Four Brothers and a Waka: Investigating Lithospheric Accommodation of Shear and Convergence Underlying the South Island of New Zealand

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Four Brothers and a Waka:
Investigating Lithospheric Accommodation of Shear and
Convergence Underlying the South Island of New Zealand

by

Daniel W. Zietlow

A.B., Rollins College, 2010

A thesis submitted to the
Faculty of the Graduate School of the
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of the requirements for the degree of
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written by Daniel W. Zietlow
has been approved for the Department of Geological Sciences

Anne F. Sheehan

Peter H. Molnar

Date ______________

The final copy of this thesis has been examined by the signatories, and we find that both the content and the form meet acceptable presentation standards of scholarly work in the above mentioned discipline.
Zietlow, Daniel W. (Ph.D., Geophysics)

Four Brothers and a Waka:

Investigating Lithospheric Accommodation of Shear and Convergence Underlying the South Island of New Zealand

Thesis directed by Dr. Anne F. Sheehan

New seismic anisotropy and $P$- and $S$-wave mantle tomography models produced from passive source seismic data provide insight into how mantle lithosphere deforms due to lateral shear and oblique convergence across the South Island of New Zealand. The shear wave splitting measurements show a zone of mantle seismic anisotropy 100 – 200 km wide with orientations of the fast quasi-shear wave nearly parallel to relative plate motion between the Pacific and Australian plates. Anisotropy of increased obliquity to relative plate motion bookend this 100 – 200 km wide zone. A model representing distributed lithospheric deformation matches the observed orientation and amounts of shear wave splitting. Likewise, given certain conditions of grain size in the asthenosphere, localized lithospheric deformation with diffuse shear in the asthenosphere also matches the shear wave splitting measurements. $P$- and $S$-wave tomograms show high-speed structure under the northwestern South Island reaching 400 – 450 km deep. This is evidence for oblique westward subduction of Pacific lithosphere since 45 Ma, consistent with estimates from finite rotations of 850 – 1000 km of subduction thought to have occurred since this time. The high-speed zone reaches 200 - 300 km below the depths of the deepest intermediate depth earthquakes, suggesting that ~ 200 - 300 km of slab below them is required to produce sufficient weight to induce the intermediate depth seismicity. Low seismic speeds seen to 200 km along the east coast of the South Island coincide with regions of volcanism and thinned lithosphere. In the central South Island directly under regions of mountain building, high speeds are found to about 200 km deep. This suggests the lack of an unstable drip due to convergence across the South Island, though is consistent with accommodation of convergence via either lithospheric thickening or intracontinental subduction.
Dedication

Whakaihia ki te whānau me te hoa
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Chapter 1

Introduction

Regions of either lateral shear or convergence between two continental plates result in distinct surface features such as mountains, plateaus, or linear escarpments. What is less apparent is how the deeper solid Earth behaves. In shear zones, does mantle lithosphere deform continuously over a zone hundreds of kilometers wide, or is lithospheric deformation localized within a few kilometers around a fault, with more diffuse shear within the underlying asthenosphere? In the case of convergent plates, how has the lithosphere been altered due to thickening or subduction processes? In this dissertation, I investigate how mantle lithosphere underlying the South Island of New Zealand behaves in both a region of lateral shear and a convergence zone via studies of seismic anisotropy and body wave tomography.

Prior investigations of New Zealand mantle seismic anisotropy show that the underlying mantle is highly anisotropic [Duclos et al., 2005; Klosko et al., 1999]. Likewise, previous studies of convergence occurring across the South Island image a nearly vertical, high-speed anomaly to about 200 km depth under the island, typical of regions undergoing convergence [Kohler and Eberhart-Phillips, 2002; Stern et al., 2000]. The narrow seismic aperture used in such studies, given the approximately 200 km width of the South Island, resulted in limited constraints on the extent of such subsurface features. My work improves on these previous studies by using a recent deployment of ocean bottom seismometers (OBSs) by the University of Colorado. This extends the seismic array aperture to about 900 km, over four times the width of the South Island, and affords us unprecedented depth resolution in the region.
The following chapters in this dissertation are stand-alone scientific articles, though Chapters 3 and 4 are heavily related. The appendices document material that enhances the main chapters, such as all the shear wave splitting measurements or travel-time measurements. In Chapter 2, I investigate upper mantle seismic anisotropy underneath the South Island through a shear wave splitting study \cite{Zietlow et al., 2014}. I find a zone of anisotropy approximately 100 - 200 km wide that is bounded by regions of distinctly different patterns of seismic anisotropy. This observed pattern of anisotropy is consistent with a model of widespread deformation in the mantle lithosphere; however, I also show that given certain mantle conditions, diffuse shear in the asthenosphere could also produce a similar pattern of anisotropy to that observed.

In Chapters 3 and 4, I investigate seismic heterogeneity in the upper mantle under the South Island through teleseismic $P$- and $S$-wave tomography \cite{Zietlow et al., 2016a,b}. In both studies, I find high-speed material under the northwest South Island and just offshore the west coast to depths between 400 and 450 km, inferred as Pacific lithosphere subducting since about 45 Ma. The high-speed zone reaches 200 - 300 km below the depths of the deepest intermediate depth earthquakes (subcrustal to $\sim$ 200 km), suggesting that about 200 - 300 km of slab below the earthquakes is required to produce sufficient weight to induce the intermediate depth seismicity. Low seismic wave-speed mantle along the east coast coincides with regions of volcanism and thinned lithosphere. Directly under the Southern Alps, the major mountain chain that resulted from oblique convergence across the South Island, high speeds are found to about 200 km depth. This suggests the lack of an unstable drip due to convergence across the South Island, though is consistent with accommodation of convergence via either lithospheric thickening or intracontinental subduction.

1.1 Four Brothers and a Waka

I derived the dissertation title “Four Brothers and a Waka” from a creation story \cite{Revington, 2012; Royal, 2012} told by the native people of New Zealand, the Māori. In the Māori tradition, the world began with the union of Papatūānuku (Mother Earth) and Rangimī (the Sky Father; also called Raki in some South Island Māori dialects) in the darkness known as Te Pō. Papatūānuku and
Raki had numerous children, though these children grew restless living in the darkness between the embrace of their parents. To alleviate their frustration, the children proposed to separate their parents by pushing Raki high above towards the heavens, leaving Papatūānuku below. The separation of Papatūānuku and Raki brought the world of light, or Te Ao Mārama.

According to the Ngāi Tahu, some time after the separation of Papatūānuku and Raki, the eldest son of the Sky Father, Aoraki, and his three brothers took a waka (canoe) down from the heavens to visit Papatūānuku. On their journey home, the waka overturned and the brothers climbed atop it for safety. As time passed, Aoraki and his brothers turned to stone where they remain today as the Kā Tiritiri o te Moana - the Southern Alps. Thus, the four brothers represent the Southern Alps and the waka represents the South Island of New Zealand.
New shear wave splitting measurements made from stations on and offshore the South Island of New Zealand show a zone of anisotropy 100 – 200 km wide. Measurements in central South Island and up to approximately 100 km offshore from the west coast yield orientations of the fast quasi-shear wave nearly parallel to relative plate motion, with increased obliquity to this orientation observed farther from shore. On the eastern side of the island, fast orientations rotate counterclockwise to become nearly perpendicular to the orientation of relative plate motion approximately 200 km off the east coast. Uniform delay times between the fast and slow quasi-shear waves of nearly 2.0 s onshore continue to stations approximately 100 km off the west coast, after which they decrease to \( \sim 1 \) s at 200 km. Stations more than \( \sim 300 \) km from the west coast show little to no splitting. East coast stations have delay times around one second. Simple strain fields calculated from a thin-viscous sheet model (representing distributed lithospheric deformation) with strain-rates decreasing exponentially to both the northwest and southeast with e-folding dimensions of 25 – 35 km (approximately 75% of the deformation within a zone 100-140 km wide) match orientations and amounts of observed splitting. A model of deformation localized in the lithosphere and then spreading out in the asthenosphere also yields predictions consistent with observed splitting if at depths of 100 – 130 km below the lithosphere, typical grain sizes are \( \sim 6 – 7 \) mm.
2.1 Introduction

Measurements of seismic anisotropy arguably offer the best way to constrain deformation deep in the Earth's subsurface [e.g., Silver, 1996]. With such measurements, it becomes possible to describe the distribution of strain in the mantle lithosphere (used here to include the relatively strong layer at temperatures less than around 1200°C), as well as the underlying asthenosphere, in active tectonic regions such as strike-slip zones. Specifically, does the lithosphere deform continuously over a broad region on the order of hundreds of kilometers, or is lithospheric deformation localized within a few kilometers around a fault that penetrates the lithosphere, with more diffuse shear spreading out within the underlying asthenosphere?

The South Island of New Zealand offers an ideal location to investigate subsurface deformation in a strike-slip zone. The tectonic history of the region is known and relatively simple (Figure 2.1). Plate reconstructions indicate ~850 km of displacement between the plates over the past 45 Ma [Cande and Stock, 2004; Sutherland, 1999]. The Alpine fault, a predominately dextral strike-slip fault running the length of South Island and the major tectonic feature of the central portion of South Island, has accommodated approximately 460 km of this displacement since 45 Ma [e.g., Sutherland et al., 2000]. Over this time period, the relative motion between the Australian and Pacific plates changed little, though a component of convergence initiated around 11 Ma [Cande and Stock, 2004]. This convergence resulted in approximately 100 km of shortening across the South Island [Walcott, 1998] and the creation of the Southern Alps. Today, geodetic measurements show that the present plate motion is resolved into approximately 36 - 39 mm/yr parallel and between 9 - 12 mm/yr perpendicular to the Southern Alps [Beavan et al., 1999].

Within the South Island, the style of deformation varies from north to south. Northeast of the South Island, westward subduction of the Pacific plate beneath the Australian plate occurs at the Hikurangi Trough, becoming increasingly oblique toward the south. This oblique convergence continues under the northern portion of the South Island beneath the Marlborough Fault Zone. Here, four major strike-slip faults accommodate much of the motion: the Wairau, the Awatere, the
Figure 2.1: Map of New Zealand and surrounding region showing the location of major tectonic features. Also shown is a summary of previous teleseismic SKS splitting measurements made by Klosko et al. [1999] (blue bars) and by Duclos et al. [2005] (red bars). Black crosses indicate null measurements, with the orientation showing the two possible fast orientations. Lines through stations show orientations of polarization of the faster quasi-shear waves. Lengths of lines are proportional to delay times. Inverted yellow triangles represent the locations of ocean bottom seismometers deployed in the MOANA experiment and green triangles represent the stations of the New Zealand National Seismograph Network used in this study. Thick black arrows show current absolute plate motion of the Pacific and Australian plates in a hotspot reference frame with no net rotation [Gripp and Gordon, 2002]. The thin black arrow shows motion of the Pacific plate relative to the Australian plate averaged over 20 Ma.
Clarence, and the Hope faults [e.g., Wallace et al., 2007]. At the southern end of the Marlborough Fault Zone, slip transfers to the Alpine fault, which accommodates 70-75% of plate motion between the Pacific and Australian plates [Norris and Cooper, 2001]. The Alpine fault extends southward into Fiordland where the Australian plate subducts beneath the Pacific plate at the Puysegur Trench and beneath Fiordland.

Previous studies of SKS phases found that the mantle underlying the South Island of New Zealand is highly anisotropic (Figure 2.1) [Duclos et al., 2005; Klosko et al., 1999]. Central portions of the South Island, as well as its northern end, exhibit orientations of fast quasi-shear waves nearly parallel to the trend of the Southern Alps. These regions also exhibit large delay times between fast and slow quasi-shear waves, ranging between 1 and 2 s. The southernmost regions of the island show orientations of the fast quasi-shear wave rotated approximately 20° counterclockwise from those measurements in the northern portion of the island, though delay times remain between 1 and 2 s. Shear wave splitting measurements made on both the Chatham Islands and Macquarie Island (both lying roughly 1000 km from the South Island) exhibit different orientations and amounts of anisotropy from what is observed on the South Island. All measurements made on the Chatham Islands yield nulls, suggesting isotropy. Those from Macquarie Island showed a NW fast axis orientation with a delay time of over one second [Klosko et al., 1999]. These results suggest that the observed anisotropy in the mantle under the South Island is not due either to older structures found throughout the South Island, as suggested by Furlong [2007], or to absolute plate motion of the Pacific or Australian plates [Klosko et al., 1999], as is observed at the East Pacific Rise [Wolfe and Solomon, 1998]. Studies of shear wave splitting of S arrivals from local earthquakes in the subduction zones yield much smaller delay times than those of SKS phases, but it has been difficult to determine whether the smaller values are caused by low anisotropy in the mantle lithosphere or by frequency-dependent splitting [Audoine et al., 2000; Karalliyadda and Savage, 2013].

Two explanations for the observed pattern of anisotropy have been proposed (Figure 2.2). In one, anisotropy is caused by shear within the mantle lithosphere across a broad zone that spans the
width of the South Island, if not wider \cite{Little2002, Molnar1999, Moore2002}, an interpretation invoked in other strike-slip regions \cite[e.g.,][]{Vauchez1991}. GPS studies also suggest that, at least for central portions of the South Island, simple shear likely occurs over a zone approximately 200 km wide \cite[e.g.,][]{Houlié2012}. Others have inferred that localized deformation in the shallow lithosphere around the Alpine fault continues through the lithosphere into the asthenosphere so that the source of the anisotropy extends to depths of 250 km or more \cite{Karalliyadda2013, Klosko1999}. Both interpretations rely on data recorded only by stations deployed across the South Island of New Zealand. As the maximum width of the South Island is approximately 200 km, this limits the lateral extent of the array aperture. Thus, to test these proposed mechanisms of deformation, we devised an experiment that expanded the array aperture.

### 2.2 The MOANA Experiment

The Marine Observations of Anisotropy Near Aotearoa (MOANA) experiment included a one-year deployment of ocean bottom seismometers (OBSs) installed off both coasts of the South Island of New Zealand from late January 2009 to early February 2010 (Figure 2.1). Deployed with approximately 100 km spacing, this array of OBSs increased the aperture of study to five times the width of the South Island. Stations NZ01 – 04 rested on oceanic crust, with the others lying on the continental shelf surrounding New Zealand. All instruments recorded continuously at a sample rate of 50 Hz. These OBSs complemented the existing land stations of the New Zealand National Seismograph Network maintained by GeoNet. All but one of the OBSs (station NZ14) were equipped with Trillium 240 seismometers with the final OBS equipped with a Trillium 40 seismometer. The former are broadband instruments with flat frequency responses between 240 seconds and 35 Hz. The latter is an intermediate period instrument with a long period corner at 40 seconds. Although 30 instruments were deployed, one did not yield usable seismograms (station NZ01), and one was never recovered (station NZ17).

