described in the text.
Figure 4.20: Same as Figure 4.19 for Vs. The thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively.
Figure 4.21: Same as Figure 4.19 for Vp/Vs. The thin and thick contour intervals are 0.01 and 0.05, respectively. The numbers correspond to features described in the text.
A third trial inversion of the body wave arrival times started with a 1D model generated by taking an average of the Priestley/Ward 3D starting model scaled with the Vp/Vs of 1.74 (Figure 4.22). The final model from this approach (Figure D.13) was obtained after 11 iterations and resulted in a reduction in residual variance from 0.1177 to 0.0659 s² (44%), and there was no significant offset in the residual mean.

A fourth and final trial inversion of the body wave arrival times started with a 1D model generated by taking a 1D average of the SIEMBRA model (Figure 4.22). The final model (Figures 4.23-4.25; a map view is shown in Figure D.14) was obtained after 15 iterations. The residual variance decreased from 0.0716 to 0.0457 s², (36%) with no offset in the residual mean.
Figure 4.23: W-E cross-sections of the Vp model generated from the inversion of body wave arrival times using the 1D model averaged from the SIEMBRA joint inversion as the starting model. The latitude of the section is shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.1 km/s and 1 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing though latitude. The numbers correspond to features described in the text.
Figure 4.24: Same as Figure 4.23 for Vs. The thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively.
Figure 4.25: Same as Figure 4.23 for Vp/Vs. The thin and thick contour intervals are 0.01 and 0.05, respectively. The numbers correspond to features described in the text.

4.4.3.1 Additional Tests of Robustness and Resolution

As noted above in the discussion of the SEC analysis, useful estimates of model robustness and resolution can be made from simple plots of observation sensitivity such as hit maps. For the regional data set, we decided to use a related and potentially more meaningful plot of the diagonal elements of the normal equation matrix, as these values indicate the total sensitivity of the inversion to changes in in a particular parameter. Plots of these this metrics (Figure 4.26; a map-view is shown in Figure D.15) show that the body waves sample most of the region very well from
the surface to about 145 km depth, with limited sampling down to about 165 km depth. Note that the longer period surface waves (e.g., from the Priestley model) will be sensitive to structure in the deepest parts of the model, but to a certain extent this sensitivity is “artificial” since we had to construct a virtual data set for these periods, and hence we do not include that dataset in our estimates of robustness.

Figure 4.26: W-E cross-section at 31°S of P-wave diagonals for the normal equations from the inversion of the 3D Priestly/Ward starting model with Vp/Vs of 1.74. Colors are the diagonals as indicated in the palette at the top of the figure. The solid blue region represents sensitivities greater than 1000 which likely represent parameters that are likely well resolved in an inversion as the damper is 500. The black circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude.

As with the SEC analysis, we carried out checkerboard tests to assess resolution, but in this case using only body wave arrival times as these form the basis for most of our interpretations. While the addition of the surface wave data will undoubtedly improve the recovery of these features, the use of virtual data will likely overestimate our ability to recover smaller scale features, and the regions covered by the SIEMBRA and CHILLAX ANT datasets are already well sampled by the body wave observations. Two checker target sizes were tested: the first composed of 5 x 5 grid points, or roughly 32-40 km cubes and the second 3 x 3 grid points, or roughly 16-24 km cubes (Figure D.16).
The results of these tests (Figures 4.27; a W-E and N-S cross section is shown in Figure D.16) suggest that the 32-40 km dimension targets are well resolved over much of the model where sensitivities (Figure 4.26) are >1000 and at both crustal and mantle depths, with diminishing resolution in the southeastern part of the model. Resolution remains reasonable at mantle depths, including those within, and potentially below, the Nazca slab, to about 150 km depth. Below that depth, ray path coverage is sparse. The 16-24 km targets (Figures 4.27; N-S cross section is shown in Figure D.16) are well recovered in the same regions as the 32-40 km targets at crustal depths but to a lesser degree at mantle depths. We conclude from these tests that features likely are well resolved at the 16-24 km level at crustal/uppermost mantle depths, and at the 32-40 km level in the mantle. For this reason, all of the features discussed below are larger than these dimensions.

An additional test that can give some indication of resolution and uncertainty is the “reconstruction” or “recovery” test (e.g., Prevot et al., 1991) which is carried out like the checkboard test but uses the actual 3D image as the target. Part the reason for this test is that the Earth is not made of checkers, and it is reasonable to ask if the data is capable of resolving features that are of similar dimension and location to the ones that we see in our images. A second reason is that LET images are known to underestimate the sizes of the actual anomalies because of trade-offs with hypocenter parameters, and the degree to which the amplitudes of anomalies are reconstructed gives us some idea of how severe that effect might be. The results of this test (Figure 4.31-4.33 & Figure D.17) show that, by and large, the features that we interpret here are well recovered but with some interesting details. First, all parts of the model are very well recovered at depths less than about 80 km (i.e., mostly crustal level features). At greater depths, and particularly within the subducted Nazca plate, Vs and Vp/Vs are well recovered in both location and amplitude, but Vp appears somewhat “washed out”, meaning that the main anomalies are evident but appear
as smaller amplitude features. There are several reasons why this might be the case, such as trade-offs with shallower structure, but may indicate that the actual variations in Vp are larger than our image would suggest.

Figure 4.27: Map-view of the results of checker resolution test for checkers with 5 x 5 grid points, or roughly 32-40 km cubes. Colors are deviations from the background as indicated in the palette at the top of the figure. The depth of the section is shown at the lower right corner of each panel. The thin and thick contour intervals are 0.5% and 2%, respectively.
Figure 4.28: Map-view of the results of checker resolution test for checkers with 3 x 3 grid points, or roughly 16-24 km cubes. Colors are deviations from the background as indicated in the palette at the top of the figure. The depth of the section is shown at the lower right corner of each panel. The thin and thick contour intervals are 0.5% and 2%, respectively.
Figure 4.29: W-E cross-sections of the results of reconstruction test using the Vp model from Figure D.13. The latitude of the section is shown at the lower right corner of each panel. Left panel shows target model and right panel the recovery result. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.1 km/s and 1 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude.
Figure 4.30: Same as Figure 4.29 for Vs. The thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively.
4.4.3.2 Summary of Results

By combining information from the models obtained with different starting models and the results of specialized tests (i.e., checkerboard, reconstruction, and sensitivity), we can discriminate robust features from possible artifacts. Plots of all the models that are not referred to directly in this section are included for comparison in Appendix D. We start with a verbal summary of these features and then discuss their potential significance.
Near the westernmost part of the model at ~71°W, where the Nazca plate begins its steep descent beneath the coast of Chile, there is a high concentration of seismicity and high Vp and Vs wavespeeds one expects from subducted oceanic lithosphere (Figures 4.19, 4.20, 4.23 & 4.24). These wavespeeds decrease to the east, between 70.5°W and 68.5°W and form a root-like structure in the upper 100 km that narrows with depth, resulting in a steep gradient from the continental crust to just above the seismic zone (labeled 1 on Figures 4.19 - 4.25). Vp and Vs within this root-like region increases with depth from crustal (Vp of 5.7 to 6.6 km/s and Vs of 3.3 to 3.8 km/s at < 50 km) to mantle (Vp of 6.6 to 8.0 km/s and Vs of 3.8 to 4.6 km/s between 50 – 100 km) values. Vp/Vs also increases from relatively low (1.68 to 1.74) to high (1.74 to 1.78) values between 50 to 100 km depth. A similar, but less prominent, low velocity feature appears to the east between 67.5°W and 66°W in the southern part of the model (labeled 1 on Figures 4.19 - 4.25). An intervening high velocity zone extends in depth from 60 km to just above the seismic zone with a Vp of 7.9 to 8.3 km/s and Vs between 4.5 to 4.8 km/s (labeled 2 Figures 4.19 - 4.25). This zone has an exceptionally high Vp/Vs of 1.76 to 1.83. In the northern part of the model, this high velocity zone extends to the east above and beyond the slab.

The seismicity in the flat part of the Nazca plate is concentrated between ~90 and ~130 km and extends for ~300 km. As noted previously, the plate appears contorted, bending back up to its shallowest depths after passing through a deeper bend. The high velocities one typically associates with a subducting slab are evident in the Vs image, with values between 4.6 to 4.9 km/s (Figures 4.20 & 4.24). Vp, however, is more complex. Particularly curious are two low Vp (7.3 to 8 km/s) zones within and below the slab (labeled 3 in on Figures 4.19 - 4.25) which are separated by relatively high velocities (8.1 to 8.5 km/s – labeled 4 in on Figures 4.19 - 4.25). The western low Vp zone correlates spatially with the JFR (e.g., Figure D.14A & D) while the eastern low Vp zone
is located where the slab bends to re-subduct steeply into the mantle. These low velocity zones appear in the Vs model as well but are much less prominent (~ 4.4 km/s). Vp/Vs is anomalously low (1.59) within the Nazca slab but uniformly high (1.74 to 1.79) in the South American mantle above it. Of particular interest is a lower Vp/Vs region above the slab between 31.8°S and 32.4°S that has a low Vp and high Vs region adjacent to where the lower velocity crustal region thins and spatially correlates with the inverted U shape in the seismicity.

Most of the cross sections shown here are either E-W or N-S, because the anomalies in map view appear to correlate with the concentrations of deeper seismicity, one may wonder if our perspective is dependent on azimuth. For this reason, we include cross sections of the model parallel and perpendicular to the strike of the subducted JFR in Figures D.18 & D.19. Double seismic zones are clearly defined in the western parts of the cross-sections and these seem to extend beneath the deflection of the slab. The lower parts of these zones are associated with low Vp/Vs. The slab has been inferred to be torn in the south at the transition from flat to normal subduction between 32°S and 33°S and ~100 km depth (e.g., Haddon & Porter, 2018). There is possible corroboration for this in our results as a relatively higher Vp/Vs ratio found from 32.5 °S to 33.5°S and 68.5°W to 67.5°W (e.g., Figure D.14A). In the cross-sections perpendicular to the JFR, where ray coverage is best between -150 km and 150 km NE (Figure D.19), we see evidence of a lower velocity zone at the 100 km position at depths below 100 km.

Finally, we note that most of the features we describe in this model also show up in the SEC “brute force” inversion where they overlap, despite having adopted a different starting model.