We used Rayleigh wave polarizations to orient the horizontal components of the OBSs \cite{Stach-
Figure 2.2: Cartoons showing the proposed mechanisms of deformation, traversing the South Island perpendicular to the Alpine fault. Light areas represent regions deforming via dislocation creep, thus producing a lattice preferred orientation. A) A model with a weak lower crust and strong mantle lithosphere in which dislocation creep occurs as in a thin-viscous sheet. B) A model with a strong lithosphere such that localized strain and lattice preferred orientation develop in the asthenosphere in a region fanning out from the base of the lithosphere.
This technique exploits the elliptical particle motion of Rayleigh waves and the fact that this motion is theoretically observed only on the vertical and radial components of ground motion. Since linearity is easier to assess than ellipticity, Stachnik et al. [2012] determined instrument orientation by cross correlating the vertical component with the Hilbert transformed radial component. The polarization analysis included first predicting the arrival time of the Rayleigh wave with a group speed of 4.0 km/s and then applying a bandpass filter from 0.02 – 0.04 Hz to the waveform in the window 20 s before and 600 s after this phase arrival. The Hilbert transformed radial components for a range of backazimuths (0 – 360°) were cross-correlated with the vertical components. The sensor orientation is assumed to correspond to the azimuth of maximum correlation. For the OBSs associated with the MOANA experiment, between 10 and 31 events were used to determine the orientations of the horizontal components.

2.3 Mantle Anisotropy and Shear Wave Splitting Methods

Within a single anisotropic layer, a traveling shear wave splits into two, orthogonal quasi-shear waves that propagate through the medium with different wave speeds, resulting in two distinct arrivals on a seismogram [Babuška and Cara, 1991, and references therein]. Shear wave splitting analyses yield measurements of both the delay time between the fast and slow quasi-shear waves (δt) and the orientation (ϕ) of the faster quasi-shear wave. These parameters provide information on the orientation and degree of anisotropy within the medium of interest.

In the upper mantle, anisotropy is dominantly a result of lattice preferred orientation (LPO), or the preferential alignment of crystalline structure in response to finite strain [e.g., Savage, 1999; Silver, 1996]. Anisotropy is commonly quantified as a percentage: 100(v_{max} – v_{min})/v_{avg}, where v_{max} is the speed of the fast quasi-shear wave, v_{min} is that of the slow quasi-shear wave, and v_{avg} is the average of the two. It is typically thought that olivine is the principal mineral involved in upper mantle anisotropy, for which percent anisotropy can reach 18% for single crystals, though crystals not aligned in the same direction yield a smaller overall percent anisotropy. Other minerals such as orthopyroxene also contribute [e.g., Kumazawa, 1969; Long and Becker, 2010; Ribe, 1992; Skemer
Deformation in the upper mantle occurs via either diffusion creep or dislocation creep [e.g., Karato and Wu, 1993]. Diffusion creep involves the diffusion of atoms along grain boundaries, whereas dislocation creep involves the migration of dislocations in the crystal lattice [Turcotte and Schubert, 2002]. In general, only dislocation creep produces the LPO and microstructure consistent with observed anisotropy in the upper mantle [e.g., Savage, 1999].

Laboratory experiments suggest that predicted lattice preferred orientation patterns in olivine aggregates depend on the finite strain [e.g., Karato et al., 2008]. At low strain and/or when dynamic recrystallization is limited, the anisotropy of a deformed aggregate can be expressed relative to the principal axes of finite strain. During progressive simple shear (e.g., the Alpine fault), the olivine a-axes initially orient approximately 45° from the plane of shear and rotate progressively toward that plane as strain accumulates; at sufficiently high strains (>150% in lab samples deformed at 1300°C) and/or cases where dynamic recrystallization is prevalent, the olivine a-axes rotate to lie within the plane of shear [e.g., Zhang and Karato, 1995]. With moderate hydrogen (“water”) content (∼500 ppm H/Si), such as might be expected in the asthenosphere or within the mantle wedge of a subduction zone, a different pattern of LPO (E-type) is observed [e.g., Karato et al., 2008; Mehl et al., 2003]. Examination of fabrics in high temperature shear zones (1000-1200°C) demonstrates that the details of how LPO evolves with progressive simple shear in the Earth match those determined in the lab and that the olivine water content under which the fabric transitions occur is consistent with measurements of water contents preserved in the natural samples [e.g., Skemer et al., 2010, 2013; Warren et al., 2008]. The agreement between the laboratory and natural observations justifies interpretations of mantle anisotropy guided by the details of the laboratory analyses. Finally, B-type and C-type LPOs observed in lab samples deformed at higher water contents, pressures, and stresses than those required to promote E-type fabric can significantly influence the interpretation of anisotropy [e.g., Karato et al., 2008].

To constrain the fast-axis orientations and delay times, we employ the method outlined by Silver and Chan [1991] as implemented in SplitLab [Wüstefeld et al., 2008]. This is a grid-search based method that determines the best fitting splitting parameters (φ and δt) by minimizing energy
on the transverse component after removing effects due to shear wave splitting. We also adopt the multiple event splitting method of Wolfe and Silver [1998] in which the resulting energy contours of individual measurements are stacked. This helps reduce effects of acknowledged shortcomings of Silver and Chan’s method in which large amounts of noise and/or small degrees of birefringence can yield incorrect values of $\varphi$ and $\delta t$ [Levin et al., 2007; Monteiller and Chevrot, 2010; Restivo and Helffrich, 1999]. In such instances, the measured value of $\varphi$ lies parallel to the backazimuth of the earthquake used in the analysis, with the result that Silver and Chan’s method can yield unrealistically large values of $\delta t$. Such measurements are typically classified as null measurements. Because the ocean is inherently a noisy environment, many splitting measurements exhibited these characteristics, which makes it difficult to distinguish null measurements from cases with low signal-to-noise ratios (defined as the ratio of the maximum amplitude on the radial component to the standard deviation of the transverse component in the analysis window). With the stacking method of Wolfe and Silver [1998], however, we were able to use such measurements to distinguish noisy waveforms and/or small delay times from nodal results.

### 2.4 Results

We examine all $SKS$, $SKKS$, and $PKS$ arrivals from earthquakes with magnitude $M_w \geq 5.0$ recorded across the MOANA array between early 2009 and early 2010, as well as by the New Zealand National Seismograph Network between 2003 and April 2012, which included twenty-eight stations on the South Island, Stewart Island, and the Chatham Islands. Since the last major study of $SKS$ phases in New Zealand [Duclos et al., 2005], several new stations were added to the network, and we could also use more than eight years of new waveform data from those previously characterized stations. Although we analyzed earthquakes of magnitude $M_w \geq 5.0$ occurring at epicentral distances between $85^\circ$ and $130^\circ$ from the South Island, we found all usable events to be of magnitude $M_w > 6.0$ with epicentral distances between $90^\circ$ and $120^\circ$. All the usable phase arrivals were $SKS$, with one usable $SKKS$ arrival. The majority of the events occurred in the Aleutian, Kuril, and Peru-Chile subduction zones, resulting in non-uniform backazimuthal coverage, especially on
the OBSs (Figure 2.3 and Table 2.1). This non-uniform backazimuthal coverage on the OBSs means that we cannot resolve any possible two-layer anisotropy [Silver and Savage, 1994].

The waveforms were bandpass filtered between 0.03 Hz and 0.08 Hz with a four pole Butterworth filter. Microseismic noise determined the high-frequency end of this filter. Since the OBSs rested atop sediment, the instruments were subject to long-period noise, which determined the low-frequency end of the filter. Furthermore, we ignore waveforms with signal-to-noise ratios less than 5, consistent with what Wüstefeld and Bokelmann [2007] found as the lowest limit for automatic detection of null measurements. The analysis window included 20 – 40 seconds of pre-event noise, with the end of the window typically limited by predicted arrivals of later seismic phases.

Individual measurements before stacking varied in quality from station to station (Figure 2.4). Typically on the OBSs, SKS arrivals were clear on the filtered waveforms, but relatively high noise levels affected some measurements (Figure 2.4b and c). This is evident by the absence of linearized particle motions as well as the large 95% confidence intervals calculated for splitting measurements at such stations. Some measurements on OBSs, however, were comparably precise to those made on land stations. These waveforms (compare Figure 2.4d and e) have high signal-to-noise ratios, linear particle motions, and tightly constrained splitting parameters.

Table 2.1: List of earthquakes used in this study. Starred events recorded on an OBS.

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Figure 2.3: Locations of earthquakes used in this study. All events shown yielded at least one usable measurement. Circles represent events used to constrain splitting parameters only on land stations. Diamonds represent events recorded on both land and ocean bottom seismometers.
Figure 2.4: Examples of splitting measurements showing (first column) the filtered waveforms with analysis windows depicted by grey shading, (second column) the waveforms corrected first for delay time and then for (third column) both delay time and orientation of the fast quasi-shear wave, the particle motion before (dashed blue line) and after (solid red line) the (fourth column) effects due to splitting are removed, and (fifth column) resulting contours of transverse energy. The filled grey contour represents the 95% confidence region of $\delta t-\phi$ space. The dashed blue lines in the seismogram in the left column are the radial components and the solid red the transverse. (a) Event 2010010 recorded at NZ08. (b) Event 2009097 recorded at NZ12. (c) Event 2009108 recorded at NZ15. (d) Event 2009222 recorded at NZ30. (e) Event 2004320 recorded on WVZ. Event details are in Table 2.1.
Some stations yielded null measurements (Figure 2.4a). In such cases, an error ellipse contour elongated in the $\delta t$ direction for $\phi$ aligned parallel or perpendicular to the backazimuth and low energy on the raw transverse component suggests either that there may be no anisotropy underlying the station or alternatively that the azimuth of the fast orientation is parallel or perpendicular to the backazimuth of the event, but with $\delta t$ poorly constrained. This is evident in the individual energy contours from station NZ09 seen in the left column of Figure 2.5. Individual contours in Figure 2.5a-d exhibit one of the shortcomings of Silver and Chan’s method, discussed in the previous section. For each SKS phase, noise and/or small birefringence at this station forces the predicted orientation of the fast quasi-shear wave along the backazimuth of the earthquake, with the predicted delay time approaching the maximum allowed time in the analysis (in this case, 4.0 s). When the contoured energy distributions are stacked, however, the resulting sum shows that these measurements are consistent with one another and yield precise estimates of $\phi$ and $\delta t$ (Figure 2.5e). Likewise, the individual contours for earthquakes with different backazimuths in Figure 2.5f-h from station NZ13 show clear anisotropy; none reveals a tightly constrained orientation or delay time, but the resulting stack (Figure 2.5j) does constrain $\phi$ and $\delta t$.

We also analyzed the individual seismograms for some measurements both before and after correcting for anisotropy with the average $\phi$ and $\delta t$ calculated from the stacked measurements at that station (Figure 2.6). The averages for $\phi$ and $\delta t$ obtained from the stack (Figure 2.7 and Table 2.2) are consistent with what the individual measurements allow, suggesting that a one-layer model is sufficient to explain the data.
Figure 2.5: (a-d) Individual energy contours from individual splitting measurements made at NZ09. (e) Resulting stack of contours shown in Figure 2.5a-d. (f-i) Individual energy contours from NZ13. (j) Stack of individual energy contours from NZ13 shown in Figure 2.5f-i. The grey contour is the 95% confidence interval. The intersection of the black lines defines the best fitting splitting parameters.
Figure 2.6: (top) Superposition of the fast ($\phi$ direction) and slow ($\phi + 90^\circ$) components and (bottom) the uncorrected and corrected particle motions for (left) event 2009321 and (right) event 2009344 at NZ09.
Figure 2.7: Summary of SKS splitting measurements made in this study. Red bars indicate orientations of the polarizations of the faster quasi-shear wave, and black crosses indicate null measurements. Grey wedges define uncertainties in measured orientations of fast quasi-shear waves (uncertainties are also found in Table 2.2). Black dots mark stations at which no usable measurements were made. Thick black arrows show current absolute plate motion of the Pacific and Australian plates in a hotspot reference frame with no net rotation [Gripp and Gordon, 2002]. The thin black arrow shows plate motion relative to the Australian plate averaged over 20 Ma. SKS splitting measurements from Klosko et al. [1999] and Duclos et al. [2005] plotted in light grey.
Table 2.2: Summary of splitting results by station, utilizing the *Wolfe and Silver* [1998] stacking method. Starred stations are those that lie in the central portion of the South Island within 200 km of the center of the shear zone. We used only these stations in the error calculations found in Table 3 and in Figures 2.8 to 2.10.

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Table 2.2 – continued from previous page

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<th>$\phi$ (°)</th>
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Stations NZ08, 22, and 23 yielded null measurements, though these are constrained by a single event with a backazimuth parallel to the predicted orientation of the fast quasi-shear wave. Some stations, including stations NZ06, 07, 14, 19, and 21, yielded no usable measurements due to high levels of noise in the waveforms. We should note that in analysis of DCZ, the north and east components of the instrument were switched, consistent with phase arrivals in the seismic record.

In central portions of the South Island and within close proximity to the Southern Alps (a zone roughly 100 – 200 km wide), the orientations of fast polarization directions are nearly parallel to relative plate motion (roughly 240°, as averaged over the past 20 Myr in a reference frame fixed to the Australian plate [Cande and Stock, 2004]). Delay times are between about 1.5 – 2.0 s (Figure 2.7). In the extreme north of the island, including the Marlborough Fault Zone, orientations...
of the fast quasi-shear wave remain nearly parallel to relative plate motion with delay times again upwards of 2.0 s; however, the mantle under some of these stations may have interacted with, or even include, the subducting Pacific plate. In the southern portion of the island, especially at stations > 100 km southeast of the Alpine fault, fast orientations are more oblique to relative plate motion. In south central portions of the South Island, including DCZ, EAZ, MLZ, PYZ and WHZ (Figure 2.7) fast orientations are rotated approximately 40° counterclockwise from plate motion. This obliquity increases to nearly 60 – 70° at stations near the southeastern coast of the South Island (ODZ, OPZ, SYZ, and TUZ). Delay times observed at these stations remain around 1.0 s. On both Stewart Island and the Chatham Islands, very little splitting is observed; delay times are only 0.4 s. The fast orientation seen on Stewart Island is consistent with the obliquity seen elsewhere on the southern portions of the South Island. Individual measurements made on the Chatham Islands yielded null or near null results, consistent with previous observations [Klosko et al., 1999]. Stacking these results gives an observed fast orientation nearly E-W, but the calculated delay time is negligibly small (0.4 s), and the number of events originating in the Aleutians and Kuril Trench may bias this result.