4.5 Discussion

The western lower velocity feature that correlates with the high Andes seems mostly likely to be an overthickened crustal root that reaches as deep as the top of the slab in the region of the
deflection. Based on the locations of the geological terranes of the region (e.g., Ramos et al., 2002; Pffiner, 2017), localized crustal thickening in this region would most likely involve the Chilenia and Cuyania terranes (Figure 1.7). While the Cuyania Terrane is characterized by mafic-ultramafic rocks (e.g., González-Menéndez et al., 2013.), the subsurface composition of the Chilenia Terrane is poorly constrained. It has been postulated to be composed either of more felsic rocks such as the La Pampa Gneisses north of the flat slab (Ribba et al., 1998; Alvarez et al., 2011) or mafic and ultramafic rocks further south of the flat slab (Ramos & Basei, 1997).

A useful tool for interpreting seismic images such as we present here is the Excel database/worksheet of Hacker & Abers (2004) and Abers & Hacker (2016), which calculates the physical properties, including seismic velocities and hydration, of minerals and rocks at different pressure and temperature (PT) conditions. Hence, it can be used to explore the extent to which a wavespeed model corresponds to a range of lithologies. This approach was used successfully by previous investigators in this region, in particular Marot et al. (2014) and Linkimer et al. (2020), and to a large extent we can interpret our model in the frameworks that they developed. Marot et al. (2014) recovered a similar root-like feature with low seismic velocities and a high \( V_p/V_s \) ratio beneath the high Andes but at shallower depths (\(~60\) km) than seen here. Using the database/worksheet of Hacker & Abers (2004), they calculated \( V_p \) and \( V_s \) for a range of relevant rocks and minerals and postulated that the crustal root beneath the Andes is hydrated, but not eclogitized, possibly due either to a pre-flat slab removal of an eclogitic root through a Rayleigh Taylor instability or to a more felsic composition of the Chilenia Terrane. Linkimer et al. (2020), used the Abers & Hacker (2016) database/worksheet and calculated \( V_s \) and \( V_p/V_s \) for a range of rocks and minerals characteristic of the mantle for PT conditions of 3 Gpa and 600°C, which correspond to the Pampean flat slab as modelled by Marot et al. (2014). They interpreted high
Vp/Vs ratios (>1.75), similar to what we see in the Andean root, as evidence of mafic and ultramafic rocks in the Chilenia and Cuyania terranes. However, the high Andes are beyond the bounds of their model.

Plotting our Vs and Vp/Vs results for the root on Linkimer et al.’s (2020) mineral graph (Figure 11 of their paper), we find that coesite is a viable mineral candidate. Coesite is formed when quartz is under high pressure and temperature and is consistent with a felsic continental crust. As Linkimer et al.’s (2020) analysis was biased towards mantle mineralogies, we investigated more minerals by creating similar plots of Vs vs Vp/Vs and an additional plot of Vp vs Vp/Vs and considered the minerals common to both (Figure 4.32 – red points in black rectangle). We found two less felsic minerals, prehnite and pargasite, that fit within our range of velocities. Prehnite is a hydrous mineral and is more stable than coesite at lower pressure and temperature conditions. Pargasite is also hydrous and has been inferred to form in regions of crustal thickening where the mantle is in contact with the crust (Hamdy et al., 2017; Kovacs et al., 2021). However, at these depths we expect lower to no percentage of these minerals, thus they should have minimal effect.

Plotting our results on Linkimer et al.’s (2020) rock graph suggests serpentine dunite (5% hydrated) as a possible explanation for what we see. There were other candidates, all hydrated, with wavespeeds just outside the limits of our observations; serpentinite wehrlite (5% hydrated), chlorite lherzolite (1.9% hydrated) and chlorite harzburgite (1.4% hydrated). We varied the percentages of hydrated chlorite and antigorite within these rocks to see if they would fall within our range of observations. The results from this analysis (Figure 4.33 – red points in black rectangle) shows that, with a slightly lower percentage of water, chlorite lherzolite (1.3% hydration) is still out of range out of range and chlorite harzburgite (1.3% hydration) is out of
range for Vp while serpentine dunite (3.7% hydration) and serpentine wehrlite (3.3% hydration) are more within the range.

Figure 4.32: Mineral analysis using the Excel worksheet from Abers & Hacker (2016). (A) Vs vs Vp/Vs and (B) Vp vs Vp/Vs. The colored boxes labeled 1, 2, 3, and 4 organize our observations as discussed in the text. Labeled minerals are within our range of observations. The gray shaded minerals are common to both plots within the ranges of our observations. Red represents calculations using a pressure of 3 GPa and temperature of 600 °C and blue the same with a pressure of 4.5 GPa and temperature of 600 °C.

While these rocks predict wavespeeds within the range of what we observe, we additionally examined felsic crustal rocks², particularly Andesite and Dacite, because of the association of this region with a thickened crustal root. As we do not know the exact mineral compositions of these rocks, we assume general compositions based on the QAPF diagram and mineral transformations based on molar masses. We also considered the mineral composition of the El Tránsito Metamorphic Complex as a possible representation of the felsic nature of the Chilenia Terrane as it is near the outcrops of the La Pampa Gneisses and are within the postulated extents of the Chilenia terrane (Alvarez et al., 2011). We expect that at these depths and PT conditions albite will transform to jadeite and quartz (e.g., Holland, 1980), anorthite to grossular, kyanite and quartz

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² www.earthchem.org
(e.g., Goldsmith, 1980) and quartz to coesite (e.g., Mirwald & Massone, 1980). We tested a spectrum of compositions for these rocks and found that all of the wavespeeds associated with the felsic rocks fit within the range of our observations; moreover, there were cases where wavespeeds from the original composition fit as well (examples Figure 4.33). Wavespeeds from several metamorphic facies, specifically amphibolite facies (1.3% hydration), epidote amphibolite facies (2% hydration), garnet amphibolite facies (1.2% hydration), granulite facies (0.5% hydration), epidote blueschist (3.1% hydration) and garnet granulite facies with no hydration, fit well within our range. Therefore, our results could be interpreted as a non-eclogitized root of either a felsic or a mafic-ultramafic composition with some combination of hydrated minerals.

Figure 4.33: Rock type analysis using the Excel worksheet from Abers & Hacker (2016). (A) Vs vs Vp/Vs and (B) Vp vs Vp/Vs. The colored boxes represent our observations labeled 1,2,3,4. Red represents a pressure of 3 GPa and temperature of 600 °C and blue represents a pressure of 4.5 GPa and temperature of 600 °C. Compositions of rocks: SW – ol57/tg27/ch13/cpx3, CL – ol66/opx16/ch15/cpx3, SD – ol60/tg30/ch10, CH – ol73/cpx16/ch11, P6 – ol20/opx70/ch10, P7 – ol20/opx80 compositions from Linkimer et al., (2020). P1 – opx20/l55/cpx5, P2 – opx40/l55/cpx5, P3 – opx60/l35/cpx5, P4 – opx80/l15/cpx5, P5 – opx100 from Qian et al. (2018) and can be found in Linkimer et al., (2020). EC, PYL and the metamorphic facies are from the Abers & Hacker (2016) database. SW2 – ol70/tg20/ch17/cpx3, CL2 – ol70/opx16/ch10/cpx3, SD2 – ol60/tg20/ch10, CH2 – ol64/opx30/ch10, EL1 – qz50/hAb20/an20/or20, EL2 – coe55/or15/lm30, based on Alvarez et al. (2011). AN1 – hAb49/an21/or15/lm30, AN2 – jd34.3/coes27.2/or20/gr9.9/ky9.7, AN3 – hAb24.5/an24.5/or15/lm20, AN4 – or21/lm20/coes19/jd17.2/gr11.5/ky11.3, DA1 – qz40/hAb20/an20/or20, DA2 – coe47.4/or20/jd14/gr9.9/ky9.2. We used the same mineral compositions for olivine (ol), chlorite (chl), clinopyroxene (cpx), orthopyroxene (opx), and garnet (gar) as Linkimer et al. (2020). atg – antigorite, qz – quartz, coe – coesite, hAb – high albite, mu – muscovite, an – anorthite, gr – grossular, jd – jadeite, ky – kyanite, or – orthoclase, hb – hornblende.
The eastern root-like feature correlates with the Sierra Pie de Palo and Sierra de Valle Fértil (Figure 1.7) but is unlikely to be associated with a deep crustal root like the Andes. Porter et al. (2012) noticed lower Vs (~ 4.4 to 4.5 km/s) in that region which they interpreted as evidence of mantle hydration. Previous studies have interpreted a monotonically increasing Vp and Vs and decreasing Vp/Vs from west to east in the slab, with an opposing trend in the overlying mantle, as progressive dehydration of the flat slab and hydration of the mantle (Wagner et al., 2006; 2008; Porter et al., 2012; Marot et al., 2014; Ammirati et al., 2015; Linkimer et al., 2020). Linkimer et al. (2020) also observe an inclined high Vp/Vs zone between 69°W and 68°W that they interpreted as a 5% hydration of depleted lherzolite or harzburgite, specifically chlorite-serpentine wehrlite and serpentine-chlorite dunite. They suggest that most of the fluid from the Nazca plate is being released in the center of the flat slab which decreases its volume until it re-subducts into the mantle. We observe fairly continuous high Vp, Vs and Vp/Vs above the flat slab which coincides with chlorite lherzolite (depleted lherzolite) with up to 1.9% hydration, chlorite harzburgite with up to 1.3% hydration, serpentinite wehrlite with up to 3.3% hydration, and pyrolite with no hydration (Figure 4.33- red points in blue rectangle). Other studies have attributed high velocities and higher Vp/Vs ratios to cold and dry depleted mantle peridotites (Wagner et al., 2005; 2006; Marot et al., 2014). Based on the petrological graphs of Marot et al. (2014) (Figure 11 in their paper), our observations for the higher velocity region plot in the range of dry peridotites but also within lower limit of their hydrous rocks with more magnesium content than iron. Thus, we infer that a cool, drier (slightly hydrated), depleted and Mg-rich, mantle exists above the flat slab.