Within about 100 km offshore of the west coast, most shear wave splitting measurements yielded fast axis orientations parallel (stations NZ11 and NZ15) or approximately parallel (station NZ12) to the orientation of relative plate motion. As seen on land for those stations within 100 km of the Alpine fault, delay times observed at the OBSs range from just over one second up to 2.4 ± 0.7 s. At stations deployed around 200 km offshore the west coast (i.e., stations NZ13, 16, and 18), delay times remain relatively large in magnitude around 1.5 s. Unlike those OBSs immediately off the coast, these stations yielded fast orientations rotated between 40 – 50° counterclockwise from relative plate motion. Beyond 300 km from the west coast, we observed a region with little to no resolvable anisotropy. The stations with minor resolvable anisotropy in this region (stations NZ05 and NZ09) show fast axes rotated almost 90° to that of relative plate motion with delay times around 0.5 s. The only useable measurement at station NZ08 yielded a null result.

The three stations deployed off the continental shelf and on oceanic lithosphere (NZ02 –
also exhibit a very different sense of anisotropy, with fast axis orientations trending almost due north and delay times around 0.7 s. Offshore of the east coast, fast directions are consistently nearly perpendicular to relative plate motion between the Pacific and Australian plates. Delay times calculated at these stations also remained consistent across the region, ranging between about 0.5 and 1.0 s.

### 2.5 Relationship of Anisotropy to Mantle Deformation

The shear wave splitting results clearly indicate a finite width over which the observed anisotropy could be due to simple shear between two plates. Off the east coast, the orientations of the fast quasi-shear waves are nearly aligned with current absolute plate motion of the Pacific plate calculated from HS3-NUVEL1A with no net rotation (Figure 2.7) [Gripp and Gordon, 2002]. Likewise, the fast orientations measured at the OBSs deployed off the west coast on oceanic lithosphere appear oriented parallel to absolute plate motion of the Australian plate. Offshore the west coast at stations deployed on the continental shelf, there is a decrease in the magnitude of the delay times as well as increase in the counterclockwise rotation of the fast axis orientations with increasing distance from the South Island. Furthermore, at distances 300 km and greater from the west coast, there appears to be little observable splitting. Thus, areas of anisotropy generated by different tectonic processes appear to bookend the region apparently affected by simple shear associated with relative plate motion. The symmetry of the anisotropy on either side of the Alpine fault in central regions of the South Island is also interesting, since this implies that strain is relatively uniform on either side.

The general trend of decreasing delay time observed off the west coast from the northeast towards the southwest (i.e., NZ11, 12, 14, 15) is mimicked on land, with delay times of 2.0 s observed in the extreme northwest of the island (QRZ) versus delay times of closer to 1.0 s observed in the southwest (i.e., JCZ). This feature may result from the more than 800 km of motion known to have occurred between the Pacific and Australian plates over the past 45 Ma, with northwestern regions seeing more relative displacement compared to those farther south.
To discriminate between the two currently proposed mechanisms of deformation shown in Figure 2.2, we compare these shear wave splitting results to predictions made from both a thin-viscous sheet model (mimicking distributed lithospheric deformation shown in Figure 2.2a) and a model in which strain within the lithosphere is localized and anisotropy is produced by distributed strain in the asthenosphere (Figure 2.2b).

### 2.5.1 Lithospheric Deformation: Thin-Viscous Sheet Model

For a thin-viscous sheet, we follow England et al. [1985]. In such a model, horizontal components of velocity are assumed to be independent of depth, a reasonable assumption when the tractions on the base of the sheet are small compared to stresses within the sheet. They assume that deformation of the sheet obeys a constitutive law of the following form: \( \dot{\varepsilon} \sim \tau^n \), where \( \dot{\varepsilon} \) is strain rate, \( \tau \) is deviatoric stress, and \( n \) is the exponent of the non-Newtonian rheological model. From England et al. [1985], the approximate velocity field as a function of distance from the deforming boundary decays as

\[
\bar{u}(x, y) \equiv U_0 \sin \left( \frac{2\pi x}{\lambda} \right) \exp \left( -\frac{4\sqrt{n}y}{\lambda} \right),
\]

(2.1)

where \( U_0 \) is the component of velocity parallel to the deforming boundary (18 mm/yr in this study, or half the total observed relative plate motion [Beavan et al., 1999]), \( x \) is the parallel distance along the boundary, \( \lambda \) is the wavelength of deformation (twice the length of a deforming boundary of finite length), and \( y \) is the perpendicular distance from the boundary. In this study, we assume that \( n = 3 \). Equation (2.1) is most accurate near the maxima and minima of the horizontal velocity (i.e., when \( x \) is close to \( \lambda/4 \)) [England et al., 1985].

With exponentially decreasing velocity and strain, approximately 75% of relative plate motion occurs where \( y < 2y_0 \), where \( y_0 \) is the e-folding width (the value of \( y \) where the velocity decays to 1/e of its original value) of Equation (2.1):

\[
y_0 = \frac{\lambda}{(4\sqrt{n}\pi)}.
\]

(2.2)

We note that England et al. [1985] formulated Equation (2.1) for a semi-infinite region in the x-y...
plane for \( y > 0 \), and therefore for only one side of a deforming boundary. As we are concerned with predicted motion on both sides of a deforming boundary, we assume symmetry in the velocity field. Thus, we define the total width of deformation due to the deforming boundary as \( 4y_0 \), again within which approximately 75% of the relative displacement of the plates has been accommodated. For a fixed exponent of the non-Newtonian rheological model \( n \), Equation (2.2) indicates that for a given value of \( y_0 \), there is a corresponding value of \( \lambda \), the along-strike wavelength of deformation. Since \( \lambda \) is twice the length of the deforming boundary, we can also predict the length of the deforming boundary from \( y_0 \), or from an inferred width of deformation.

### 2.5.2 Asthenospheric Deformation

For the case of shear in the asthenosphere, strain rate decreases with depth in a fan-like shape beneath the lithosphere-asthenosphere boundary (Figure 2.2b). Where the strain rate becomes sufficiently small, dislocation creep is no longer the dominant mechanism of deformation, and lattice preferred orientation does not develop. Assuming that strain from absolute plate motion is negligible compared to the strain generated in the fan-like shape beneath the fault, and assuming a constant viscosity in the asthenosphere, we adopt a toroidal velocity field of the following form, in which the only non-zero component is that parallel to what is assumed to be a fixed boundary [Savage et al., 2004]:

\[
\dot{u}_y = U_0 \left( 1 - \frac{2\theta}{\pi} \right),
\]

where \( U_0 \) again is the component of velocity parallel to a strike-slip boundary \( (U_0 = 18 \text{ mm/yr}) \), \( \theta = \arcsin(z/r) \), \( r = (y^2 + z^2)^{1/2} \), \( z = \text{depth} \), and \( y = \text{perpendicular distance from the fault} \). The only non-zero component of the strain rate tensor for this velocity field is [Savage et al., 2004]:

\[
\dot{\varepsilon} = \pm \frac{U_0}{\pi r}.
\]

To determine whether or not a transition from dislocation to diffusion creep is feasible at a specified depth given the strain rates predicted from Equation (2.4), we use the synthesis of laboratory experiments given by Hirth and Kohlstedt [2003]. The strain rate of minerals in a
steady state is

\[ \dot{\varepsilon} = A\sigma^m d^{-p} C_{OH}^r \exp \left[ -\frac{(E^* + PV^*)}{RT} \right], \]  

(2.5)

where \( \dot{\varepsilon} \) is the strain rate, \( A \) is a pre-exponential factor, \( \sigma \) is shear stress, \( m \) is the stress exponent, \( p \) is the grain size exponent, \( d \) is grain size, \( C_{OH} \) is water content, \( r \) is the water content exponent, \( E^* \) is the activation energy, \( P \) is pressure, \( V^* \) is the activation volume, \( R \) is the gas constant, and \( T \) is temperature in Kelvin. In a dislocation creep regime at constant OH concentrations (1000 H/10\textsuperscript{6}Si), \( A = 90 \) (for stress in MPa, \( C_{OH} \) in H/10\textsuperscript{6}Si, and grain size in \( \mu \)m), \( m = 3.5 \pm 0.3 \), \( p = 0 \), \( r = 1.2 \), \( E^* = 480 \pm 40 \) kJ/mol, and \( V^* = 11\times10^{-6} \) m\textsuperscript{3}/mol. In a diffusion creep regime at constant OH concentrations (again, 1000 H/10\textsuperscript{6}Si), \( A = 1.0\times10^6 \) (again, for stress in MPa, \( C_{OH} \) in H/10\textsuperscript{6}Si, and grain size in \( \mu \)m), \( m = 1 \), \( p = 3 \), \( r = 1 \), \( E^* = 335 \pm 75 \) kJ/mol, and \( V^* = 4.0\times10^{-6} \) m\textsuperscript{3}/mol.

We calculate pressure by assuming it is 1 GPa at the base of the crust with a pressure gradient of 30 MPa/km below this. At the lithosphere-asthenosphere boundary, we assume a temperature of 1573 K, with an adiabatic gradient of 0.5 K/km deeper into the asthenosphere. The unknowns for which we must solve in Equation (2.5) are shear stress and grain size.

To solve for these unknown parameters, we first predict strain rate as a function of depth using Equation (2.4). Given this strain rate, the temperature and pressure at a specified depth, and the parameters for dislocation creep listed above, we next predict the shear stress required to yield the same strain rate given by Equation (2.4) using the flow law given in Equation (2.5) for dislocation creep (i.e., when \( p = 0 \)). Then, using this predicted shear stress, we determine the corresponding grain size that gives the same strain rate using Equation (2.5) in its diffusion creep form (i.e., when \( p = 3 \)). For instance, if the transition between dislocation creep and diffusion creep occurred 100 km into the asthenosphere (\( z_0 = 100 \) km, where \( z_0 \) is the distance from the lithosphere-asthenosphere boundary to the depth of the transition between dislocation and diffusion creep directly under the deforming boundary), this model would require a shear stress of \( \sim 0.2 \) MPa and a grain size of \( \sim 6 \) mm. For a transition at 150 km into the asthenosphere (\( z_0 = 150 \) km), a shear stress of \( \sim 0.2 \) MPa and a grain size of \( \sim 7 \) mm is required, and a transition at 200 km below the lithosphere-
asthenosphere boundary also requires a shear stress of \( \sim 0.2 \) MPa, but a grain size of \( \sim 8 \) mm. Although strain rate decreases with depth, stresses increase to compensate for the influence of activation volume in the factor that includes it in Equation (2.5).

### 2.5.3 Predicted Patterns of Anisotropy

To generate a predicted pattern of anisotropy from the models discussed above, we follow relationships outlined in Ribe [1992] and Kaminski and Ribe [2002]. We first convert the predicted strain rates from each model into strain following McKenzie and Jackson [1983], who showed that in the case of simple shear, strain can be calculated by simply multiplying strain rate by the time over which the strain has accumulated. Modeling by Savage et al. [2007] suggest that the widespread anisotropy observed on the South Island is primarily a result of strike-slip deformation, with little effect from convergence. Thus, we ignore the compressional component of motion between the Pacific and Australian plates and assume all the strain accumulation is due to finite simple shear parallel to relative motion between the Pacific and Australian plates. Following Sutherland [1999], about 850 km of strike-slip motion has occurred between the Pacific and Australian plates. We assume the time over which this strain accumulated is equal to the time required to displace 850 km of the fault at the present rate of 36 mm/yr, yielding about 24 Ma and maximum strain rates on the order of \( 10^{-14} - 10^{-15} \) s\(^{-1}\). Although the predicted strain in the asthenosphere includes a plunge, we examine only the horizontal projection of that strain.

Numerical models of polycrystalline flow [Kaminski and Ribe, 2002] suggest that at temperatures appropriate for the asthenosphere, lattice preferred orientation develops in three stages. For strains less than 0.2, the LPO follows the evolution of the finite strain ellipsoid. For strains between 0.2 and 1.2, however, the LPO evolves faster than the finite strain ellipsoid, which Kaminski and Ribe [2002] parameterize using a model for dynamic recrystallization. We estimate orientations for such strains from the mean orientation of the olivine a-axis in simple shear given in Kaminski and Ribe [2002]. For strains larger than 1.2, olivine a-axes reach a steady-state orientation within the shear plane parallel to the shear direction. We note that Zhang and Karato’s [1995] experiments,
used by Kaminski and Ribe [2002], were conducted at high temperatures (1300°C), and although others have shown similar relationships at slightly lower temperatures of 1200°C [Bystricky et al., 2000], tests of Kaminski and Ribe’s [2002] parameterization have not been made for lithospheric temperatures.

Using laboratory-based flow laws for single crystals, Ribe [1992] calculated percent anisotropy as a function of natural strain for olivine aggregates. We used Ribe’s [1992] relationship for an aggregate composed of 70% olivine (equation 24 of Ribe [1992]) to convert our calculations of strain from each model to percent anisotropy. To predict approximate delay times from percent anisotropy, we simply multiply percent anisotropy by the layer thickness over shear wave speed, assuming that for predicted strain larger than 1.2, percent anisotropy is 10%, which is reasonable given estimates of percent anisotropy from Pn studies [Bourguignon et al., 2007; Scherwath et al., 2002]; however, since the delay time between fast and slow quasi-shear waves depends on both the magnitude of anisotropy and the thickness of the anisotropic layer, we do not emphasize the absolute magnitude of the predicted delay times. Instead, we use them to address relative variations in delay time as a function of distance from the deforming boundary. In the case of a thin-viscous sheet, we assume a lithospheric thickness of approximately 100 km, based on estimates of lithospheric thickness used in other studies [e.g., Stern et al., 2000]. By making such an assumption, we also ignore any crustal contribution to anisotropy, of which average delay times range from approximately 0.2 to 0.3 s for the South Island [Karalliyadda and Savage, 2013]. There is some agreement between the fast directions of local S phases made in Karalliyadda and Savage [2013] and the teleseismic SKS measurements made in our study, especially for southern portions of the South Island. Yet the smaller delay times calculated by Karalliyadda and Savage [2013] than those measured using SKS phases suggest that the bulk of the anisotropy must at least be deeper than the crust. For anisotropy in the asthenosphere, we assume an isotropic lithospheric thickness of 100 km, and we estimate a thickness of the underlying anisotropic zone as a function of the value chosen for \(z_0\) as discussed in the previous section. We assume a shear wave speed of 4.5 km/s.
2.6 Discussion

Our shear wave splitting measurements (Figure 2.7) indicate that large anisotropy with the orientation of the fast quasi-shear wave nearly parallel to the direction of relative plate motion spans a zone approximately 100 – 200 km wide in central portions of the South Island. We estimate relevant parameters for our two models, namely, the width of deformation ($4y_0$) and corresponding length of the deforming boundary ($\lambda/2$) for the thin-viscous sheet model, and shear stress ($\sigma$) and grain size ($d$) for a model of asthenospheric shear. Calculated RMS misfits for the orientation of the fast-quasi shear wave and delay time between our shear wave splitting measurements and these predictions for the central portion of the South Island (Figure 2.8a-b and Table 2.3) show that both the lithospheric deformation and asthenospheric shear models can be made to fit the data comparably well. This is not surprising since we chose model parameters so that the predicted patterns match the observed patterns.