One of the reasons Porter et al. (2012) suggested progressive dehydration is because they imaged low Vs zones in the oceanic mantle, similar to our Vs anomalies, which they associated with serpentinization of the oceanic mantle. However, we find that the Vp/Vs ratio is too low, and
Vp too reduced relative of Vs to permit an interpretation of serpentinization. The Vp anomalies in the slab are of particular interest as they are more prominent than the Vs anomalies. Portner et al. (2017) recovered low teleseismic Vp below the slab, which they attribute to the entrainment of hot asthenosphere from the JFR hot spot that is related to a hole in the slab centered at 32°S, 64°W and at about 350 km depth. A significant difference between our image and theirs is that their low Vp anomaly does not extend into the slab and is continuous along the length and width of the slab. Liu & Gao (2022) also see low Vs zones under the slab that they infer to be asthenosphere. The resistivity model of Burd et al. (2013) provides evidence of a plume rising from the 410 km discontinuity in a northeasterly direction that encounters the slab at 250 km depth where it is resubducting into the mantle. Their model also shows low resistivity beneath the slab around 30.5°S and 31.5°S. One may argue from these results that hot mantle is impinging on the slab and either thinning or tearing it (e.g., Gao et al., 2021). However, this does not explain why Vp would be more affected than Vs as Vs is more sensitive than Vp to melt and increasing temperature (Trampert et al., 2001; Tryggvason et al., 2002; Nugraha et al., 2019). Nevertheless, this could provide some context for the low velocity zone to the east which shows a significant decrease in Vp, Vs, and seismicity. There is only a slightly higher Vp/Vs in that region which could suggest that while asthenosphere is heating the slab, no melting is occurring.

The significant decrease in Vp compared to Vs found in the west spatially correlates with the JFR, suggesting that an explanation might lie in whatever distinguishes the JFR from the surrounding “normal” oceanic lithosphere. An increase in silica content has been shown to decrease both Vp and Vs with Vp being more sensitive (e.g., Marot et al., 2014). Silica can be deposited on the ocean floor prior to entering the subduction zone, as most of the dissolution of silica occurs in the upper water column above ~1000m. Dissolution decreases from ~1000m to
6000m before increasing again below 6000m (e.g., Rothwell, 2005; Petró et al., 2018). As most of the Nazca plate is above 6000m depth, there is probably no significant difference in the amount of silica deposited on the flat Nazca plate and the bathymetric highs. Another source for differing silica content could be in the formation of the JFR (ocean island basalts or OIB) vs the Nazca plate (mid oceanic ridge basalts or MORB). The JFR was created by the Juan Fernandez hotspot (Bello-Gonzales et al., 2018) which is believed to be sourced from a fixed weak (low temperature) deep-rooted mantle plume (Rodrigo & Lara 2014; Lara et al., 2018; Reyes et al., 2019). This resulted in the formation of theolitic basalts, alkalic basalts, trachy-basalts, pricritic basalts and basanites which have a silica content between 44 and 50 wt % (Gerlach et al., 1986; Farley et al., 1993; Reyes et al., 2017; Lara et al., 2018; Reyes et al., 2019), although lower concentrations of silica (between 36 and 40 wt %) have also been found in the JFR (Natland, 2003). By comparison, MORBs have a mean silica content of ~50 wt % (Hacker et al., 2003a; Gale et al., 2013). Estimates of silica content in MORBs associated with the East Pacific Rise are variable: Pandey et al. (2009) found lower silica content (~38 to 42 wt %) while other studies found an average silica content of 50 wt % at the rise itself (Reynolds et al., 1992; Niu et al., 1999), along the interior of the Nazca plate (Rhodes et al., 1976) and at the Chile Ridge (Karsten et al., 1996; Bach et al., 1996). Thus, the formation of the JFR does not appear to involve an anomalously high amount of silica compared to the rest of the Nazca Plate. Wagner et al. (2005; 2006; 2008) recovered anomalously high Vs, low Vp and high Vp/Vs above the slab which they attributed to orthopyroxene enrichment due to silica rich fluid metasomatizing mantle olivine. They suggested that a significant source of silica could be from erosion of the forearc as the JFR subducts. This would not explain the anomaly we find as it is deeper in the oceanic mantle. However, orthopyroxene found in the rocks that make up the JFR (Natland, 2003) could possibly be responsible for the signal we see. Considering
Linkimer et al.’s (2020) petrological analysis, our high Vs and low Vp/Vs ratios coincide with pyroxenites enriched with orthopyroxene as well as eclogite. On their graph they also plotted rock compositions from Qian et al.’s (2018) analysis of orthoenstatite which coincide with our range of observations. Linkimer et al.’s (2020) result assumed PT conditions above the slab so we recalculated using PT conditions for within and below the slab (~4.5 Gpa and ~600°C) based on the model of Marot et al. (2014). Our mineral analysis (Figure 4.32 – blue points) shows that enstatite, the Mg component of orthopyroxene, is within our range of our observations for box 4 (green rectangle). Ca tschermak and diopside, minerals that can be found in clinopyroxene where diopside is the Mg component, are also within range. We also see that glaucophane, which is a hydrous anisotropic mineral that is stable at colder subduction zones (e.g., Bang et al., 2021), falls within our range for box 3 (red rectangle). Glaucophane has been proposed as an explanation for low velocity zones at the top of subducting oceanic crust (Bezacier et al., 2010; Mookherjee & Bezacier, 2012), however, our low velocities are also observed well within the mantle. Our rock analysis (Figure 4.33 – blue points) shows no rock composition corresponding with box 3 but eclogite is just within our range of wavespeeds for box 4 and rocks with higher concentrations of orthopyroxene (without chlorite) fit comfortably in box 4. Thus, we infer that while rocks enriched with orthopyroxene could explain the part of the slab with higher Vp and Vs, they cannot explain the low Vp region. Another possible mechanism that could increase silica content is magmatic differentiation, as the more buoyant felsic components could rise and create a low Vp anomaly. While there does not seem to be any process in the formation of the JFR that would have this result, it has been suggested that the local flexing of the plate as it subducts can result in magma penetration and the formation of petit spot volcanism near the trench (Hirano et al., 2006). Petit
hot spot volcanoes were found near the JFR (Hirano et al., 2013) and could provide a means for preferential magmatic differentiation of the JFR in the outer rise prior to subduction.

Dense fracture networks in dry rocks have also been proposed to reduce Vp more significantly than Vs, decreasing the Vp/Vs ratio, with the opposite effect observed for saturated rocks (e.g., O’Connell & Budiansky, 1974; Nugraha et al., 2019). The region of the JFR is highly fractured and faulted (e.g., Kopp et al. 2004) and the basalts in the JFR are known to be highly vesicular (Devey et al., 2000; Natland, 2003). However, we expect permeability and porosity to decrease with depth owing to increasing pressure. Additionally, Porter et al. (2012) suggest that compressive stresses act on the upper part of the slab as it transitions from normal to flat subduction, reducing permeability. We do observe largely extensional focal mechanisms within the Pampean flat slab which could imply reactivation of fracture and fault networks as the slab flattens. These dense faults and fracture networks associated with the JFR could also contribute to another mechanism for reducing P wave velocity through the presence of supercritical fluids. Supercritical fluids occur beyond critical temperatures and pressures where the fluid has properties of both a liquid and a gas. At normal subduction zones, supercritical fluids are related to a mixture of aqueous fluid and hydrous melts in the mantle (e.g., Kessel et al., 2005; Klimm et al., 2008; Zheng & Herman, 2014) and also to the “melt-like” fluid produced as minerals are more easily dissolved in water at higher temperature and pressure (Ni et al., 2017). This could result in a more silica rich fluid in the slab that could also lead to our hypothesized orthopyroxene enrichment. Flat slab subduction zones are not associated with melt and are much cooler environments than normal subduction zones, and there is evidence of volatile rich supercritical fluids being released from flat slabs (Mungall, 2002; Bissig et al., 2003; Newell et al., 2015). Thus, we consider the possibility of supercritical fluids trapped in the flat slab derived from an influx of fluids infiltrating the JFR.
There is evidence that fluids penetrate at least into the upper oceanic mantle through faults and fractures associated with the JFR (Kopp et al., 2004). Higher saturation of supercritical fluids (water/steam and CO\textsubscript{2}) have been found to decrease V\textsubscript{p} more than V\textsubscript{s} and in turn result in a low V\textsubscript{p}/V\textsubscript{s} ratio at upper crustal depths in sandstone (e.g., Xue & Ohsumi, 2004; Daley et al., 2007) and volcanic regions (Tryggvason et al., 2002; Nugraha et al., 2019). Tryggvason et al. (2002) also observed high seismicity in the region of the interpreted supercritical fluids. Additionally, CO\textsubscript{2} in its gaseous state lowers V\textsubscript{p} to a greater extent than V\textsubscript{s} (e.g., Agofack et al., 2018) which would result in lower V\textsubscript{p}/V\textsubscript{s} ratios. This may provide an explanation for the lower V\textsubscript{p}/V\textsubscript{s} region we see starting at 31.8°S above the slab and supports our hypothesis that seamounts in the Nazca Plate are releasing CO\textsubscript{2} that is percolating up to and fracking the lower crust. If water were being released, we would expect a high V\textsubscript{p}/V\textsubscript{s} ratio as V\textsubscript{s} is more significantly reduced than V\textsubscript{p} by hydration. We note that Linkimer et al. (2020) also see a lower V\textsubscript{p}/V\textsubscript{s} zone (1.70 – 1.73) at 31.8°S, which they attribute to orthopyroxene enrichment. While intriguing, we emphasize that the velocity observations for supercritical fluids and CO\textsubscript{2} gas were made at crustal depths, and extrapolations to mantle depths should be made with caution.

Perhaps the most surprising result of the analysis discussed here is related to something we do not image. Specifically, there does not seem to be any evidence for the hypothesized high wavespeed body beneath the high Andes that provided the initial motivation for this project. Quite the contrary, we find an extensive lower wavespeed body that we attribute to the formation of an overthickened Andean crustal root. The relationship of this root to the bending of the slab is enigmatic because any remnant of South American lithosphere beneath the high Andes appears to have been sheared off or otherwise entrained by the Nazca plate, and it is difficult to conceive of the remaining continental crust providing much in the way of either a load or barrier to flexure.
So, what happened to the systematic pattern in residuals that we observed in Chile from events in the flat slab beneath Argentina? The east-west gradient of residuals could be explained by a similar gradient in wavespeeds increasing to the west, and the existence of lower than expected velocities to the east of Chile satisfies that criterion. However, an examination of residuals from some of the CHILLAX arrivals from the flat slab, including the M4.9 event on 2015/12/26 discussed in Chapter 1, show that much of that signal remains unexplained. In the case of the M4.9 event, the original 0 to -2 s residual gradient is now closer to 1 to -1 s, indicating the effect of the root has been mostly to demean that signal from these events. This is precisely the effect we were hoping to avoid by taking the “gentle” versus “brute force” approach to accommodating these residuals. However, despite taking this route, many of these observations were still classified as outliers during the processing.

One possibility is that these residuals are caused not by heterogeneity, as we have assumed, but by anisotropy. The variations in raypath azimuths in the combined body wave dataset should allow us to recover an isotropic background, and, in a sense, a potentially large remnant anisotropic signal corroborates that presumption. We note that an azimuthal anisotropy study by Anderson et al. (2004) using SKS shear wave splitting suggests E-W fast polarization in the region of the flat slab, which is consistent with the residual gradient we observe between Argentina and Chile. Mapping anisotropy at the scale required here is a far from trivial endeavor, however, and will have to wait for the next dissertation.
5. CONCLUDING REMARKS

The Pampean flat slab subduction zone is an interesting phenomenon and understanding the dynamics of such a feature holds implications for the wider study of Earth Science past, present and future. In carrying out this dissertation, unexpected findings were made far beyond what we originally sought out to investigate, which is the way of how research often works. The initial discovery of anomalous travel times propelled us along the path of testing the hypothesis of a lithospheric root trapped above the Pampean flat slab. In the data collection phase, a curious clustering of earthquakes was found in the flat slab; 4 NNE-SSW trending parallel lineations that intersect the JFR and are separated about 50 km from each other. Even more curious is the near vertical correspondence with these linear trends and the seismicity seen in the continental crust. These trends correlate to minor ridges seen on the Nazca plate and we postulate that volatiles are being released from these ridges, through dehydration or decarbonation, increasing pore pressure in the upper crust resulting in the coupling between the deep and shallow seismicity. The regional images created also show a region of low Vp/Vs within the mantle that correlate with an inverted U shape seen in the crustal seismicity that could be evidence of CO₂ rising from the slab to the crust. The images also revealed some other interesting results, specifically, a low wavespeed structure, interpreted as the crustal root of the Andes, that reaches the depths of the flat slab and two low Vp anomalies within and below the flat slab. The two low Vp anomalies are very remarkable because the Vs is not as affected. We postulate that the eastern Vp anomaly is due to hot asthenosphere rising but no melting because the Vs is not significantly reduced such that the Vp/Vs ratio is low. The western low Vp anomaly correlates with the JFR and we hypothesize that it is either due to an increase in silica content or supercritical fluids trapped within the JFR.
Intriguingly, much of the original travel time anomaly that initiated this dissertation is still present. We postulate that it could be due to some feature being smaller than what our model could resolve or that there is some anisotropic effect that our assumption of isotropy cannot account for. Either way, while we do have a better image of the Pampean flat slab region, there remains more work to do which we will leave for future investigations.
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A. PASSIVE SOURCE SEISMIC INTERFEROMETRY

In this appendix we go through the theory of how the Green’s Function (GF) is extracted from cross-correlation by first looking at general signal processing theory for extracting the impulse response in a discrete time series and then looking at how it relates to the earth and ambient noise in 1D, 2D and 3D. Lastly, we show that the Green’s Function can also be extracted from the time derivative of the cross-correlation function.

A.1 Extracting the Impulse Response from Discrete Time Signals

This section is based on theory from Proakis & Manolakis (2007).

\[ y(n) = h(n) * x(n) \]  \hspace{1cm} (A.1)

where \( h(n) \) is the impulse response of the linear system and \( * \) represents convolution.

In an LTI system the cross-correlation function, \( r_{yx}(t) \), of the output, \( y(n) \), and the input, \( x(n) \), has a value of \( h(n) * r_{xx}(n) \) where \( r_{xx}(n) \), is the autocorrelation function, that is, when \( y(n) = x(n) \).
Proof:

The cross-correlation function is defined as:

\[
 r_{yx}(t) = \int_{-\infty}^{\infty} y(n)x(n - t) \, dn = \sum_{n=-\infty}^{\infty} y(n)x(n - t) = y(t) * x(-t) \\
= h(t) * x(t) * x(-t)
\]  

(A.2)

* and the time reversal of \( x(t) \) represents cross-correlation.

But

\[
 x(t) * x(-t) = \int_{-\infty}^{\infty} x(n)x(n - t) \, dn = \sum_{n=-\infty}^{\infty} x(n)x(n - t) = r_{xx}(t)
\]  

(A.3)

So, the cross-correlation of the output signal and the input signal is:

\[
 r_{yx}(t) = h(t) * r_{xx}(t)
\]  

(A.4)

when \( x(t) \) is white noise (what we attempt for with ambient noise tomography), the autocorrelation is the Dirac delta function, so the impulse response can be extracted from the cross-correlation:

\[
 r_{yx}(t) = h(t)
\]  

(A.5)

A.2 Extracting the Green’s Function from the Cross-Correlation Function

The Earth can be considered an LTI system, so as above, we can extract the Green’s Function (GF) from the cross-correlation of ambient noise. This section is a summary of theory from Wapenaar et al. (2010).

Consider two receivers along the x-axis at \( A \) and \( B \) recording the direct waves radiating from an impulsive source (Figure A.2). Assuming a constant velocity and a lossless medium, the response (the GF) at \( A \) is \( G_{SA}(t) \) and at \( B \) is \( G_{SB}(t) \). For these responses there is an overlap of the wave paths from the source to the receiver at \( A \). If we cross-correlate the common path is cancelled resulting in the receiver at \( A \) acting as a virtual source with a response \( G_{AB}(t) \) at \( B \).
Figure A.2: Direct wave paths from an impulsive source recorded by receivers A and B. The signal from source to Receiver is shown by the green arrow. The signal to Receiver B is shown by the green and yellow arrow where the green part represents the overlap between the signal traveling to A and B and the yellow represents the signal between A and B.

Thus, the characteristics of the actual source (such as position, time, and velocity) do not need to be known in order to obtain the GF propagating from A to B.

\[ G_{AB}(t) = G_{SB}(t) \ast G_{SA}(-t) \]  
(A.6)

If the source, \( s(t) \), is not impulsive, then the responses at A and B become:

\[ u_{SA}(t) = G_{SA}(t) \ast s(t) \]  
(A.7)

\[ u_{SB}(t) = G_{SB}(t) \ast s(t) \]  
(A.8)

The autocorrelation of the source function is:

\[ S_s(t) = s(t) \ast s(-t) \]  
(A.9)

So, the cross-correlation of the responses is:

\[ u_{SB}(t) \ast u_{SA}(-t) = [G_{SB}(t) \ast s(t)] \ast [G_{SA}(-t) \ast s(-t)] = G_{AB}(t) \ast S_s(t) \]  
(A.10)

In the 2D and 3D cases we consider multiple independent sources surrounding the two receivers fired sequentially from different azimuths (Figure A.3). If we sum the cross-correlations for all the sources, the sources that contribute the most are in the first Fresnel zone, as such, the waves that travel between A and B will add constructively and the waves that do not (outside the Fresnel zone) will add destructively which gives a time-symmetric response:

\[ \sum u_{SB}(t) \ast u_{SA}(-t) = [G_{AB}(t) + G_{BA}(-t)] \ast S_s(t) \]  
(A.11)
where $G_{AB}(t)$ is the GF at $B$ for a virtual source at $A$ and represents the causal part of the response (positive lag), and $G_{BA}(-t)$ is the time-reversed GF at $A$ for a virtual source at $B$ and represents the acausal part of the response (negative lag).

In terms of ambient noise, the noise sources occur simultaneously so the responses are naturally summed. Assuming that the noise sources are uncorrelated (independent):

$$< u_B(t) * u_A(-t) > = [G_{AB}(t) + G_{BA}(-t)] * S_N(t) = C_{AB}(t)$$

(A.12)

Figure A.3: Noise sources at all azimuths recorded at receivers A and B. Gray highlighted areas are the first Fresnel zones where there is constructive interference.

In an attempt to achieve this, normalization and whitening are done before cross-correlation.
A.3 Extracting the Green’s Function from the Time Derivative of the Cross-correlation Function

Another approach to extracting the GF from ambient noise is to take the time derivative of the cross-correlation function (Lobokis & Weaver, 2001; Weaver & Lobkis, 2004, Sneider, 2004; Roux et al., 2005; Sabra et al., 2005a,b; Bardos et al., 2008; Lin et al., 2008). This approach retrieves an anti-symmetric response (Wapenaar et al., 2010; Boschi & Weemstra, 2015). We can transform from anti-symmetric to the symmetric and vice versa by differentiating in the time domain (Wapenaar et al., 2010).

The time derivative of the cross-correlation function is defined as

\[
\frac{dC_{AB}(t)}{dt} = -G_{AB}(t) + G_{BA}(-t)
\]

(A.13)

Such that,

\[
G_{AB}(t) = -\frac{dC_{AB}(t)}{dt}, t \geq 0
\]

(A.14)

\[
G_{BA}(t) = -\frac{dC_{AB}(-t)}{dt}, t \geq 0
\]

(A.15)

Proof:

S. Roecker derived the equation for the cross-correlation function following from Yao & van der Hilst (2009).

![Figure A.4: A plane wave with an azimuth $\theta$ (ray paths – solid red arrows, wave front – red dashed line). The black triangles are the stations A and B with an azimuth of $\varphi$ from A to B. (Based on Yao & van der Hilst, 2009).](image)
Consider a plane wave travelling with angular frequency $\omega$ at an azimuth of $\theta$ along the path of station A and station B (Figure A.4). The time delay between A and B is $\delta t$ with a phase delay of $\delta \phi = \omega \delta t$. The waves with amplitude, $A(\omega, \theta)$, can be described as:

\[ W_A = A(\omega, \theta) \sin(\omega t) \quad (A.16) \]
\[ W_B = A(\omega, \theta) \sin(\omega(t - \delta t)) = A(\omega, \theta) \sin(\omega t - \delta \phi) \quad (A.17) \]

In the time domain, the cross correlation between $W_A$ and $W_B$ over the period $T$ is defined as

\[ C_{AB}(\tau) = \int_0^T W_A(t)W_B(t + \tau)dt \quad (A.18) \]

\[ C_{AB}(\tau) = A^2(\omega, \theta) \int_0^T \sin(\omega t) \sin(\omega(t - \delta t + \tau))dt \quad (A.19) \]

From

\[ \cos(a + b) = \cos(a) \cos(b) - \sin(a) \sin(b) \quad (A.20) \]
\[ \cos(a - b) = \cos(a) \cos(b) + \sin(a) \sin(b) \quad (A.21) \]

We get

\[ \sin(a) \sin(b) = \frac{\cos(a - b) - \cos(a + b)}{2} \quad (A.22) \]