A thin-viscous sheet model with a deformation width between $4y_0 = 100$ and 140 km (Figure 2.9b-c and Figure 2.10) yields a predicted pattern of anisotropy similar to that observed (Figure 2.9a). These widths of deformation correspond to a deforming boundary length ($\lambda/2$) around 544 – 762 km, consistent with the approximate length of the boundary, as well as with $\sim$ 850 km of displacement thought to have occurred between the Pacific and Australian plates over the past 45 Ma [Cande and Stock, 2004; Sutherland, 1999]. Likewise, a model of shear in the asthenosphere (Figure 2.9d-e and Figure 2.10) with a shear stress in the asthenosphere of approximately 0.21 – 0.22 MPa and with an average grain size on the order of 6 ± 1 to 7 ± 1 mm (corresponding to $z_0$ = 100 to 130 km) also yields a similar pattern of anisotropy to that observed.

Given the long period filter used in making the shear wave splitting measurements, individual SKS waves sample a wider diameter along the geometrical ray path (i.e., larger Fresnel zone) than shorter period seismic waves. Thus, the width of the deforming zone may be narrower than the 100 – 200 km zone we infer. Rümpker and Ryberg [2000] show SKS phases with a dominant period of 16 s (similar to those we analyzed) arriving at a station directly above an anisotropic region sample
Figure 2.8: Calculated RMS misfit in predicted orientation of the fast quasi-shear wave relative to motion between the Australian and Pacific Plates and predicted delay time for A) a model of lithospheric deformation and B) a model for asthenospheric deformation.
Figure 2.9: Comparison of observed shear wave splitting to predictions of the orientation of the fast quasi-shear wave relative to motion between the Pacific and Australian plates. Lengths correspond to delay time ($\delta t$). A) Measurements made in this study. B-C) Predictions from a model for lithospheric deformation with a deforming zone width of $4y_0 = 100$–$140$ km, respectively. D-E) Predictions from a model for asthenospheric deformation with $\sigma = 0.21$ MPa and $d \approx 6$ mm (corresponding to $z_0 = 100$ km), and with $\sigma = 0.22$ MPa and $d \approx 7$ mm (corresponding to $z_0 = 130$ km), respectively.
Figure 2.10: A) Orientations of polarizations of fast quasi-shear waves (relative to plate motion between the Pacific and Australian plates) versus distance from the center of the shear zone. Negative and positive distances are for stations west and east of the center of the shear zone, respectively. B) Delay time versus distance from the center of the shear zone. In both A) and B), red circles indicate those stations with fast direction orientations within 20° of the orientation of relative motion between the Pacific and Australian plates. These represent those measurements from the central portion of the South Island reflecting simple shear along a deforming zone.
a region 140 km wide at the surface. This sampled region becomes 180 km wide at 100 km depth and approximately 240 km wide at 200 km depth. Given the 100 km spacing of the seismic stations and the clear difference in measured anisotropy between onshore the South Island and offshore the east coast, there is no reason to think the width of the deforming zone is significantly narrower than 100 – 200 km wide.

Thus the question remains, given these model parameters and other currently available data on the mantle characteristics of the South Island of New Zealand, are predictions from either model plausible? Specifically, is the lithosphere underlying the South Island and offshore thick enough to yield large delay times similar to those observed from the teleseismic shear wave splitting measurements? Are the shear stresses and grain sizes calculated here viable for producing anisotropy at asthenospheric depths?

A shear wave speed of 4.5 km/s and 10% anisotropy requires a 90 km thick anisotropic layer to yield a delay time of 2 s, similar to the largest delay times measured in this study. The crust offshore and in coastal areas of the South Island is approximately 25 km thick [Salmon et al., 2013]. Under the Southern Alps, the thickness increases to just over 40 km. Furthermore, recent work suggests that at temperatures below 700 ± 100°C, olivine will not flow significantly at geologic stresses [e.g., Mei et al., 2010] and therefore will not produce a lattice preferred orientation. The modeling work of Savage et al. [2007] places the 700°C isotherm in the 40 km depth range for this region. Therefore, this would require a total lithospheric thickness of 130 km (a 40 km thick isotropic region overlying a 90 km thick anisotropic region) for the anisotropy to be completely accommodated within the lithosphere.

Determining the thickness of lithosphere remains a controversial topic, though evidence from P-wave delays suggest that the lithosphere beneath the axis of the South Island may have been thickened, perhaps by two times, from an original thickness of 100 – 120 km [Stern et al., 2000]. While this would imply a somewhat deeper lithosphere-asthenosphere boundary than that inferred by other studies of surrounding regions [e.g., Fichtner et al., 2010; Rychert and Shearer, 2009], P-wave tomography using local and regional events yielded a high velocity region to 180 km depth.
Kohler and Eberhart-Phillips, 2002], suggesting lithosphere to this depth. Thus, we deduce that the lithosphere could host the observed anisotropy over most of the study region, especially given that larger delay times under portions of the Southern Alps might reflect thickened lithosphere.

Our model of localized lithospheric deformation with shear in the asthenosphere predicts shear stresses and grain sizes that fall within currently accepted ranges for such values. From the model strain rates and shear stresses, we calculate an effective viscosity of the asthenosphere on the order of 1020 Pa s, consistent with other studies [e.g., Conrad and Behn, 2010]. Furthermore, estimates of grain size in the asthenospheric upper mantle range from several millimeters to centimeters [Becker et al., 2008; Hirth and Kohlstedt, 2003; Podolefsky et al., 2004], making a grain size close to 6 – 7 mm (as required in this study) realistic.

More problematic is determining the grain size required for a transition from dislocation creep to diffusion creep and therefore, the transition from a region susceptible to the formation of lattice preferred orientation (dislocation creep) to one that is not (diffusion creep, see Section 2.3). Some estimate this transition at about 200 km depth with a grain size around 1 cm [Hirth and Kohlstedt, 2003]. Later work investigated the effects of hydration on the depth of this transition as well as grain size in oceanic mantle [Behn et al., 2009]. Behn et al. [2009] found that, for mantle lithosphere older than 60 Myr with an olivine water content of 1000 H/10^6Si, grain size reached a minimum of 15 – 20 mm at ~ 150 km depth or 20 – 30 mm at ~ 400 km depth, values greater than that predicted for the transition between dislocation creep and diffusion creep. Taking into account the errors associated with the parameters used in Equation (2.5), we calculate that the transition between the creep mechanisms occurs for ranges of shear stress from 0.06 – 0.61 MPa and grain size from about 5 – 7 mm at z_0 = 100 km. For z_0 = 130 km, shear stress ranges from 0.05 – 0.67 MPa, with grain size ranging from about 5 – 8 mm. These ranges for grain size are less than the steady state grain sizes proposed by Behn et al. [2009]; we do not have independent constrains on how large grain size could actually be in this region.

Concerning the role of water content on the evolution of LPO, laboratory experiments show that the influence of water on predicted anisotropy is largest when B-type or C-type fabrics are
produced [e.g., Karato et al., 2008, and references therin]. In these cases, the fast quasi-shear wave can be polarized normal, not parallel, to the orientation of shearing (horizontal flow with B-type fabric; vertical flow with C-type). In the South Island of New Zealand (at least in regions like the Southern Alps that lie outside of subduction zones), the water content is likely not high enough to induce these fabrics. Furthermore, the shear stresses in the asthenosphere calculated from our localized lithospheric deformation model are low relative to those required to promote B-type and C-type fabric. Thus, we infer that hydration of the asthenosphere does not strongly affect anisotropy beneath New Zealand.

Although this SKS splitting study alone cannot definitively discriminate between distributed versus localized modes of lithospheric deformation in the South Island of New Zealand, further analyses that can place a depth extent on the anisotropy will aid in this debate. A study of Pn travel times also utilizing the data collected during the MOANA experiment [Collins and Molnar, manuscript in preparation, 2013] shows fast Pn propagation for paths crossing the southeast side of the South Island to be aligned NNW-SSE, similar to that of fast quasi-shear waves. Thus, the Pn data suggest that some of the shear wave splitting observed at OBSs southeast of the island is within the lithosphere. Fast propagation of Pn for paths beneath most of the island and offshore to the northwest (in a zone 150 – 300 km wide) is approximately parallel to the relative motion between the Australian and Pacific plates. Hence for that region, it is similar to the shear wave splitting measurements reported here. Splitting from local S-phases on both the MOANA OBSs and GeoNet stations also suggest a possible lithospheric origin for the observed anisotropy [Karalliyadda, manuscript in preparation, 2013].

Given these new shear wave splitting measurements and the available data on the characteristics of the mantle under the South Island of New Zealand, it appears that either distributed lithospheric or asthenospheric deformation is plausible. If anisotropy is the result of distributed deformation of the mantle lithosphere, we expect a width of deformation around $4y_0 = 100-140$ km. If anisotropy is the result of localized lithospheric deformation with pervasive shear in the asthenosphere, grain sizes on the order of 6 – 7 mm are required at depths between 100 and 130
km below the lithosphere.

2.7 Conclusions

New shear wave splitting measurements made on and offshore the South Island of New Zealand show an approximately 100 – 200 km wide zone of 1.0 – 2.0 s delay times with orientations of the fast quasi-shear waves nearly parallel to the direction of relative motion between the Pacific and Australian plates. Regions of distinctly different patterns of anisotropy bound this zone of high anisotropy. Approximately 300 km off the west coast, little to no shear wave splitting is observed. Offshore the east coast, the anisotropy is clearly not a result of simple shear between the Pacific and Australian plates and is consistent with its being due to current absolute plate motion of the Pacific plate or frozen into the lithosphere.

We used these new shear wave splitting results to test two models of deformation to explain the observed anisotropy: (1) a thin-viscous sheet model mimicking distributed deformation in the mantle lithosphere and (2) a model mimicking localized deformation in the mantle lithosphere around a fault that extends to the asthenosphere, where deformation is then accommodated in a fan-like pattern centered beneath the fault. A model of a thin-viscous sheet with a width of deformation around 100 – 140 km could explain the observed pattern of anisotropy. Likewise, if grain size at depths of 100 – 130 km into the asthenosphere were \( \sim 6 – 7 \) mm, localized lithospheric deformation with diffuse shear in the underlying asthenosphere also fits our measurements equally well. In any case, deformation in the upper mantle along this major shear zone in the South Island of New Zealand appears confined to a 100-200 km wide region.

2.8 Acknowledgements

We thank Josh Stachnik and Zhaohui Yang for their assistance in orienting the horizontal components of the OBSs and Megan Anderson for her help with the SplitLab software. We also thank the captain and crew of the R/V Thomas G. Thompson (cruise TN229) in 2009 and of the R/V Roger Revelle (cruise RR1002) in 2010. All 30 OBSs were manufactured by Scripps Insti-
tution of Oceanography and are part of the United States National Ocean Bottom Seismograph Instrumentation Pool. These data are available for download from the IRIS Data Management Center. We acknowledge the New Zealand GeoNet project and its sponsors EQC, GNS Science, and LINZ for also providing data in this study. The figures were made with the Generic Mapping Tools (GMT) software by Wessel and Smith [1998]. The (former) New Zealand Foundation for Research, Science and Technology, and the National Science Foundation Continental Dynamics program supported this work under grants EAR-0409564, EAR-0409609, and EAR-0409835. We also thank editor Tom Parsons and two anonymous reviewers for constructive comments.
Table 2.3: Calculated RMS misfits between shear wave spitting measurements made in this study and predictions from both a thin-viscous sheet model representing lithospheric deformation (LD) and a model representing asthenospheric deformation (AD). Only those measurements in the central portion of the South Island and within 200 km of the Alpine fault were included in the error calculation (see Table 2.2).

<table>
<thead>
<tr>
<th>LD</th>
<th>RMS Misfit $\phi$ (°)</th>
<th>RMS Misfit $\delta t$ (s)</th>
<th>AD</th>
<th>RMS Misfit $\phi$ (°)</th>
<th>RMS Misfit $\delta t$ (s)</th>
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<tr>
<td>$4y0=40km$</td>
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<td>$z0=50km$</td>
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Chapter 3

Teleseismic $P$-wave tomography of South Island, New Zealand upper mantle: Evidence of subduction of Pacific lithosphere since 45 Ma

A $P$-wave speed tomogram produced from teleseismic travel-time measurements made on and offshore the South Island of New Zealand shows a nearly vertical zone with wave speeds that are 4.5% higher than the background average reaching to depths of approximately 450 km under the northwestern region of the island. This structure is consistent with oblique west-southwest subduction of Pacific lithosphere since about 45 Ma, when subduction beneath the region began. The high-speed zone reaches about 200 - 300 km below the depths of the deepest intermediate depth earthquakes (subcrustal to $\sim$ 200 km) and therefore suggests that $\sim$ 200 - 300 km of slab below them is required to produce sufficient weight to induce the intermediate depth seismicity. In the southwestern South Island, high $P$-wave speeds indicate subduction of the Australian plate at the Puysegur trench to approximately 200 km depth. A band with speeds $\sim$ 2 - 3.5% lower than the background average is found along the east coast of the South Island to depths of $\sim$ 150 - 200 km, and underlies Miocene or younger volcanism; these low speeds are consistent with thinned lithosphere. A core of high speeds under the Southern Alps associated with a convergent margin and mountain building imaged in previous investigations is not well resolved in this study. This could suggest that such high speeds are limited in both width and depth and not resolvable by our data.
3.1 Introduction

Initiating around 45 Ma, oblique convergence between the Australian and Pacific plates occurs across the South Island of New Zealand [e.g., Norris et al., 1990; Sutherland, 1995]. Subduction zones exist to the north and south of the South Island (Figure 3.1). At the Hikurangi trench, oblique west-southwestward subduction of the Pacific plate beneath the Australian plate accommodates convergence in the northern South Island. In the southern South Island, accommodation of convergence occurs via eastward subduction of oceanic Australian plate beneath the Pacific plate at the Puysegur trench. The more than 800 km of motion between the plates since 45 Ma [Sutherland, 1995, 1999; Sutherland et al., 2000] includes about 460 km of strike slip along the Alpine fault, a predominately dextral fault running the length of the South Island (Figure 3.1). Around 11 Ma, a component of convergence initiated [Cande and Stock, 2004], resulting in ~70 km of shortening across the central South Island, and the uplift of the Southern Alps. Thus, the full budget of relative plate motion across the region can be calculated [e.g., Cande and Stock, 2004], with the geologic evolution of the central part of the South Island responding largely to right-lateral shear between the Pacific and Australian plates, and oblique subduction dominating Cenozoic tectonics in the northern part of the island. What is less clear is how the mantle lithosphere responded to such relative plate motion.

Subduction in the south at the Puysegur trench initiated during the Miocene with about 170 km of subduction since that time [Eberhart-Phillips and Reyners, 2001; Sutherland et al., 2000]. Subduction in the northern part of the South Island is much older and initiated as early as 45 Ma with as much as 1000 km of subducted lithosphere [Cande and Stock, 2004; Sutherland, 1995; Sutherland et al., 2000]. Studies of seismicity show that earthquakes associated with subduction occur to depths of about 300 km under the North Island and about 200 km under the northwestern South Island [Anderson and Webb, 1994]. A few isolated earthquakes have occurred as deep as ~600 km [Adams, 1963] beneath the North Island and no intermediate depth earthquakes associated with subducting Pacific lithosphere occur farther south than about 42°S (Figure 3.1). Fault plane
Figure 3.1: Location of seismographs used in this study. The light blue squares are ocean bottom seismometers associated with the MOANA array and the purple squares are the New Zealand National Seismograph Network maintained by GeoNet. Stations CASS, CROE, KELY, and FREW are temporary stations deployed by Victoria University of Wellington. Twenty-nine of the OBSs were equipped with Trillium 240 seismometers (a broadband instrument) with the final OBS (station NZ14) equipped with a Trillium 40 seismometer (an intermediate period instrument). All instruments recorded continuously at a sample rate of 50 Hz. We note that among the 30 OBSs deployed, station NZ01 did not yield usable seismic data and station NZ17 was not recovered. Bathymetric features including the Campbell Plateau, Challenger Plateau, Chatham Rise, the Hikurangi Trench, and the Puysegur Trench, along with the cities of Christchurch and Dunedin, and the Alpine fault, are labeled for reference. Earthquakes since 2009 with $m_b > 4$ and focal depths greater than 40 km highlight the inclined seismic zones associated with subduction of the Pacific plate beneath the North Island and northern South Island and of the Australian plate beneath Fiordland in southwest New Zealand. Red triangles mark locations of Miocene or younger volcanism. The thick arrows show current absolute plate motion of the Australian and Pacific plates in a hotspot reference frame [Gripp and Gordon, 2002] and the thin arrow shows relative motion between the Australian and Pacific plates.
solutions of the intermediate depth earthquakes typically show downdip extension, as is common within subducting slabs [Isacks and Molnar, 1971]. Thus, the generation of such intermediate depth earthquakes is governed by gravity acting on the excess density in the downgoing slab at some depth below the seismicity. Given the known total amount of plate convergence [Sutherland, 1995], it is possible to determine the amount of subducted slab below the intermediate depth earthquakes.

The approximately 200-km-wide array aperture defined solely by on-land seismograph stations limited previous investigations. In this study we incorporate data recorded by a temporary deployment of ocean bottom seismometers that expands the array aperture to over four times the width of the South Island and enhances both lateral and depth resolution of the $P$-wave speed structure under the island.

### 3.2 Data and Methods

#### 3.2.1 Data

The Marine Observations of Anisotropy Near Aotearoa (MOANA) experiment consisted of a one-year deployment of ocean bottom seismometers (OBSs) installed off both coasts of the South Island of New Zealand from late January 2009 to early February 2010 (Figure 1). This array of 30 OBSs was deployed with approximately 100 km spacing, with a full array aperture of over 900 km NW - SE, increasing the extent of the network to over four times the width of the South Island. Four stations (NZ01 - NZ04) lay atop oceanic crust. The rest were deployed on either the Challenger Plateau (west of the island) or the Campbell Plateau (east of the island), submerged continental crust of the Australian and Pacific plates, respectively. Concurrently with this OBS array, four temporary broadband stations were deployed on land around the Arthur’s Pass region (CASS, CROE, KELY, and FREW; Figure 1). This temporary array complemented the permanent New Zealand National Seismograph Network maintained by GeoNet [Petersen et al., 2011]. Some stations exhibited reversed polarity on the vertical component (Figure 3.2). We corrected for this before making any travel-time measurements.
Figure 3.2: Example waveforms filtered from 0.05 - 0.1 Hz (left) and unfiltered (right) recorded from an earthquake with $m_b = 6.7$ in the Banda Sea, with a backazimuth of $305^\circ$. Solid red lines denoted OBSs. The polarities of vertical components for KHZ and QRZ (denoted by a dashed black line) are reversed and were corrected before making travel-time picks.
Eight different seismometer types were deployed between the OBSs, the temporary land stations, and the permanent GeoNet array. Convolving the eight different instrument responses with a delta function showed negligible difference in responses between the various stations in the time domain and thus, we did not correct for instrument response. For the OBS stations we attempted to correct for the effect of noise recorded by vertical components and generated by both deformation of the seafloor from infragravity waves and tilt of the instrument following the methods of Crawford and Webb [2000] and Webb and Crawford [1999]. Infragravity waves produce their most significant seafloor pressure fluctuations at periods much longer than those of interest in this body wave study (> 30 sec or < 0.03 Hz) and removal of the infragravity signal did not improve the signal quality. Likewise, removing tilt noise showed no improvement in the OBS seismic records, likely due to the poor coherence (< 0.5) between the horizontal and vertical components of the OBSs across the spectrum. Thus, we did not make corrections for long-period seafloor noise or tilt noise.

3.2.2 Travel-Time Measurements

We initially examined seismograms (Figure 3.2) from all events of \( m_b \geq 5.5 \) occurring from February 2009 to February 2010 with epicentral distances between 25° – 100°. Of these, we observed clear \( P \)-wave arrivals on more than 50% of the 57 stations from 53 of these earthquakes in the filter band 0.05 - 0.1 Hz (20 - 10 s) and 25 earthquakes in the filter band 0.08 - 0.12 Hz (12.5 - 8.3 s), similar to filter bands used in other OBS teleseismic body wave studies [e.g., Tanaka et al., 2009; Toomey et al., 1998; Wolfe et al., 2011]. From analysis of ambient noise recorded on the seismometers used in this experiment [Yang et al., 2012], these frequency bands are higher than the band in which infragravity waves dominate the noise, but includes the primary microseism noise peak. Despite the long period of these filter bands and their overlap with microseismic noise, they yielded the most earthquakes with adequate signal-to-noise ratios for which we could measure arrival times on both land stations and OBSs. To fill undersampled backazimuths, we also searched for clear \( PP \) arrivals. In total, we measured 2230 \( P \) and 129 \( PP \) arrival times in the band 0.05 -
0.1 Hz and 1027 $P$ and 39 $PP$ arrival times in the band 0.08 - 0.12 Hz on 56 earthquakes using dbxcor, a cross-correlation, beam-forming program (Figure D.1) [Pavlis and Vernon, 2010].

Many of these earthquakes occurred in the subduction zones offshore Japan, Sumatra, and Fiji-Tonga. In order to ensure an even distribution of earthquakes, so as not to give undue weight to a particular backazimuth, we separated the 56 earthquakes into 10° backazimuth and 10° distance bins and selected one earthquake from each bin. This yielded 29 earthquakes and a total of 1171 $P$ and 129 $PP$ picks in the band 0.05 - 0.1 Hz and 621 $P$ and 39 $PP$ picks in the band 0.08 - 0.12 Hz (Figure 3.3) for an RMS travel-time residual of 0.845 s.

### 3.2.3 Crustal and Sediment Corrections

Without the application of crustal corrections in teleseismic $P$-wave tomography, erroneous model structure may appear near the Moho and uppermost mantle [e.g., Humphreys and Clayton, 1990]. To calculate crustal thicknesses we used the recent work by Collins and Molnar [2014], who measured $Pn$ travel times from regional earthquakes recorded across both the MOANA array and the GeoNet array. Because the $Pn$ time term accounts for delays through both crust and sediment and given the large thickness of sediment known to exist on either side of the South Island [e.g., Ball et al., 2014; Kennett et al., 1975; Wood and Woodward, 2002], we also corrected for a sediment layer beneath the OBSs using sediment thicknesses from Whittaker et al. [2013] before calculating the crustal thickness. Given sediment and crustal thicknesses (Table D.1), we calculated travel times through the crust and sediment for each $P$-wave arrival by taking into account the angle of incidence. We assumed the total correction is the sum of the travel-time difference between a $P$-wave traveling in the mantle versus sediment plus the difference between a $P$-wave traveling in the mantle versus crust (see Appendix D.1). Following Collins and Molnar [2014], we assumed a crustal $P$-wave speed of 6.2 km/s and a mantle speed of 8.1 km/s. At stations where $Pn$ measurements were not published (NZ02, NZ03, NZ04, APZ, DCZ, KHZ, NNZ, PYZ, and THZ), we assumed a crustal thickness given by Salmon et al. [2013]. We removed the mean correction for crust and sediment from the travel-time residual for each earthquake. We also examined the effect that
Figure 3.3: A) Locations and azimuthal distribution of earthquakes used in this study. Circles denote events used for $P$-wave travel times, and diamonds denote earthquakes for which $PP$ picks were made. All events were $m_b > 5.5$ and occurred from 2009 - 2010. B) Azimuthal distribution of earthquakes used in this study. The radial axis is the number of earthquakes in each $10^\circ$ azimuthal bin.
inverting for additional station terms had on the model. With inclusion of such terms, we found that magnitudes of speed anomalies in the top approximately 200 km of the model decreased when station terms were allowed. The general pattern of travel-time residuals did not change, however, when station terms were included. Thus, we do not invert for station terms in the results discussed here, but do tabulate such station corrections in Table D.1.

3.2.4 Patterns of Travel-Time Residuals

Patterns of teleseismic $P$-wave travel-time residuals, which vary in magnitude from approximately -2.5 s to +3 s after application of crustal and sediment corrections, reveal lateral variations in $P$-wave speeds (Figures 3.4 and D.2). To demonstrate lateral heterogeneity, we present maps of residuals for earthquakes in three backazimuths: those from the northeast (Figure 3.4a-b and Figure D.2a), south (Figure 3.4c-d and Figure D.2b-c), and northwest (Figure 3.4e-f and Figure D.2d-e). Stations on the South Island record early arrivals from events that lie to the northeast, compared with OBSs off both coasts (Figure 3.4a-b and Figure D.2a), but exceptions include the OBSs just off the central west coast (NZ14 and NZ15), which record early arrivals. This pattern suggests that relatively high speeds underlie the northwest part of the island and seafloor just west of it, though rays to these stations passing directly through the Pacific slab beneath the North Island may contribute to the early arrivals at stations on the northern end of the South Island [e.g., Williams et al., 2013].

$P$-wave travel-time residuals, ranging in magnitude from approximately -2 s to +1.5 s, from earthquakes from the south (Figure 3.4c-d and Figure D.2b-c) show late arrivals at stations on the South Island and at OBSs off the east coast. Stations along the west coast show a mix of early and late arrivals, with OBSs within about 200 km of the west coast showing early arrivals. These early arrivals, with residuals $< -1$ s, at northwestern OBSs are again consistent with a high-speed zone beneath the northwestern part of the South Island and its offshore region.

For paths from the northwest (Figure 3.4e-f and Figure D.2d-e), OBSs deployed off the west coast show late arrivals compared with most land stations. The sign of the residuals recorded on
Figure 3.4: Measured $P$-wave travel-time residuals in the frequency band 0.05 - 0.1 Hz after application of a crustal and sediment correction for six events from a range of backazimuths. Figure D.2 shows residuals measured in the band 0.08 - 0.12 Hz. Earthquakes: A) near Samoa with backazimuth of 32° ($m_b = 7.1$); B) near Fiji with backazimuth 22° ($m_b = 5.7$); C) near the South Sandwich Islands with backazimuth 177° ($m_b = 6.2$); D) also near the South Sandwich Islands with backazimuth 171° ($m_b = 6.2$); E) from offshore the Solomon Islands with backazimuth of 334° ($m_b = 6.3$); and F) from offshore Japan with backazimuth of 333° ($m_b = 6.5$). Areas of the circles are proportional to the magnitudes of the travel-time residuals. Black arrows point from the earthquake epicenter. A negative (positive) travel-time residual indicates an early (late) arrival with respect to average times for the AK135 reference model [Kennett et al., 1995].
OBSs off the east coast depends on their distance from the South Island. *P*-waves from earthquakes occurring within approximately 60° of the island (Figure 3.4e and Figure D.2d) arrive early off the east coast, whereas those from earthquakes greater than 60° away seem to arrive late at some stations (Figure 3.4f and Figure D.2e). Residuals again range from approximately -2 to +1.5 s.

### 3.2.5 Tomographic Method

We used the finite-frequency tomography code of *Schmandt and Humphreys* [2010] to invert *P*-wave arrival-time measurements. This code uses sensitivity kernels to account for volumetric variations in wave speed based on 1D ray tracing in the AK135 reference model [*Kennett et al.*, 1995]. It considers sensitivity only in the first Fresnel zone, calculated by an approximation of the Born theoretical “banana-doughnut” kernel of *Dahlen et al.* [2000]. The radius, *R*, of the first Fresnel zone can be approximated as

\[
R = \sqrt{\left(\frac{v}{f}\right) \frac{\delta(D - \delta)}{D}},
\]

where *v* is wave speed, *f* is frequency, *δ* is distance along the ray path, and *D* is the total ray path length [*Spetzler and Snieder*, 2004]. For a 5000 km long ray path (typical for a *P*-wave produced by a shallow earthquake 45° away) and assuming a mantle *P*-wave speed of 8.1 km/s and a center frequency of 0.075 Hz (∼13 s), the radius of the first Fresnel zone would be about 100 km at 100 km depth and about 175 km at 300 km depth. Thus, resolving features with lateral dimensions as small as 100 km is not likely.

The region for which we determined lateral variations in wave speed ranges in latitude from approximately 35°S to 51°S and in longitude from approximately 158°E to 177°W. The model structure extends in depth from 60 - 600 km. We did not invert for structure shallower than 60 km because we corrected the travel-time measurements for crustal thickness and sediment prior to the inversion. A trapezoidal mesh (to account for the curvature of the Earth) of 70 x 71 horizontal nodes and 13 vertical nodes (representing the number of depth slices) parameterizes the structure. This resulted in a node spacing of about 20 km in the center of the structure and about 30 - 35 km
towards its edges. All node spacings are less than the approximate width of the first Fresnel zone sensitivity kernel at the frequency band of the measured $P$-waves. With over 30 times more nodes than travel-time measurements, the inferred structure will appear highly smoothed; however, the underdetermined nature of the model is offset by the width of the Fresnel zone radii. For instance, since at 100 km depth the radius of the first Fresnel zone for a center frequency of 0.075 Hz is about 100 km, about 50 nodes in this depth layer are sampled by one ray.