Setting $a = \omega t$ and $b = \omega(t - \delta t + \tau)$

\[ a - b = \omega t - \omega(t - \delta t + \tau) = \omega(\delta t - \tau) \quad (A.23) \]
\[ a + b = \omega t + \omega(t - \delta t + \tau) = 2\omega t - \omega(\delta t - \tau) \quad (A.24) \]

So, the cross-correlation function becomes,

\[ C_{AB}(\tau) = \frac{A^2(\omega, \theta)}{2} \int_0^T \cos(\omega(\delta t - \tau)) - \cos(2\omega t - \omega(\delta t - \tau))dt \quad (A.25) \]

Integrating

\[ \int_0^T \cos(\omega(\delta t - \tau)) dt = T \cos(\omega(\delta t - \tau)) \quad (A.26) \]
and
\[
\int_0^T \cos(2\omega t - \omega(\delta t - \tau)) dt = \frac{1}{2\omega} [\sin(2\omega T - \omega(\delta t - \tau)) - \sin(-\omega(\delta t - \tau))] = 0
\]  
\[
\frac{1}{2\omega} [\sin(4\pi - \omega(\delta t - \tau)) - \sin(-\omega(\delta t - \tau))] = 0
\]
gives
\[
C_{AB}(\tau) = \frac{A^2(\omega, \theta)}{2} T \cos(\omega(\delta t - \tau)) = \frac{A^2(\omega, \theta)}{2} T \cos(\omega(\tau - \delta t))
\]  
(A.28)

So, for plane waves at all azimuths we get an equation for the cross-correlation function which is similar to equation 1 in the Yao and van der Hilst (2009) paper if we set \(E_p(\omega, \theta) = \frac{A^2(\omega, \theta)}{2} T\):

\[
C_{AB}(\omega, t) = \int_0^{2\pi} E_p(\omega, \theta) \cos (\omega(t - \delta t)) d\theta
\]  
(A.29)

where \(E_p(\omega, \theta)\) is a general expression for the amplitude at a particular frequency and azimuth.

The theoretical Green’s Function, according to Dahlen & Tromp (1998) for a far-field fundamental mode surface wave is approximately:

\[
G_{AB}(\omega, t) = A \cos \left(k_{AB} \Delta_{AB} - \omega t + \frac{\pi}{4} \right)
\]  
(A.30)

where \(A\) represents amplitude, \(k_{AB} = \omega/c\) is the path averaged wavenumber; \(k_{AB} \Delta_{AB} = \omega \Delta_{AB} / c = \omega t_{AB} = \delta \phi_{AB}\) is the phase delay between A and B so

\[
G_{AB}(\omega, t) = A \cos \left(\delta \phi_{AB} - \omega t + \frac{\pi}{4} \right)
\]  
(A.31)

If we take the derivative of the cross-correlation function we get,

\[
\frac{dC_{AB}(\omega, t)}{dt} = -\int_0^{2\pi} \omega E_p(\omega, \theta) \sin(\omega(t - \delta t)) d\theta
\]  
\[
\int_0^{2\pi} \omega E_p(\omega, \theta) \sin(\delta \phi_{AB} - \omega t) d\theta
\]  
(A.32)
This resembles the theoretical GF where the $\frac{\pi}{4}$ phase shift in the theoretical GF is assumed to be resolved by the integration over the azimuths.

Thus, the cross-correlation function is related to the GF as follows:

\[
G_{AB}(\omega, t) = -\frac{dC_{AB}(\omega, t)}{dt}
\]  
(A.33)
B. THE INVERSE PROBLEM

The following discussion reviews the basics of inverse theory used throughout the analyses discussed in this dissertation, and is based on discussions with S.Roecker (personal communication), as well as Tarantola & Valette (1982b), Menke (1989), Aster et al. (2005), Tarantola (2005) and Press et al. (2007), Nollet (2008), Shearer (2012) and Ammon et al. (2020).

The processes by which we make predictions of observations (e.g., through the laws of physics) by specifying boundary or starting conditions are often referred to as “forward modelling”. The reverse of this process, using observations to deduce these conditions, is referred to as “inverse modeling”. Usually we perform inverse modeling, or “solve the inverse problem”, by first assuming a certain physical state, calculating predictions for that state with the forward problem, and then adjust that state (the “model”) to minimize an objective function that typically depends on the differences between the predictions made by the model and the actual observations.

The forward problem can be described as:

\[
g(m) = d
\]

where \( g \) is the operator (some function that relates \( m \) to \( d \) in the physical system under study), \( m \) are the model parameters and \( d \) are the observations.

Taking the Taylor expansion of \( g(m) \) about a starting point \( m_0 \):

\[
g(m) = g(m_0) + \sum_i \frac{\partial g}{\partial m_i}(m_i - m_{0i}) + O(m^2)
\]

We typically assume that the higher order terms are either negligible or can be managed through a series of linear steps (iterations) so that we only need to solve the linear problem:

\[
g(m) - g(m_0) = \sum_i \frac{\partial g}{\partial m_i}(m_i - m_{0i}) = d - d_0 = \text{residual}
\]

where \( d \) is the actual observation and \( d_0 \) is the predicted observation at \( m_0 \). In matrix form:
\( G \Delta m = \Delta d \)  

where \( G \) is the matrix that holds the sensitivities of the observations to the model parameters, \( \Delta d \) is the vector of residuals (the difference between the observed and predicted values) and \( \Delta m \) is the vector of perturbations (changes in the model required to reduce the residuals).

The simplest form of the inverse problem is:

\[ \Delta m = G^{-1} \Delta d \]

where we solve for \( \Delta m \) and add it to the current model parameters. For a nonlinear forward problem, we repeat this over some number of iterations until we arrive at a value within a statistically relevant threshold. However, this approach will work only if \( G \) is invertible, that is, if \( G \) is a square matrix meaning that the number of observations is equal to the number of unknowns. Generally, we have to deal with an ill posed problem that may be overdetermined (more observations than unknowns), underdetermined (more unknowns than data so multiple models can fit the data) or mixed determined (one part of the model is overdetermined and another is underdetermined). The preferred method for solving such a problem is the least squares method:

\[ \min || \Delta d - G \Delta m ||^2 \]

This method is attractive because it often is computationally more tractable as the size of the \( G \) matrix remains the same no matter the number of observations. On the other hand, this method is sensitive to outliers (blunders) because the minimization of a square causes it to preferentially accommodate large residuals.

The least squares method aims to minimize the sum of the squares of the residuals, \( E \), that is, the square of the Euclidean norm of the residual vector, \( e \), so:

\[ E = e^T e \]
Thus, for a system of linear equations $Ax = b$, where $A$ is the coefficient matrix, $x$ is the vector of unknowns and $b$ is the vector of constants, then $e = b - Ax$ and

$$E = (b - Ax)^T (b - Ax) \quad (B.8)$$

Since

$$(x + y)^T = x^T + y^T \quad (B.9)$$

and

$$x^T y = y^T x \quad (B.10)$$

then

$$E = b^T b - 2A^T x^T b + A^T Ax x \quad (B.11)$$

The minimum occurs when the derivative of $E$ is 0. Using the following properties for matrix derivatives:

$$\frac{d (x^T x)}{dx} = 2x \quad (B.12)$$

$$\frac{d (x^T a)}{dx} = a \quad (B.13)$$

We get:

$$\frac{\partial E}{\partial x} = -2A^T b + 2A^T Ax = 0 \quad (B.14)$$

which gives the following system of equations:

$$A^T Ax = A^T b \quad (B.15)$$

These are called the normal equations of the least squares problem. Now, applying the theory of normal equations to $G(\Delta m) = \Delta d$ we get:

$$G^T G \Delta m = G^T \Delta d \quad (B.16)$$

$G^T G$ is a square matrix and, if invertible (i.e., no singularities):

$$\Delta m = (G^T G)^{-1} G^T \Delta d \quad (B.17)$$
Depending on the application, we may not want to solve the least squares problem directly from the normal equations because of a tendency for high condition numbers that form near singular matrices that enhances errors. A more robust option can be to use singular value decomposition, as it is more robust when dealing with noisy data (Appendix B.1).

Practically, we have to consider noise and uncertainties in the data where small eigenvalues can cause even small uncertainties to have a significant effect on the solution, leading to divergence. One option is to stabilize the solution by adding a damper (a constant), \( \varepsilon^2 \), to the diagonal of \( G^T G \): 

\[
\Delta m = (G^T G + \varepsilon^2 I)^{-1} G^T \Delta d
\]

where \( I \) is the identity matrix. This “damped” least squares solution prevents ‘overstepping’ by keeping changes from one iteration to the next from being too large. Adopting a Bayesian point of view, Tarantola & Valette (1982b) show how the damper takes into consideration model and measurement uncertainty. They derive a least squares solution as:

\[
m_{k+1} = m_k + \left[ G_k^T \cdot C_{dd}^{-1} G_k + C_{mm}^{-1} \right]^{-1} \cdot \left[ G_k^T \cdot C_{dd}^{-1} \cdot (d_0 - g(m_k)) - C_{mm}^{-1} \cdot (m_k - m_0) \right]
\]

where \( m_{k+1} \) is the model at the \( k + 1 \) iteration, \( m_k \) is the current model, \( G_k \) is the linear operator, \( C_{dd} \) is the covariance of the data (uncertainties in the observations), \( C_{mm} \) is the covariance of the model (anticipated uncertainties in the predicted values), \( d_0 - g(m_k) \) is the current residual (the difference between the observation and the prediction from the current model) and \( m_0 \) is the starting model. This equation is similar to the previous least squares equation with the addition of the covariance matrices and the extra term \( C_{mm}^{-1} \cdot (m_k - m_0) \). This extra term keeps the solution close to the starting model and practically is useful only if there is confidence in the starting model. The diagonal elements of the \( C_{dd} \) matrix correspond to the variances of the
observations and the off diagonals to the covariance between the different observations. Usually, we presume the observations are independent of each other so \( C_{dd} \) is a diagonal matrix. The diagonals of the \( C_{mm} \) matrix represent a threshold within which the predicted value should lie (variances of the prior probability density for the individual model parameters), and the off diagonals represents coupling between the points in the model (in practical terms, the smoothness of the model). One could enforce a type of smoothing by specifying off diagonal elements in this matrix, but for a number of reasons related to stability and efficiency we here prefer to use a diagonal matrix and enforce coupling through a posteriori smoothing. Hence, we define the matrices as follows:

\[
C_{dd} = \sigma_{dd}^2 I \quad \text{(B.20)}
\]

\[
C_{mm} = \sigma_{mm}^2 I \quad \text{(B.21)}
\]

Note that these covariance matrices allow us to give each observation or model parameter its own uncertainty, but for illustration purposes we consider constant uncertainty, \( \sigma^2 \). So:

\[
m_{k+1} = m_k + \sigma_{dd}^2 \left[ G_k^T \sigma_{dd}^{-2} G_k + \sigma_{mm}^{-2} I \right]^{-1} \cdot \left[ G_k^T \sigma_{dd}^{-2} (d_0 - g(m_k)) \right] \quad \text{(B.22)}
\]

\[
m_{k+1} = m_k + \sigma_{dd}^2 \left[ G_k^T G_k + \frac{\sigma_{dd}^2}{\sigma_{mm}^2} I \right]^{-1} \cdot \left[ G_k^T \sigma_{dd}^{-2} (d_0 - g(m_k)) \right] \quad \text{(B.23)}
\]

\[
m_{k+1} = m_k + \left[ G_k^T G_k + \frac{\sigma_{dd}^2}{\sigma_{mm}^2} I \right]^{-1} \cdot \left[ G_k^T (d_0 - g(m_k)) \right] \quad \text{(B.24)}
\]

\[
m_{k+1} = m_k + \left[ G_k^T G_k + \varepsilon^2 I \right]^{-1} \cdot \left[ G_k^T (d_0 - g(m_k)) \right] \quad \text{(B.25)}
\]

where \( \frac{\sigma_{dd}^2}{\sigma_{mm}^2} = \varepsilon^2 \), which is the damper (note that we can apply the above approach to variable uncertainties by scaling the columns of the G matrix).

We see from the above that the damper is the ratio between the uncertainty in the data to the uncertainty in the model (note that one can also think of “uncertainty” in this context as “level
of confidence”). Therefore, a large damper will be a result of a large uncertainty in the data and/or a small uncertainty in the model and a small damper will be a result of the opposite. Note that the larger the damper the smaller the steps which suggests we are more confident in the previous model than the data, so we want to stay close to it. The trade-off is that larger dampers can result in more steps for optimization. Conversely, if we have less confidence in our previous model, we can take larger steps by using a smaller damper with the trade-off that a damper which is too small could lead to divergence. The addition of the damper also governs a trade-off between the fit and resolution of the solution where the damper lowers the uncertainty in the solution but also results in a loss in resolution (Appendix B.1).

As data collection techniques have improved over the years, we often have to deal with much larger problems where solving the least squares problem explicitly as described above can be computationally intractable. In this case, we take advantage of $G$ being a sparse matrix and store the non-zero values and their position in the matrix. We also apply a version of the LSQR algorithm (Paige & Saunders 1982 a,b), an iterative projection method that finds an approximate solution to the damped least-squares problem:

$$\begin{bmatrix} A \\ \lambda I \end{bmatrix} x = \begin{bmatrix} b \\ 0 \end{bmatrix} \quad (B.26)$$

which can be written as

$$\begin{bmatrix} I & A \\ A^T & -\lambda^2 I \end{bmatrix} \begin{bmatrix} r \\ x \end{bmatrix} = \begin{bmatrix} b \\ 0 \end{bmatrix} \quad (B.27)$$

where $r = b - Ax$. LSQR applies the Lanczos algorithm for $k + 1$ iterations to reduce the above to a lower diagonal system, such that the above equation becomes:

$$\begin{bmatrix} I & B_k \\ B_k^T & -\lambda^2 I \end{bmatrix} \begin{bmatrix} t_{k+1} \\ y_k \end{bmatrix} = \begin{bmatrix} B_1 e_1 \\ 0 \end{bmatrix} \quad (B.28)$$
\[
\begin{bmatrix}
  r_k \\
  x_k
\end{bmatrix} =
\begin{bmatrix}
  U_{k+1} & 0 \\
  0 & V_k
\end{bmatrix}
\begin{bmatrix}
  t_{k+1} \\
  y_k
\end{bmatrix}
\]

(B.29)

where \( B_k \) is the lower diagonal with Lanczos scalars \( \alpha_k \) and \( B_{k+1}, U_k \) and \( V_k \) are the matrices that hold the Lanczos vectors, \( r_k = b - Ax_k, x_k = V_k y_k, t_{k+1} = B_1 e_1 - B_k y_k \)

Note that \( y_k \) is the damped least squares solution to

\[
\begin{bmatrix}
  B_k \\
  \lambda I
\end{bmatrix}
\begin{bmatrix}
  y_k
\end{bmatrix} =
\begin{bmatrix}
  B_1 e_1 \\
  0
\end{bmatrix}
\]

(B.30)

which can be solved using orthogonal transformations (QR Factorization) and gives an approximate solution to \( Ax = b \).

Although this approach is reliable and efficient, we lose the ability to explicitly calculate the covariance and resolution matrices, which may be used to assess the quality of the solution. Instead, we can apply sensitivity tests that provide estimates of the quality of the solution and verification that the data can actually recover the parameters seen in the final model. Sensitivity tests can use a known structure and the original observations to create a synthetic dataset which is then used to recover the structure using the same inversion process done to get the final model. The difference between the known structure and the recovered structure gives insights into the resolution. Two popular types of trial models often used are; (1) A checkerboard test where the known structure is equally spaced alternating negative and positive perturbations (checkers) to some reference model and (2) the final model itself in a reconstruction test to see if the specific variations of wavespeed (magnitude and geometry) can be recovered. The former test mimics a point spread function reminiscent of the resolution matrix, while the latter is often used to test how well relevant geometries and scale lengths (i.e., acknowledging that the Earth is not made of checkers) can be recovered. Additionally, specific features in a model can be specified and recovered as a test of robustness.
B.1 Singular Value Decomposition of the Least-Squares Problem

The following is based on S.Roecker (personal communication), Menke (1989), Aster et al. (2005) and Press et al. (2007).

The singular value decomposition (SVD) of an $m \times n$ matrix $G$ is:

$$ G = U \Sigma V^T $$  \hspace{1cm} (B.31)

where $U$ is an $m \times m$ orthogonal matrix spanning the data space (m observations), $V$ is an $n \times n$ orthogonal matrix spanning the model space (n parameters), and $\Sigma$ is an $m \times n$ matrix of singular values (non-negative real scalars) typically arranged in descending order.

We can derive the SVD of matrix $G$ by first defining a square matrix, $S$ as

$$ S = \begin{bmatrix} 0 & G \bar{G} \\ \bar{G} & 0 \end{bmatrix} $$  \hspace{1cm} (B.32)

where $\bar{G}$ is the complex conjugate transpose of $G$. Since $S$ is a square matrix and $S^T = S$, there exists an orthogonal set of eigenvectors $w$ and real eigenvalues $\lambda$, such that:

$$ Sw_i = \lambda_i w_i $$  \hspace{1cm} (B.33)

The nontrivial solution for the eigenvalues is:

$$ (S - I\lambda_i)w_i = 0 $$  \hspace{1cm} (B.34)

and will only occur if the determinant of the bracketed quantity is zero. This gives the characteristic equation:

$$ det(S - I\lambda_i) = 0 $$  \hspace{1cm} (B.35)

We can solve for $\lambda$ and $w$ by breaking $w$ into two vectors $u$ and $v$. That is:

$$ Sw_i = \lambda_i w_i $$  \hspace{1cm} (B.36)

becomes

$$ \begin{bmatrix} 0 & G \\ \bar{G} & 0 \end{bmatrix} \begin{bmatrix} u_i \\ v_i \end{bmatrix} = \lambda_i \begin{bmatrix} u_i \\ v_i \end{bmatrix} $$  \hspace{1cm} (B.37)
Giving us two coupled equations:

\[ Gv_i = \lambda_i u_i \]  
\[ \tilde{G}u_i = \lambda_i v_i \]  

(B.38)  
(B.39)

There are two sets of solutions: (1) the zero eigenvalues case, and (2) the nonzero eigenvalues case. In the zero case \( u \) and \( v \) are independent, \( Gv_i = 0 \) and \( \tilde{G}u_i = 0 \). In the non-zero case, we have an orthogonal set of eigenvectors with real eigenvalues and can describe the \( u \) eigenvectors as matrix \( U \) and the \( v \) eigenvectors as matrix \( V \). We can divide the \( U \) and \( V \) matrices into a “p” space (nonzero eigenvalues case) and “o” space (zero eigenvalues case):

\[ U = (U_p, U_o) \]  
\[ V = (V_p, V_o) \]  

(B.40)  
(B.41)

We now write the equations as:

\[ GV_p = U_p \Lambda_p \]  
\[ \tilde{G}U_p = V_p \Lambda_p \]  
\[ GV_o = 0 \]  
\[ \tilde{G}U_o = 0 \]  

(B.42)  
(B.43)  
(B.44)  
(B.45)

So, equation \( Gv_i = \lambda_i u_i \) becomes:

\[ GV = G[V_p, V_o] = [U_p, U_o] \begin{bmatrix} \Lambda_p & 0 \\ 0 & 0 \end{bmatrix} \]  

(B.46)

Multiply by \( V^T \)

\[ GVV^T = [U_p, U_o] \begin{bmatrix} \Lambda_p & 0 \\ 0 & 0 \end{bmatrix} \begin{bmatrix} V_p \\ V_o \end{bmatrix} \]  

(B.47)

But \( VV^T = I \) and \( U_o \) and \( V_o \) are multiplied by zeros and do not contribute to \( G \) and the overall inverse problem so the singular value decomposition of \( G \) is:

\[ G = U_p \Lambda_p \tilde{V}_p \]  

(B.48)
This is equivalent to \( G = UAV^T \) but instead constructing \( G \) with only the “p” space.

Let’s relate this to the inverse problem. If there are no zero singular values, then the generalized inverse operator is:

\[
G_g^{-1} = V_p \Lambda_p^{-1} U_p^T
\]  

(B.49)

So, for the equation

\[
Gm = d
\]

(B.50)

\[
m = G_g^{-1} d
\]

(B.51)

\[
m = V_p \Lambda_p^{-1} U_p^T d
\]

(B.52)

This is the least squares solution.