We applied both model norm damping and spatial smoothing to help regularize the inverse problem, which then minimizes the cost function

$$E = \|Am - d\|^2 + \gamma\|Lm\|^2 + \varepsilon\|m\|^2, \quad (3.2)$$

where $A$ is a matrix that contains the partial derivatives relating travel-time residuals to the model parameters, $m$ is a vector containing the perturbations to the velocity model, $d$ are the differential travel-time data, $\gamma$ is the smoothing parameter, $L$ is the smoothing matrix where the weight of smoothing between neighboring nodes decreases with inverse distance, and $\varepsilon$ is the damping parameter [Schmandt and Humphreys, 2010]. We used a standard trade-off analysis between variance reduction and model norm to determine suitable damping and smoothing parameters (Figure D.3). Varying the smoothing parameter had little effect on the resulting trade-off curve, especially for damping values $\geq 5$, likely due to inherent smoothing from using low-frequency data.

### 3.2.6 Model Resolution

To test the expected model resolution given our culled arrival time measurements (29 earthquakes, see Section 3.2.2) and damping and smoothing parameters, we used synthetic hypothesis and checkerboard tests (Figures 3.5, 3.6, D.4 and D.5). We also show representative ray coverage through the central portion of South Island (Figure 3.6f-g). The synthetic checkerboard test (Figures D.4 and D.5) consisted of cubes with approximately 200-km-long sides with alternating positive and negative amplitudes of +5% or -5% in a neutral background of the reference speeds at each depth, AK135. We chose the size of the cubes to be on the order of the width of the first
Fresnel zone at 100 km depth. In general, with the same source locations and stations as in the inversion of real data and no added synthetic noise, the locations of the checkers are recovered, but the amplitudes of anomalies are underestimated. Maximum amplitude recovery ranges from approximately 20% to 96%, with amplitude recovery above 80% for the top 125 km. At 330 km and below, amplitude recovery is around 30% or less.

Although commonly used to test resolution, checkerboard tests are poor for identifying possible artifacts in the results of the inversion [e.g., Eilon et al., 2015]. Thus, we also determined expected model resolution for a hypothetical structure representing a high-speed anomaly spanning nearly the length of the South Island between the Hikurangi and Puysegur subduction zones (Figures 3.5 and 3.6). We used 200-km-wide input structures reaching two different depths, 330 km (the depth below which we found amplitude recovery to fall below about 30% from the checkerboard test) and 565 km. The amplitude of the anomaly was set to +4% of the average background speed at each depth. The background average (i.e., the average speed perturbation at all nodes less than 4%) was about 0.1%. We carried out this test both with added synthetic noise, assigned randomly to the arrival times and approximated as a normal distribution with RMS equal to the RMS differential travel time of the best fitting model (RMS = 0.5 s), and with no synthetic noise. The addition of synthetic noise did not alter the recovered location of the anomalous body, but it did impact the amplitude by forcing the recovered anomaly to be larger in amplitude compared to the no-noise case.

In general, amplitudes are still underestimated; for a slab reaching to 330 km the maximum recovery (the maximum fraction of the input speed anomaly) averaged over the depths of 60 - 330 km is ~ 88%. Vertical smearing, a problem common in teleseismic tomography, is also evident, especially at depths below 330 km. For a slab reaching to 565 km depth, recovery of structure below about 300 km ranges from about 30% at the bottom of the model to about 60% at 330 km depth, but the average amplitude recovery over the depth range 60 - 330 km increased to ~ 91%. This suggests that although structure deeper than about 300 km is not as well resolved as shallower structure, it is possible that travel times through a deep structure leak into shallower structure and
Figure 3.5: Recovery of 200-km-wide synthetic input structures (+4% $V_p$ of the background model of AK135) representing high-speed anomalies to 330 km deep (top panels) and 565 km deep (bottom panels). Synthetic travel times were calculated by tracing rays through the synthetic structure using the same station and event distribution as observed. Random noise of RMS = 0.5 s was added to the synthetic travel times. The synthetic travel times were then inverted for a 3D structure using the same method as with the measured arrival times. Depth slices at 90, 165, 245, 330, and 420 km are shown.
Figure 3.6: Cross-sections through the synthetic $P$-wave tomographic structure demonstrating resolution. A-B) Input and resolution of a 200-km-wide high-speed zone (+4\% $V_p$ of the background model, AK135) extending to 330 km depth. C-D) Input and resolution of a 200-km-wide high-speed zone extending to 565 km. E) Map showing location of these cross-sections. F-G) Example ray coverage for a transect through the center of both our array and the South Island of New Zealand. Rays within 2° of the transect are plotted.
enhance inferred speeds within structures at shallow depths. The width of the inferred high-speed zone increases with depth. This is likely the result of decreasing resolution with depth as well as an increase in the radius of the first Fresnel zone with depth. As calculated from Equation (3.1), the radius of the first Fresnel zone would be about 100 km at 100 km depth and about 175 km at 300 km depth. This test and calculation suggest that lateral variations in speed below about 300 - 350 km should be resolvable, though not as well as those shallower depths, and due to the increasing width of Fresnel zones, lateral dimensions of resolvable wavespeed anomalies should broaden with depth.

3.3 Results

We show the P-wave tomogram under the South Island of New Zealand in 11 depth slices ranging from 60 - 465 km (Figure 3.7) and 3 cross-sections traversing the central portion of South Island perpendicular to the surface expression of the Alpine fault (Figure 3.8). One measure of the quality of a tomographic inversion is the data variance reduction. The inferred structure accounts for 71% of 0.714 s² variance of observed residuals. With an average error in the P-wave travel-time measurements of about 0.3 s and an RMS travel-time residual of 0.845 s, the best possible variance reduction would be ∼ 87%. Thus the inferred structure explains much of the variance in the measured travel-time residuals.

To the west and south of the city of Christchurch (Figure 3.1), a low-speed region (approximately -2 to -3.5%) reaches depths of 150 km (Figure 3.7a-d). Similar magnitude low speeds are also apparent in the upper mantle beneath Dunedin. For steep rays and rays from the east, late arrival times recorded by station ODZ (Figure 3.9a), just north of Dunedin and southwest of Christchurch (Figure 3.1), are consistent with low wave speeds along the east coast. This low-speed region is defined clearly at depths shallower than 150 – 200 km, but not at greater depths. Comparably low speeds also exist off the west coast, but because they are on the edge of the array are less reliable than anomalies found in the center of our model array.

In the northern South Island, a nearly vertical high-speed body, on average 4.5% higher
Figure 3.7: Depth slices of $P$-wave speed variations from the tomographic inversion utilizing stations shown in Figure 3.1 and earthquakes shown in Figure 3.3. Depth slices at 60, 90, 125, 165, 205, 245, 285, 330, 375, 420, and 465 km are shown.
Figure 3.8: Cross-sections through the $P$-wave tomogram. Map inserts show the locations of Profiles A-C, which traverse the South Island perpendicular to the surface expression of the Alpine fault. We did not invert for structure above 60 km. Black circles in A mark earthquake hypocenters in New Zealand since 2009 with $m_b > 4$ and between 40$^\circ$S and 42$^\circ$S.
Figure 3.9: Lower hemisphere stereographic projections of observed (top row) and predicted (middle row) $P$-wave travel-time residuals and their differences (bottom row), plotted by incident angle at the Moho, for stations ODZ (A - C), NNZ (D - F), and NZ14 (G - I). Stations are chosen to show residuals near the east coast low-speed anomaly (ODZ), top of subducting Pacific plate (NNZ), and the west coast high wave speed anomaly (NZ14). Areas of circles are proportional to magnitudes of the travel-time residuals.
than reference speeds, is evident to a depth of 400 - 450 km, and becomes weaker and shallower southward (Figures 3.7 and 3.8). Another nearly vertical feature is imaged in the southwestern South Island, to a depth of ∼ 200 km (Figure 3.7a-f). We note that although magnitudes of anomalies are large (especially for the northwestern feature) and exceed 4.5% higher than reference speeds, the best average amplitude recovery found in synthetic tests using a high-speed region of similar shape and depth was ∼ 74 - 91%. Thus, although speeds in the northern feature appear to be on average approximately 4.5% higher than reference speeds, they could be 5 - 6% faster, if not more if confined to a narrower region than we can resolve, as in the feature recovery tests shown in Figures 3.5 and 3.6. Such a large anomaly in speed is consistent with the magnitudes of residuals. For instance, travel-time measurements made at station NNZ from earthquakes to the northwest and with angles of incidence around 20° are earlier by 2 s or more than the average for the region studied here (Figure 3.9d). If such a 2 s residual resulted from passage through a layer 300 km thick (between depths of 100 and 400 km) at a 20° angle of incidence, for which the $P$-wave travel time would be ∼ 40 s, the average speed would be 5% greater than normal. If the structure beneath the surrounding offshore region represented a global average of the mantle, with $P$-wave residuals of ∼ 1 s there, then arrivals at NNZ roughly 3 s earlier than normal would require average speeds in the upper 300 km of the mantle beneath NNZ be higher by 7.5% than beneath other regions, or ∼ 5% higher if the high-speed zone extended over a depth range of 450 km. Such large anomalies are consistent with passage of rays through subducted lithosphere [e.g., Fry et al., 2014; Mitronovas and Isacks, 1971; Zhao, 2012].

Squeeze tests are useful for investigating the minimum depth extent of the structure required to satisfy the travel-time data set. We performed the squeeze tests in two stages: 1) we inverted residuals by forcing them into a model region less than a specified depth, and 2) we inverted the resulting residuals from stage 1, but in a model with a relaxed depth constraint. Seismicity due to subducting Pacific lithosphere in the northwest of the South Island decreases to a minimum around 300 km deep [Anderson and Webb, 1994] and thus, we squeezed the model to 330 km. Inversion of the residuals from that structure shows that under the northwestern portion of the South Island,
the travel-time data require speeds below a depth of 330 km that are higher than average by at least 2% (Figure 3.10).

The high speeds extending offshore the north and central west coast of the South Island might be contaminated by the inclusion of travel-time measurements made on the OBSs for which we might have overestimated thicknesses of crust and sediment, particularly stations NZ14 and NZ15. Almost all arrivals at NZ14 (Figure 3.9g) and NZ15 are 1 - 2 s early, with the earliest from earthquakes to the northeast and east. Measurements of \( Pn \) speeds at stations NZ14 and NZ15 show large anisotropy of approximately 8%, with the fastest speeds for horizontal paths trending approximately parallel to N60\(^\circ\)E [Collins and Molnar, 2014]. Zietlow et al. [2014] also measured large SKS split times (> 2 s) in this region, with similar orientations of fast quasi-\( S \) waves. This suggests \( P \)-wave anisotropy in the uppermost mantle beneath this offshore region may contribute to the travel-time advances from teleseisms. To address this possibility, consider a 100 km thick anisotropic layer, \( P \)-wave anisotropy of 8% with high speeds for horizontal paths aligned NE–SW. For a teleseismic \( P \)-wave arriving with a 45\(^\circ\) angle of incidence, the average travel time would be 17 - 18 s, and with average anisotropy of 4%, such \( P \)-waves could be advanced by \( \sim 0.7 \) s for teleseismic paths from N60\(^\circ\)E or S60\(^\circ\)W. To ensure that this high-speed body is not caused by a systematic error at these stations or that early arrivals at these stations result from anisotropy, we performed an inversion with stations NZ14 and NZ15 removed. Without these stations, the resulting structure (Figure D.6) shows this NE-trending high-speed feature along the northwest coast, but with both its lateral extent (based on the core of +4.5% \( Vp \) speeds) and the magnitude of the maximum anomaly reduced: the region with speeds of +4.5% \( Vp \) reaches only to about 250 km depth compared to about 450 km depth when arrivals from these two stations are included.

We investigated the effect that structure outside of the region might have on our tomogram, namely subducted Pacific lithosphere under the North Island, or structure at the bounce point of the \( PP \) ray paths. High speeds associated with subducted Pacific lithosphere are a likely consequence of low temperatures in the slab, but could also be enhanced by anisotropy within or adjacent to the slab. To test the influence of rays through such high-speed zones, we removed arrival times from
Figure 3.10: A squeeze test demonstrating that the travel-time data require high speeds below a depth of 330 km under the northwestern South Island. A-B) Results of inverting the observed travel-time residuals produced by squeezing the travel-time residuals to less than 330 km depth; C-D) results of inverting the predicted residuals from A and B with a relaxed depth constraint; E-F) tomograms produced from adding the squeezed (A-B) and relaxed (C-D) travel-time residuals.
the 3 earthquakes occurring near Tonga (backazimuths from 10° - 30° in Figure 3.3) and also the 3 earthquakes that produced the PP ray paths. Eliminating travel-times from these events does not affect the location of the major high- or low-speed features discussed previously (Figure 3.11), but the lateral extent of the high-speed body in the northwest becomes smaller, and the core of the region with +4.5% $V_p$ does not reach as deep. These two tests (Figures 3.11 and D.6) suggest that although anisotropy and the passage of waves through high-speed regions outside the model space may contribute to the magnitude of high-speed anomalies in this body, they cannot account for all the observed variation in travel-time data. We infer that the high-speed zones beneath the northern and southern parts of the South Island and the low-speed zone under its east coast remain robust features required by the arrival times.

Finally, we compared our data set to previous studies [e.g., Fry et al., 2014; Kohler and Eberhart-Phillips, 2002; Stern et al., 2000] by removing the OBSs and inverting for a velocity structure using arrival times only from the land stations (Figure D.7). Note that we did not incorporate data used in these earlier investigations into this inversion in order to see if these features can be resolved using an independent data set. The major features from the previous studies, including the high speeds in the northwestern part of the island, the low speeds along the east coast around Christchurch and Dunedin, and the high-speed feature in the southwest, are all evident in the tomogram presented here. The high-speed zone in the northwestern part of the South Island and offshore is poorly defined below ~200 km, compared with tomograms using OBSs (Figures 3.7, 3.8, 3.10, 3.11 and D.6). A major, and perhaps somewhat surprising, difference is the lack of a high-speed zone underneath the Southern Alps in this tomogram (near 44°S, 170°E), imaged in earlier investigations [e.g., Fry et al., 2014; Kohler and Eberhart-Phillips, 2002; Stern et al., 2000] and seen in the tomogram presented here only weakly, with anomalies of +1% or less. We return to this lack of high speeds imaged below the mountains later.
Figure 3.11: Depth slices at 60, 90, 125, 165, 205, 245, 285, 330, 375, 420, and 465 km of a $V_p$ tomogram for which earthquakes from near Tonga and the $PP$ events were excluded. Eliminating these events tests the influence of paths that traverse the Tonga subduction zones and structure near $PP$ bounce points. Compare this with Figure 3.7.
3.4 Discussion

The low-speed features around both Christchurch and Dunedin (Figures 3.7 and 3.8) coincide with regions of elevated heat flow (70 - 74 mW m$^{-2}$ near Christchurch and 90 - 92 mW m$^{-2}$ near Dunedin), as well as regions that experienced Miocene and younger volcanism (≈ 6 Ma near Christchurch and ≈ 10 Ma near Dunedin; Figure 3.1) [Godfrey et al., 2001, and references therein]. The high heat flow, especially in the Dunedin region, implies a thinned lithosphere [Godfrey et al., 2001], suggesting that the low-speed region is anomalously warm compared to that under the South Island where the lithosphere is thicker. Other studies, such as those based on geochemical investigations that invoke lithospheric removal to explain the composition and origin of the Cenozoic intraplate volcanism [Hoernle et al., 2006; Timm et al., 2009], have proposed a thinned lithosphere underneath Christchurch and Dunedin.