Proof:

If we multiply \( Gm = d \) by \( G^T \) then:

\[
G^TGm = G^Td
\]

(B.53)

\[
m = (G^TG)^{-1} G^Td
\]

(B.54)

If there is a \( U_o \) space and no \( V_o \) space, then:

\[
G^TG = V_p \Lambda_p U_p^T U_p \Lambda_p V_p^T = V_p \Lambda_p V_p^T
\]

(B.55)

So

\[
(G^TG)^{-1} = V_p \Lambda_p^{-2} V_p^T
\]

(B.56)

and

\[
m = V_p \Lambda_p^{-2} V_p^T V_p \Lambda_p U_p^T d = V_p \Lambda_p^{-1} U_p^T d
\]

(B.57)

\[
m_g = G_g^{-1} d
\]

(B.58)

**B.2 The Resolution and Covariance Matrices**

If we substitute \( Gm = d \) into the generalized inverse, then:

\[
m_g = G_g^{-1} Gm
\]

\[
= V_p \Lambda_p^{-1} U_p^T U_p \Lambda_p V_p^T m
\]

\[
= V_p V_p^T m
\]

(B.59)
If there is no $V_0$ space then $V_p V_p^T = I$ so the inverse problem is perfectly resolved that is, $m_g = m$. The product $V_p V_p^T$ is the Model Resolution Matrix. If $V_p V_p^T \neq I$, that is, there are zero eigenvalues so there are off diagonal terms then there is smearing of the solution (‘fuzziness’).

If we adjust the generalized inverse to include uncertainties in the data and model, we can write the problem in terms of covariance where we multiply each term by a weighting term (its transpose) as follows:

$$\langle \Delta m_g \Delta m_g^T \rangle = G_g^{-1} \langle \Delta d \Delta d^T \rangle G_g^{-1}$$

(B.60)

where $\langle \rangle$ represents the expected value. Letting the data covariance be a constant, $\sigma_d^2$,

$$\langle \Delta m_g \Delta m_g^T \rangle = \sigma_d^2 G_g^{-1} G_g^{-1} = \sigma_d^2 V_p \Lambda_p^{-1} U_p^T U_p \Lambda_p^{-1} V_p^T$$

$$= \sigma_d^2 V_p \Lambda_p^{-2} V_p^T$$

$$= \sigma_d^2 (G^T G)^{-1}$$

(B.61)

Hence, we can recover an estimate of the uncertainties in the solution (the covariance of the solution). From this equation we can see that for a non-linear solution, the uncertainty in the model is sensitive to the singular values. If the singular values of $\Lambda_p^{-2}$ are small then the linear steps will be too large and will overshoot the solution and the uncertainties will be large, leading to divergence. We can regularize the solution by adding a damper to the singular values to compensate for small singular values, decreasing the uncertainty. However, this affects the resolution. Consider

$$m_d = (G^T G + \epsilon^2 I)^{-1} G^T d$$

(B.62)

$$m_d = V_p E \Lambda_p^{-1} U_p^T d$$

(B.63)

where $E$ is a diagonal matrix with elements $\frac{\lambda_i^2}{\lambda_i^2 + \epsilon^2}$. If we substitute $Gm = d$, then:

$$m_d = V_p EV_p^T m$$

(B.64)
So, when deriving the resolution matrix, instead of $V_p V_p^T$ as before, there is an extra term which causes a degradation in the resolution, resulting in a trade-off between the resolution and error.
C. DERIVATION OF THE EIKONAL EQUATION

We calculate travel time tables using the eikonal equation, which relates the travel time of a ray in a medium to its velocity structure. The eikonal equation is a high frequency approximation of the wave equation in a weakly inhomogeneous medium (small velocity gradients). Based on theory from Slawinski (2007), Robinson & Clark (2017) and Ammon et al. (2020).

The acoustic wave equation in an isotropic, homogenous medium is:

\[ \frac{\partial^2 u}{\partial x^2} + \frac{\partial^2 u}{\partial y^2} + \frac{\partial^2 u}{\partial z^2} = \frac{1}{v^2} \frac{\partial^2 u}{\partial t^2} \]  

(C.1)

where \( u(x, y, z, t) \) represents a scalar disturbance or motion. Substituting the vector differential operator:

\[ \nabla = \left( \frac{\partial}{\partial x}, \frac{\partial}{\partial y}, \frac{\partial}{\partial z} \right) \]  

We get,

\[ \nabla^2 u(x, t) = \frac{1}{v^2} \frac{\partial^2 u(x, t)}{\partial t^2} \]  

(C.3)

where \( x \) are the position coordinates. In an inhomogeneous medium, \( v \) would be a function of \( x \) so

\[ \nabla^2 u(x, t) = \frac{1}{v(x)^2} \frac{\partial^2 u(x, t)}{\partial t^2} \]  

(C.4)

Consider the plane wave solution:

\[ u(x, t) = A(x)e^{i\omega(T(x)-t)} \]  

(C.5)

where \( A(x) \) denotes the amplitude of the disturbance (displacement) and \( T(x) \) is the travel time as a function of position. Substituting this into the inhomogeneous wave equation we get:

\[ \nabla^2 [A(x)e^{i\omega(T(x)-t)}] = \frac{1}{v(x)^2} \frac{\partial^2}{\partial t^2} [A(x)e^{i\omega(T(x)-t)}] \]  

(C.6)

Using the product rule for differentiation and for simplicity letting \( A(x) = A \) and \( T(x) = T \):

\[ \nabla^2 [Ae^{i\omega(T-\tau)}] = \frac{1}{A^2} \frac{\partial^2}{\partial t^2} [Ae^{i\omega(T-\tau)}] \]
\[ \nabla u(x, t) = \nabla A e^{i w (T(x) - t)} + i w A \nabla T e^{i w (T - t)} \]  
(C.7)

\[ \nabla^2 u(x, t) = e^{i w (T - t)} [\nabla^2 A - w^2 A (\nabla T)^2 + i (2 w \nabla A \nabla T + w A \nabla^2 T)] \]  
(C.8)

\[ \frac{\partial u}{\partial t} = -i w A e^{i w (T - t)} \]  
(C.9)

\[ \frac{\partial^2 u}{\partial t^2} = -w^2 A e^{i w (T - t)} \]  
(C.10)

Separating the real and imaginary parts we get two equations:

\[ \nabla^2 A(x) - w^2 A(x) [\nabla T(x)]^2 = -\frac{w^2 A(x)}{v(x)^2} \]  
(C.11)

and

\[ 2w \nabla A(x) \nabla T(x) + w A(x) \nabla^2 T(x) = 0 \]  
(C.12)

Dividing the first equation by \( w^2 A(x) \)

\[ \frac{\nabla^2 A(x)}{w^2 A(x)} - [\nabla T(x)]^2 = -\frac{1}{v(x)^2} \]  
(C.13)

Considering the high frequency limit as \( w \to \infty \), the term on the left tends to 0 and

\[ [\nabla T(x)]^2 = \frac{1}{v(x)^2} \]  
(C.14)

Expanding the \( \nabla \) operator leaves us with the eikonal equation:

\[ \left( \frac{\partial T(x)}{\partial x} \right)^2 + \left( \frac{\partial T(x)}{\partial y} \right)^2 + \left( \frac{\partial T(x)}{\partial z} \right)^2 = \left( \frac{1}{v(x)} \right)^2 \]  
(C.15)

Where \( \frac{1}{v} \) is defined as slowness, \( T(x, y, z) \) is the travel time to a point \( (x, y, z) \).

We can solve the eikonal equation numerically to generate travel time tables which tell us the time from a starting point (usually a seismic station) to any point in the medium. This in turn allows us to calculate the spatial gradient of time anywhere in the medium, which we can then use...
to estimate ray paths (rays are perpendicular to isochrons, and hence follow the local travel time gradient).

Another way to think about the above equation is to recall that a wavefront represents an isochron, and the ray direction, which, is perpendicular to the wavefront, will travel in the direction of the gradient of time (maximum change of time) or $\nabla T(x)$. If we think of slowness as a vector, then it should be the case that:

$$\frac{\partial T(x)}{\partial x} = s_x$$  \hspace{1cm} (C.16)

and similarly for $y$ and $z$. Hence, we can see that the eikonal equation simply equates the magnitudes of different vectors that represent the same quantity.
D. ADDITIONAL IMAGES FROM TOMOGRAPHY
(CHAPTER 4)