The high-speed feature imaged in the southwestern South Island results from subduction of Australian lithosphere beneath the Pacific plate at the Puysegur trench. The nearly vertical high-speed zone extends to depths of ≈ 200 km (Figure 3.7a-e and Figure 3.12). This is consistent with the tomogram of Eberhart-Philips and Reyners [2001], as well as studies of seismicity that show a nearly vertical seismic zone [e.g., Anderson and Webb, 1994].

The lack of high speeds underneath the Southern Alps, imaged here as +1% or less, is unexpected. Teleseismic $P$-wave delays measured by Stern et al. [2000] suggest an approximately 100 km wide high-speed body with speeds up to 7% fast centered at ≈ 120 km depth with a depth extent of about 100 km directly beneath the thickest crust below the Southern Alps. Kohler and Eberhart-Phillips [2002] found a 2 - 4% high-speed anomaly east of the Alpine fault underneath the mountains to no deeper than 200 km and varying in width from 60 to 100 km. From a combined study of surface waves and body waves, Fry et al. [2014] suggest high speeds in lithospheric mantle under the Southern Alps in a zone ≈ 100 km wide and no deeper than ≈ 125 km. Much of the signal in this region is likely overwhelmed by the higher speeds over larger regions to the north and west, where high-speed anomalies are ≈ 4 times greater than those under the Southern Alps (an
Figure 3.12: 3D plot of the $P$-wave tomogram showing the $+4\%$ $V_p$ isosurface. The bounding box has a depth extent of 350 km. Seismicity $> 40$ km from the USGS catalog of $M > 3$ since 1990 highlights the earthquakes related to each subducting slab. This shows the lack of deep seismicity under the northern South Island compared to the depth of the subducted lithospheric slab.
average of 4.5% fast versus 1% fast) and reach to greater depths. Plots of ray paths (Figure 3.6f) also show limited crossing ray coverage at depths less than 100 km, resulting in limited resolution at shallow depths. Since previous studies suggest that high speeds are not likely below about 150 - 200 km deep beneath the Southern Alps and are no wider than about 100 km, it is likely that our data cannot resolve such a feature. Our inability to image high speeds under the Southern Alps may also corroborate the findings of these previous studies that the high-speed zone beneath the Southern Alps does not extend deeper than about 150 - 200 km.

The dominant feature in the tomogram (Figures 3.7, 3.8 and 3.12) is the large, high-speed anomaly beneath and west of the northern South Island, interpreted here as subducted Pacific lithosphere. This feature connects to the southern end of subducted slab under the North Island. The nearly vertical high-speed zone is consistent with studies of seismicity at the northern end of the South Island [e.g., Anderson and Webb, 1994] that suggest that the slab bends towards the vertical below approximately 100 km depth, with earthquakes occurring to about 300 km depth under the North Island and to about 200 km depth under the northwestern South Island (Figure 3.12 and Figure 3.13a). The high-speed zone does appear southeast of the known location of seismicity (Figure 3.8a and Figure 3.12), suggesting that absolute motion of the Pacific plate (northwest in a hotspot reference frame [Gripp and Gordon, 2002]; Figure 3.1) overrode the down-going lithospheric slab. Furthermore, the depth to which at least 4.5% higher than reference speeds reach (∼ 450 km deep; Figure 3.8a and Figure 3.12) is greater than the deepest earthquakes in this region [e.g., Anderson and Webb, 1994]. As shown earlier, high speeds from structures such as the Tonga-Kermadec slab contribute to the spatial extent and magnitude of speed anomalies in this body; even when ray paths passing through Pacific lithosphere beneath the North Island or the Tonga-Kermadec slab are removed, however, a core with speeds at least 4.5% higher than average are still imaged to at least 250 km (Figure 3.11), and with speeds of +2 - 3% imaged to at least 400 km. This implies more subducted lithosphere than earthquakes suggest.

Sutherland [1995] showed that finite rotation of the Pacific, Australian, and Antarctic plates brings features on these plates into accord at 45 Ma, suggesting that relative motion between these
plates began at that time. The pole position for finite relative displacement of the Pacific and Australian plates for 45 Ma is at 49.8°S, 178.4°E (roughly 1000 km away from the northwestern end of the South Island; Figure 3.13a); the angle is 49.0 ± 3.0° [Sutherland, 1995]. The estimated amount of convergence (C) is then

\[ C = R\Theta \sin(\Phi), \quad (3.3) \]

where R is the radius of the Earth, Θ is the angle of rotation in radians, and Φ is the angle subtended by the arc length from the pole of rotation to the point of interest. From Equation (3.3), between 850 and 1000 km of Pacific plate lithosphere has been subducted obliquely southwestward beneath the northern end of the South Island since 45 Ma. For oblique, southwestward subduction, we estimate about 500 km of gently dipping subducted slab lies between the Hikurangi trench and the high speeds in the northwestern South Island (Figure 3.7), and about 600 km of gently dipping subducted slab lies between the trench and the high speeds off the central west coast of the island. The nearly horizontal segment of slab presumably reaches a depth of ~ 100 km. Oblique convergence of 850 - 1000 km suggests that the vertical extent of the subducted slab should reach 400 - 600 km depth under the northwestern South Island (Figure 3.13b), with increasing amounts of subduction and greater depths farther northeast, where greater convergence has occurred.

Under the northwestern South Island, earthquakes occur to about 200 km depth, but subducted lithosphere reaches about 400 - 450 km depth (Figure 3.8a). Subducted lithosphere off the central west coast of the South Island should reach depths of about 200 km (Figure 3.8b-c), but in a region where intermediate depth earthquakes do not occur. In general, intermediate depth earthquakes associated with subducting lithospheric slabs show fault plane solutions with down-dip extension, suggesting the slab descends into the asthenosphere under its own weight [Isacks and Molnar, 1971]. The greater depth of the high-speed zone than the deepest earthquakes implies that about 200 - 300 km of slab is needed below the deepest seismicity in order to produce intermediate depth earthquakes.
Figure 3.13: A) Location of the pole of rotation describing finite rotations of the Pacific and Australian plates since 45 Ma (black circle) [Sutherland, 1995], radial distance from the pole (black squares and dashed lines), and location of seismicity of M > 4 since 2009 from the USGS catalog. B) Plot of seismicity as a function of depth and distance from the 45 Ma pole of rotation. The dashed vertical lines highlight seismicity under the northern South Island. The solid black lines are the expected greatest depth of a subducted slab given the amount of oblique convergence, assuming 500 km (bottom line) or 600 km (top line) of slab lies nearly horizontal under New Zealand and that the slab turns vertical at a depth of 100 km.
3.5 Summary

We inverted teleseismic $P$-wave travel-time residuals measured on seismic stations both on and offshore the South Island of New Zealand to yield a $P$-wave tomogram reaching to a depth of 465 km. The inclusion of data from ocean bottom seismometers not only was essential for obtaining resolution for the offshore region, but also provided additional ray paths and increased resolution for the onshore parts of the region studied. A low $P$-wave speed feature along the east coast between the cities of Christchurch and Dunedin and extending to a depth of 150 - 200 km marks thinned lithosphere where Miocene and younger volcanism has occurred. Subduction beneath the southwestern South Island at the Puysegur trench is imaged to about 200 km depth. A lack of high speeds beneath the Southern Alps, imaged here as +1% or less, was unexpected and may suggest a shallow depth extent (< 150 km) of such high speeds detected by others with more dense networks in this region [Fry et al., 2014; Kohler and Eberhart-Phillips, 2002; Stern et al., 2000].

The dominant feature of the tomogram is a large extent of high speeds beneath the northwestern South Island and just west of it that we interpret as subducted Pacific lithosphere. High speeds, 4.5% above reference speeds, reach depths of about 450 km, and provide evidence for oblique westward subduction of Pacific lithosphere since about 45 Ma. The depth to which high-speed material reaches combined with the amount of subducted Pacific lithosphere compared to the depth of seismicity suggests that about 200 - 300 km of slab is required below the deepest seismicity to cause intermediate depth (subcrustal to 200 km) earthquakes.

3.6 Acknowledgements

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Investigation of mantle deformation due to oblique convergence from teleseismic S-wave tomography of South Island, New Zealand upper mantle

An S-wave tomogram produced from teleseismic travel-time measurements made on and offshore the South Island of New Zealand shows high speeds in the mantle under the northwestern South Island reaching to depths of 400 km and is consistent with oblique subduction of Pacific lithosphere since 45 Ma. A core of high speeds (7 - 10% faster than average) under the central Southern Alps associated with a convergent margin and mountain building appears ~ 150 km wide to no deeper than 150 km. The limited depth extent of the high-speed body suggests the lack of an unstable mantle drip (i.e., detached lithosphere), though the tomogram is consistent with accommodation of convergence via either lithospheric thickening or intracontinental subduction. Low-speed structures to depths of approximately 200 km along the east coast represent regions of thinned lithosphere that underlie Miocene and younger volcanism.

4.1 Introduction

Oblique convergence between the Australian and Pacific plates occurs beneath the South Island of New Zealand, with mountain building, thrust faulting and/or folding, and crustal thickening characterizing the surface and crustal response to this convergence; however, what is less apparent is how thickening or subduction processes deform the mantle lithosphere under the island. In the South Island the Alpine fault, a predominately dextral strike-slip fault running the length of the island, is the major tectonic feature of the area and absorbs the majority of motion between the
Australian and Pacific plates. Based on plate reconstructions, more than 800 km of displacement occurred between the plates over the past 45 Ma, with slip on the Alpine fault accommodating at least 450 km of this displacement [Sutherland, 1995, 1999; Sutherland et al., 2000] and possibly more than 700 km [Lamb et al., 2016]. During this time the relative motion between the Australian and Pacific plates appears to have changed little, though a component of convergence that initiated around 11 Ma [Cande and Stock, 2004] resulted in \( \sim 70 \) km of shortening across the South Island and the uplift of the Southern Alps. Geodetic measurements show around 40 mm/yr of right-lateral strike slip along the Alpine fault and around 10 mm/yr of convergence perpendicular to the fault [Beavan et al., 2002].

Subduction zones to the north and south of the South Island bookend the Alpine fault. To the north, oblique subduction of the Pacific plate beneath the Australian plate is of opposite polarity of subduction of the Australian plate beneath the Pacific plate in the south [Sutherland, 1995]. Convergence between the Australian and Pacific plates initiated around 45 Ma [Cande and Stock, 2004; Sutherland, 1995; Sutherland et al., 2000]. Seismicity under the North Island to depths of about 300 km and under the northern end of the South Island to depths of about 200 km suggests the subducting slab turns near-vertical at about 100 km depth [Anderson and Webb, 1994]. Subduction in the south is highly oblique and younger than in the north, initiating during the Miocene with about 170 km of plate subducted since this time [Eberhart-Phillips and Reyners, 2001; Sutherland, 1995].

Previous teleseismic studies found evidence for a nearly vertical, high seismic wave-speed anomaly under the central portion of the Southern Alps that may extend to at least 200 km deep [Kohler and Eberhart-Phillips, 2002; Stern et al., 2000]. These studies indicate that the high-speed anomaly is relatively symmetric and is either subvertical or dipping steeply to the west. Stern et al. [2000] summarized two possible interpretations for lithospheric accommodation of convergence underneath the South Island: intracontinental subduction or lithospheric thickening (Figure 4.1). The total shortening of \( \sim 70 \) km across the South Island suggests that if subduction of a slab occurred, the high-speed zone ought not reach deeper than \( \sim 150 - 200 \) km (roughly 70 km of
thickening of a 100 km thick lithosphere). Moreover, if a slab dipped northwestward, as many have suggested [e.g., Beaumont et al., 1996; Beavan et al., 1999], seismic stations off the central west coast of New Zealand should record early arrivals. Conversely, with only modest shortening, a high-speed anomaly at depths of 300 - 400 km would require that thickened mantle lithosphere had become unstable and stretched deeper into the mantle. Fry et al. [2014] inferred a combination of intracontinental subduction and lithospheric thickening accommodates shortening across the South Island along an older, mid-Cenozoic plate boundary.

Figure 4.1: Proposed mechanisms of accommodation of lithospheric shortening. A) A northwestward dipping intracontinental subduction zone [Wellman, 1979]; B) thickened mantle lithosphere as imagined by Houseman et al. [1981]; C) thickened mantle lithosphere that became unstable and developed into a drip. The grey region in each cartoon represents the high-speed zone that would be imaged in the tomography. Adapted from Stern et al. [2000].

A recent teleseismic P-wave study (Chapter 3) that used data from an onshore - offshore seismic experiment with station spacing around 100 km (larger station spacing than in the experiments described above) found a lack of high P-wave speeds in the central portion of the South Island under the Southern Alps, suggesting that high-speed structure associated with convergence does not extend deeper than 150 - 200 km [Zietlow et al., 2016a]. The significant finding from this study was the depth to which high-speed structure reached under the northwestern South Island and just
offshore the central west coast. Zietlow et al. [2016a] (Chapter 3) inferred high speeds to about 450 km deep in this region consistent with oblique westward subduction of Pacific lithosphere since 45 Ma. The depth to which high speeds reach combined with the known convergence across New Zealand suggests that about 200 - 300 km of slab is required below intermediate depth seismicity in order to produce sufficient weight to induce the intermediate depth earthquakes. In the study presented here, we analyze S-wave data recorded on the temporary deployment of ocean bottom seismometers used in Zietlow et al. [2016a] (Chapter 3) to allow broader resolution of lithosphere under the South Island and investigate deformation due to oblique convergence across the island.