Figure D.1A: Map-view of Vs model generated from dispersion measurements from SIEMBRA using a Vp/Vs ratio of 1.73 and the “all data” 1D model (Figure 4.1). The depth of the section is shown at the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively. The black circles represent the hypocenter locations within 10 km of a horizontal plane passing though that depth.
Figure D.1B: Same as D.1A without hypocenter locations.
Figure D.1C: W-E cross-sections of Vs model with the latitude of the section shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing though that latitude.
Figure D.1D: Same as D.1C without hypocenter locations.
Figure D.2A: Map-view of Vp for SIEMBRA joint inversion model with the depth of the section shown at the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively. The black circles represent the hypocenter locations within 10 km of a horizontal plane passing through that depth.
Figure D.2B: Same as D.2A but for Vs.
Figure D.2C: Similar to D.2A but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.2D: Same as D.2A without hypocenter locations.
Figure D.2E: Same as D.2B without hypocenter locations.
Figure D.2F: Same as D.2C without hypocenter locations.
Figure D.2G: Same as Figure 4.4 but without hypocenter locations.
Figure D.2H: Same as Figure 4.5 but without hypocenter locations.
Figure D.2I: Same as Figure 4.6 but without hypocenter locations.
Figure D.3A: Map-view of Vp for SIEMBRA-ESP-CSN “brute force” model with the depth of the section shown at the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively. The black circles represent the hypocenter locations within 10 km of a horizontal plane passing through that depth.
Figure D.3B: Same as D.3A but for Vs.
Figure D.3C: Similar to D.3A but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.3D: Same as D.3A without hypocenter locations.
Figure D.3E: Same as D.3B without hypocenter locations.
Figure D.3F: Same as D.3C without hypocenter locations.
Figure D.3G: Same as Figure 4.8 but without hypocenter locations and observations labeled.
Figure D.3H: Same as Figure 4.9 but without hypocenter locations.
Figure D.3i: Same as Figure 4.10 but without hypocenter locations.
Figure D.4A: Map-view of Vp for model created from the 131 events used to calculate the residuals from SIEMBRA-ESP events to the CSN stations. Depth of the section shown at the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively. The black circles represent the hypocenter locations within 10 km of a horizontal plane passing though that depth.
Figure D.4B: Same as D.4A but for Vs.
Figure D.4C: Similar to D.4A but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.4D: W-E cross-sections of Vp for model created from the 131 events used to calculate the residuals from SIEMBRA-ESP events to the CSN stations. The latitude of the section is shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.1 km/s and 1 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude.
Figure D.4E: Similar to D.4D but for $V_s$ and the thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively.
Figure D.4F: Similar to D.4D but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.5A: Map-view of Vp model created from the CSN dataset with the starting model in Figure D.3. The depth of the section shown at the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively. The black circles represent the hypocenter locations within 10 km of a horizontal plane passing though that depth.
Figure D.5B: Same as D.5A but for Vs.
Figure D.5C: Similar to D.5A but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.5D: W-E cross-sections of Vp model created from the CSN dataset with the starting model in Figure D.3. The latitude of the section is shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.1 km/s and 1 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing though that latitude.
Figure D.5E: Similar to D.5D but for Vs and the thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively.
Figure D.5F: Similar to D.5D but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.6A: Map-view of Vp for SIEMBRA-ESP-CSN “gentle” model with the depth of the section shown at the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively. The black circles represent the hypocenter locations within 10 km of a horizontal plane passing through that depth.
Figure D.6B: Same as D.6A but for Vs.
Figure D.6C: Similar to D.6A but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.6D: Same as D.6A without hypocenter locations.
Figure D.6E: Same as D.6B without hypocenter locations.
Figure D.6F: Same as D.6C without hypocenter locations.
Figure D.6G: Same as Figure 4.12 but without hypocenter locations and observations labeled.
Figure D.6H: Same as Figure 4.13 but without hypocenter locations.
Figure D.6I: Same as Figure 4.14 but without hypocenter locations.
Figure D.7: N-S cross-sections of the results of the checkerboard resolution test for the SIEMBRA-ESP-CSN “brute force” model. The longitude of the section is shown at the lower right corner of each panel. Colors are deviations from the background as indicated in the palette at the top of the figure. The contour intervals are 0.5%.
Figure D.8A: Map-view of Priestly/Ward Vs model with the depth of the section shown at the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively.
Figure D.8B: W-E cross-sections of Priestly/Ward Vs model with the latitude of the section shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude.
Figure D.9A: Map-view of Vs model generated from virtual data from the Priestly/Ward Vs model (Figure D.8) and ANT data from the CHILLAX and SIEMBRA networks. The depth of the section is shown at the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively. The black circles represent the hypocenter locations within 8 km of a horizontal plane passing though depth.
Figure D.9B: W-E cross-sections of Vs model generated from virtual data from the Priestly/Ward Vs model (Figure D.8) and ANT data from the CHILLAX and SIEMBRA networks. The latitude of the section is shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude.
Figure D.10: W-E cross-sections of the Vp/Vs model generated by inverting with a starting surface wave model scaled by the IASP91 Vp/Vs. The latitude of the section is shown at the lower right corner of each panel. Colors are ratio values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.01 and 0.05, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude. The blue background shows that the IASP91 Vp/Vs is too high for the region.
Figure D.11A: Map-view of Vp for model generated from the joint inversion of body waves and surface waves from the regional dataset using the starting surface wave model scaled with a Vp/Vs of 1.74. The depth of the section is shown at the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively. The black circles represent the hypocenter locations within 8 km of a horizontal plane passing though depth.
Figure D.11B: Same as D.11A but for Vs.
Figure D.11C: Similar to D.11A but for $V_p/V_s$, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.11D: Same as D.11A without hypocenter locations.
Figure D.11E: Same as D.11B without hypocenter locations.
Figure D.11F: Same as D.11C without hypocenter locations.
Figure D.11G: W-E cross-sections of the Vp model. The latitude of the section is shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.1 km/s and 1 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing though latitude.
Figure D.11H: Similar to D.11G but for Vs and the thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively.
Figure D.11I: Similar to D.11G but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.11J: Same as D.11G but without hypocenter locations.
Figure D.11K: Same as D.11H but without hypocenter locations.
Figure D.11L: Same as D.11I but without hypocenter locations.
Figure D.12A: Map-view of $V_p$ for final model from joint inversion using only body wave data after surface wave data was exhausted. The depth of the section is shown at the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively. The black circles represent the hypocenter locations within 8 km of a horizontal plane passing though that depth.
Figure D.12B: Same as D.12A but for Vs.
Figure D.12C: Similar to D.12A but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.12D: Same as D.12A without hypocenter locations.
Figure D.12E: Same as D.12B without hypocenter locations.
Figure D.12F: Same as D.12C without hypocenter locations.
Figure D.12G: Same as Figure 4.19 but without hypocenter locations and observations labeled.
Figure D.12H: Same as Figure 4.20 but without hypocenter locations.
Figure D.121: Same as Figure 4.21 but without hypocenter locations.
Figure D.13A: Map-view of Vp for model generated from regional body wave data using the starting model as a 1D model averaged from the Priestley/Ward starting model. The depth of the section is shown at the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively. The black circles represent the hypocenter locations within 8 km of a horizontal plane passing through depth.
Figure D.13B: Same as D.13A but for Vs.
Figure D.13C: Similar to D.13A but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.13D: Same as D.13A without hypocenter locations.
Figure D.13E: Same as D.13B without hypocenter locations.
Figure D.13F: Same as D.13C without hypocenter locations.
Figure D.13G: W-E cross-sections of the Vp model. The latitude of the section is shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.1 km/s and 1 km/s, respectively. The white circles represent the hypocenter locations within 10 km of a vertical plane passing though latitude.
Figure D.13H: Similar to D.13G but for Vs and the thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively.
Figure D.13I: Similar to D.13G but for Vp/Vs, colors are ratio values as indicated in the palette at the top of
the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.13J: Same as D.13G but without hypocenter locations.
Figure D.13K: Same as D.13H but without hypocenter locations.
Figure D.13L: Same as D.13I but without hypocenter locations.
Figure D.14A: Map-view of Vp for model generated from regional body wave data using the 1D model averaged from the SIEMBRA joint inversion as starting model. The depth of the section is shown at the lower right corner of each panel. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively. The black circles represent the hypocenter locations within 8 km of a horizontal plane passing though that depth.
Figure D.14B: Same as D.14A but for Vs.
Figure D.14C: Similar to D.14A but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.14D: Same as D.14A without hypocenter locations.
Figure D.14E: Same as D.14B without hypocenter locations.
Figure D.14F: Same as D.14C without hypocenter locations.
Figure D.14G: Same as Figure 4.23 but without hypocenter locations and observations labeled.
Figure D.14H: Same as Figure 4.24 but without hypocenter locations.
Figure D.14I: Same as Figure 4.25 but without hypocenter locations and observations labeled.
Figure D.15: Map-view of P-wave diagonals for the normal equations from the inversion using the Priestley/Ward 3D starting model. Colors are the diagonals as indicated in the palette at the top of the figure. The solid blue region represents sensitivities greater than 1000 which likely represent parameters that could be resolved in an inversion as the damper is 500. The black circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude.
Figure D.16A: Map-view of checker model with checkers with 5 x 5 grid points, or roughly 32–40 km cubes for Checkerboard Test using the data from the inversion with the Priestley/Ward 3D starting model. Colors are deviations from the background as indicated in the palette at the top of the figure. Red and Blue boxes represent regions perturbed by +5%, respectively. Yellow triangles are the seismic stations.
Figure D.16B: W-E cross-section of D.16A.

Figure D.16C: N-S cross-section of D.16A.
Figure D.16D: Map-view of checker model with checkers the second 3 x 3 grid points, or roughly 16-24 km cubes. Red and Blue boxes represent regions perturbed by +5%, respectively. Yellow triangles are the seismic stations.

Figure D.16E: W-E cross-section of D.16D.
Figure D.16F: N-S cross-section of D.16D.

Figure D.16G: W-E cross-sections of Figure 4.27. The latitude of the section is shown at the lower right corner of each panel. The contour intervals are 0.2%.
Figure D.16H: N-S section of Figure 4.27. The longitude of the section is shown at the lower right corner of each panel. The contour intervals are 0.5%.

Figure D.16I: N-S cross-section of Figure 4.28. The contour intervals are 0.5%.
Figure D.17A: Map-view of Vp from reconstruction test using the model from Figure D.13. The depth of the section is shown at the lower right corner of each panel. Left panel shows target model and right panel the recovery result. Colors are deviations from a one-dimensional background as indicated in the palette at the top of the figure. The thin and thick contour intervals are 1% and 10%, respectively. The black circles represent the hypocenter locations within 8 km of a horizontal plane passing though that depth.
Figure D.17B: Same as D.17A but forVs.
Figure D.17C: Similar to D.17A but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.18A: Map showing paths for the cross-sections (red lines) parallel to the JFR. Black circles represent hypocenter locations. Yellow triangles are seismic stations.
Figure D.18B: Cross-sections for Vp for model in Figure 4.19 with the depth of the section shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.1 km/s and 1 km/s, respectively. The black circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude.

Figure D.18C: Similar to D.18B but for Vs model in Figure 4.20 and the thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively.
Figure D.18D: Similar to D.18B but for Vp/Vs model in Figure 4.21, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.

Figure D.18E: Same as D.18B but for model in Figure 4.23.
Figure D.18F: Same as D.18C but for model in Figure 4.24.

Figure D.18G: Same as D.18D but for model in Figure 4.25.
Figure D.19A: Map showing paths for the cross-sections (red lines) perpendicular to the JFR for the models in Figures 4.19-4.21. Black circles represent hypocenter locations. Yellow triangles are seismic stations.
Figure D.19B: Vp with the depth of the section shown at the lower right corner of each panel. Colors are velocity values as indicated in the palette at the top of the figure. The thin and thick contour intervals are 0.1 km/s and 1 km/s, respectively. The black circles represent the hypocenter locations within 10 km of a vertical plane passing through latitude.
Figure D.19C: Same as D.19B but for Vs and the thin and thick contour intervals are 0.05 km/s and 0.5 km/s, respectively.
Figure D.19D: Similar to D.19B but for Vp/Vs, colors are ratio values as indicated in the palette at the top of the figure and the thin and thick contour intervals are 0.01 and 0.05, respectively.
Figure D.19E: P-wave diagonals for the normal equations. Colors are the diagonals as indicated in the palette at the top of the figure. The solid blue region represents sensitivities greater than 1000 which likely represent parameters that could be resolved in an inversion as the damper is 500. The black circles represent the hypocenter locations within 10 km of a vertical plane passing through the latitude.
Figure D.19F: Same as D.19E but for $V_p/V_s$. 