4.2 Data and Methods

4.2.1 Data

Consisting of a one-year deployment from late January 2009 to early February 2010, the Marine Observations of Anisotropy Near Aotearoa (MOANA) experiment comprised 30 ocean bottom seismometers (OBSs) installed off both coasts of the South Island of New Zealand (Figure 4.2) which complemented the permanent New Zealand National Seismograph Network [Petersen et al., 2011]. The MOANA array had approximately 100 km station spacing with a full array aperture of over 900 km NW - SE, increasing the extent of the seismic network to over four times the width of the South Island. Seventeen stations (NZ05 - NZ21) were deployed to the west of the South Island on the Challenger Plateau and nine stations deployed to the east on the Campbell Plateau. Both plateaus are submerged continental crust of the Australian and Pacific plates, respectively. Four stations (NZ01 - NZ04) lay atop oceanic crust to the west of the South Island and south of the Challenger Plateau. Amongst the 30 OBSs deployed, station NZ01 did not yield usable seismic data and we did not recover station NZ17. To orient the horizontal components of the OBSs, we used Rayleigh wave polarizations [Stachnik et al., 2012]. In addition to the OBSs, the MOANA experiment also included four temporary broadband land stations (CASS, CROE, KELY, and FREW; Figure 4.2).
Figure 4.2: Location of seismographs used in this study. The light blue squares are ocean bottom seismometers associated with the MOANA array and the purple squares are the New Zealand National Seismograph Network maintained by GeoNet. Stations CASS, CROE, KELY, and FREW were temporary stations deployed by Victoria University of Wellington. Bathymetric features including the Campbell Plateau, Challenger Plateau, Chatham Rise, the Hikurangi Trench, and the Puysegur Trench, along with the cities of Christchurch and Dunedin, and the Alpine fault, are labeled for reference. Seismicity since 2009 of $m_b > 4$ and focal depth greater than 40 km highlights the seismic zones associated with each subducting slab. Red triangles mark locations of Miocene or younger volcanism. The thin black arrow shows current relative plate motion between the Pacific and Australian plates and the two thick black arrows show absolute plate motion of each plate in a hotspot reference frame [Gripp and Gordon, 2002]. From Zietlow et al. [2016a] (Chapter 3).
We examined seismograms (Figure 4.3) recorded across MOANA from all events of $m_b \geq 5.5$ with epicentral distances between $25^\circ - 100^\circ$ occurring from February 2009 to February 2010. Similar to other OBS teleseismic body wave studies [e.g., Wolfe et al., 2011], we examined these seismograms in multiple frequency bands, though found few usable events for frequencies greater than 0.12 Hz ($< \sim 8$ s). Of the seismograms examined, we observed clear $S$-wave arrivals on more than 50% of stations from 35 earthquakes in the filter band 0.05 - 0.1 Hz (20 - 10 s) and 12 earthquakes in the filter band 0.08 - 0.12 Hz (12.5 - 8.3 s). To help fill undersampled backazimuths and epicentral distances, we also examined the seismograms for clear $SKS$ arrivals and found 4 events in the band 0.05 - 0.1 Hz and 1 in the band 0.08 - 0.12 Hz.

Many of the observed earthquakes that yielded clear $S$-wave arrivals occurred in the subduction zones offshore Fiji-Tonga, Japan, and Sumatra. So as not to preferentially weight a particular backazimuth, we separated the earthquakes into $10^\circ$ backazimuth and $10^\circ$ distance bins and selected one earthquake from each bin. This yielded 25 earthquakes with usable picks in the band 0.05 - 0.1 Hz and 11 earthquakes in the band 0.08 - 0.12 Hz (Figure 4.4). In total, we measured 1321 $S$ and 213 $SKS$ arrival times (RMS $\sim 2.16$ s) using dbxcor, a waveform cross-correlation, beam-forming program (Figure G.1) [Pavlis and Vernon, 2010].

### 4.2.2 Corrections for Crust and Sediment

In teleseismic tomography, erroneous model structure may appear near the Moho and uppermost mantle without correction for crustal thickness [e.g., Humphreys and Clayton, 1990]. Given the amount of sediment known to exist either side of the South Island [e.g., Ball et al., 2014; Uenzelmann-Neben et al., 2009; Wood and Woodward, 2002; Wood et al., 2000], we also corrected for delays through low-speed sediment. To calculate crustal thickness underneath each station, we used recent $Pn$ travel-time measurements [Collins and Molnar, 2014] from earthquakes recorded across the MOANA array and the permanent land stations. At stations where $Pn$ measurements were not published (NZ02, NZ03, NZ04, APZ, DCZ, KHZ, NNZ, PYZ, and THZ), we assumed a crustal thickness given by Salmon et al. [2013]. We calculated sediment thickness from Whit-
Figure 4.3: Example waveforms unfiltered (left) and filtered from 0.05 - 0.1 Hz (right) recorded from an $m_b = 6.8$ earthquake offshore Java (backazimuth of $283^\circ$ and distance of $66^\circ$). Red bars denote OBSs.
Figure 4.4: A) Location and azimuthal distribution of earthquakes used in this S-wave tomography study. B) Azimuthal distribution of earthquakes used in this study. The radial axis is number of earthquakes. All events are $m_b > 5.5$ and occurred from 2009 - 2010.
taker et al. [2013]. Given the derived crust and sediment thicknesses under each station, we then calculated crust and sediment corrections (Table G.1) on a by-event basis assuming a shear wave speed in the mantle of 4.5 km/s and in the crust of 3.6 km/s for continental crust and 3.8 km/s for oceanic crust, and appropriate angles of incidence. Although shear wave speed in sediment varies greatly with depth [e.g., Hamilton, 1976; Ruan et al., 2014], we found that varying the shear wave speed in sediment between 50 m/s and 2 km/s did not greatly alter the final tomographic image of shear wave speeds. Thus, we assumed a shear wave speed in sediment of 1 km/s for the sediment correction.

4.2.3 Patterns of Travel-Time Residuals

After application of corrections for crust and sediment, patterns of teleseismic $S$-wave travel-time residuals reveal lateral variations in $S$-wave speeds (Figures 4.5 and G.2) and vary in magnitude between approximately +6 and -6 s. Recordings of earthquakes that occurred to the northwest of the South Island show early arrivals at stations on land, at OBSs just offshore the east coast, and at NZ14 and NZ15 compared to all other OBSs that record late arrivals (Figure 4.5a-b). In the northwestern part of the South Island, residuals are upwards of $\sim 5$ s fast (e.g., NNZ). This pattern suggests that relatively high speeds underlie the northwest part of the island, though rays to these stations passing directly through the Pacific slab beneath the North Island may also contribute to these observed early arrivals [e.g., Williams et al., 2013].

For earthquakes from the south (Figure 4.5c-d and Figure G.2a), residuals show late arrivals, ranging between +2 and +3 s, on most land stations as well as on the OBSs offshore the east coast relative to other stations. Most OBSs offshore the west coast show early arrivals. These early arrivals, with residuals $< -2$ s, at northwestern land stations and OBSs are again consistent with a high-speed zone beneath the northwestern part of the South Island and its offshore region. Residuals from earthquakes from the southwest (Figure G.2b) show a similar pattern to residuals from earthquakes to the south, though residuals in the south central region (e.g., EAZ and MLZ) record early arrivals instead of late.
Figure 4.5: Plot of measured travel-time residuals in the band 0.05 - 0.1 Hz after application of a crustal and sediment correction for 6 events from different backazimuths. Figure G.2 shows residuals measured in the band 0.08 - 0.12 Hz. Earthquakes A) near Fiji with backazimuth of 15° ($m_b = 6.6$); B) near Fiji with backazimuth of 22° ($m_b = 5.7$); C) south of Tierra del Fuego with backazimuth of 151° ($m_b = 6.0$); D) near the Sandwich Islands with backazimuth of 171° ($m_b = 6.2$); E) from offshore the Solomon Islands with backazimuth of 334° ($m_b = 6.3$); and F) from offshore Japan with backazimuth of 333° ($m_b = 6.5$). Black arrows point from the earthquake. A negative (positive) travel-time residual indicates an early (late) arrival.
OBSs deployed off the west coast record late arrivals compared to most land stations that record early arrivals for earthquakes from the northwest (Figure 4.5e-f and Figure G.2c-d). The residuals recorded on some eastern land stations (e.g., OPZ, SYZ, and TUZ) and the OBSs off the east coast differ depending on angle of incidence. $S$-waves from earthquakes occurring within approximately 60° of the island (Figure 4.5e and Figure G.2c-d) arrive early at the most eastern land stations and OBSs off the east coast, whereas those from earthquakes greater than 60° away arrive late (Figure 4.5f). The different distributions of residuals for different azimuths indicate a depth limit to the high speeds under the South Island. Residuals range between approximately +3 and -3 s from earthquakes less than 60° away and between approximately +5 and -5 s from earthquakes greater than 60° away.

4.2.4 Tomographic Method

We used the finite-frequency tomography code of Schmandt and Humphreys [2010] to invert the $S$-wave arrival-time measurements. This code uses sensitivity kernels to account for volumetric variations in wave speed based on 1D ray tracing in the AK135 reference model [Kennett et al., 1995]. It considers sensitivity only in the first Fresnel zone, calculated by an approximation of the Born theoretical “banana-doughnut” kernel of Dahlen et al. [2000]. The radius, $R$, of the first Fresnel zone can be approximated as

$$ R = \sqrt{\frac{V}{f}} \delta \frac{D - \delta}{D}, $$

where $V$ is wave speed, $f$ is frequency, $\delta$ is distance along the ray path, and $D$ is the total ray path length [Spetzler and Snieder, 2004]. Assuming a mantle $S$-wave speed of 4.5 km/s and a center frequency of 0.1 Hz (10 s), for a 5000 km long ray path (typical for an $S$-wave produced by a shallow earthquake 45° away) the radius of the first Fresnel zone would be about 70 km at 100 km depth and over 100 km at 300 km depth. Thus, resolution of features with lateral dimensions less than 100 km is not likely.

The region for which we performed our tomographic inversion ranges in latitude from approx-
imately 35°S to 51°S and in longitude from approximately 158°E to 177°W. The model structure extends in depth from 60 - 600 km. We did not invert for structure shallower than 60 km because we corrected the travel-time measurements for crustal thickness and sediment prior to the inversion. A trapezoidal mesh (that accounts for the curvature of the Earth) of 70 x 71 horizontal nodes and 13 vertical nodes (representing the number of depth slices) parameterizes the structure. This resulted in a node spacing of about 20 km in the center of the structure and about 30 - 35 km towards its edges. All node spacings are less than the approximate width of the first Fresnel zone sensitivity kernel (∼ 100 km at the top of the model and ∼ 300 km at the bottom) at the frequency band of the measured S-waves.

We applied both model norm damping and spatial smoothing to help regularize the inverse problem, which minimizes the cost function

\[ E = \| Am - d \|^2 + \gamma \| Lm \|^2 + \varepsilon \| m \|^2, \]  

(4.2)

where \( A \) is a matrix that contains the partial derivatives relating travel-time residuals to the model parameters, \( m \) is a vector containing the perturbations to the speed model, \( d \) are the differential travel-time data, \( \gamma \) is the smoothing parameter, \( L \) is the smoothing matrix where the weight of smoothing between neighboring nodes decreases with inverse distance between the nodes, and \( \varepsilon \) is the damping parameter [Schmandt and Humphreys, 2010]. We used a standard trade-off analysis between variance reduction and model norm to determine suitable damping and smoothing parameters (Figure G.3).

### 4.2.5 Model Resolution

Given our residual measurements and damping and smoothing parameters, we tested model resolution using synthetic hypothesis and checkerboard tests (Figures 4.6, 4.7, G.4 and G.5). To demonstrate the spatial coverage given our array geometry and travel-time data, we also show plots of the “hit quality” parameter (Figure 4.8) as depth slices ranging from 60 - 515 km. “Hit quality” is a measure of both hit count and backazimuth coverage ranging from 0 - 1, where a node
rated “1” is sampled by several rays from each backazimuth sextant [Schmandt and Humphreys, 2010]. Measurements of “hit quality” indicate that the on-land portion of our model, as well as approximately 300 km offshore either side of the central part of the South Island, is well sampled (i.e., hit quality > 0.8) to 465 km deep.

The synthetic checkerboard test (Figures G.4 and G.5) consisted of cubes approximately 200 km on a side with alternating positive and negative amplitudes of ±10% in a neutral background of the reference speeds at each depth, AK135. We chose the dimensions of the cubes to be approximately the width of the first Fresnel zone at 100 km depth. In general, with the same source locations and stations as in the inversion of real data and no added synthetic noise, the locations of the checkers are recovered, but the amplitudes of anomalies are underestimated. Maximum amplitude recovery ranges from approximately 20% to 88%, with recovered amplitudes greater than 50% limited to the top 165 km of the model. Amplitude recovery is approximately 30% or less below depths of 330 km.

Checkerboard tests are common to test model resolution, but are poor for identifying possible artifacts in the results of the inversion [e.g., Eilon et al., 2015] and typically represent the extreme limit of model resolution. Thus, we also determined expected model resolution of a 200-km-wide hypothetical structure representative of a high-speed anomaly spanning nearly the length of the western South Island and extending to 330 km (the depth below which the checkerboard test indicated resolution became less than 30%) or 565 km depth (Figures 4.6 and 4.7). We set the amplitude of the high-speed anomaly at +10% of the reference speed at each depth, imbedded in a background average of -0.02%. This test also included synthetic noise, approximated as a normal distribution with standard error equal to the RMS misfit of travel times for the best fitting model (RMS = 1.35 s) and assigned randomly to the arrival times. Typically, the recovered anomaly was larger in magnitude with the addition of synthetic noise compared to the no-noise case. In general, though, amplitudes are still underestimated.

For the slab to 330 km, the average recovery of speed magnitude for depths of 60 - 330 km is ~ 86% of the input speed anomaly (Figure 4.6 and Figure 4.7a-b). With deeper slabs extending
Figure 4.6: Resolution of synthetic input structure (+10% $V_s$ of the reference model, AK135) simulating vertical slabs of high-speed material with widths 200 km that extend from 60 km to 330 km (top panels) and from 60 km to 565 km (bottom panels). Synthetic travel times were calculated by tracing rays through the synthetic structures using the same station and event distribution as observed. Random noise of RMS = 1.35 s was added to the synthetic travel times. The synthetic travel times are then inverted for speed perturbation using the same method as with the measured arrival times. Depth slices at 90, 165, 245, 330, and 420 km are shown.
Figure 4.7: Cross-sections through the synthetic tomographic structure in Figure 4.6. A-B) Input and resolution of a 200-km-wide high-speed zone (+10% Vs of the reference model, AK135) extending to 330 km depth. C-D) Input and resolution of a 200 km wide high-speed zone extending to 565 km. E) Map showing location of the cross-sections.
Figure 4.8: Plots of the “hit quality” parameter at depth slices from 60 - 515 km demonstrating ray coverage. Cooler colors represent better resolved regions.
to 565 km (Figure 4.6 and Figure 4.7c-d), recovery over the depth range 60 - 330 km increased to an average of \( \sim 91\% \) of the input speed anomaly and did not fall below about 30\% until deeper than 465 km in the model. This test suggests that travel-times through deep structure could leak into shallower structure and enhance inferred speeds within the shallower structure.

The hypothesis tests of a slab extending to either 330 km or 565 km depth show that the width of the inferred high-speed zone increases with depth. This is likely the result of decreasing resolution with depth as well as changes in the radius of the first Fresnel zone with depth (Equation (4.1)). As we found above, the radius of the first Fresnel zone is \( \sim 70 \) km at 100 km depth and over 100 km at 300 km depth. Thus, we can resolve lateral variations in speed below about 350 km, though not as well as those at shallower depths, and due to the increasing width of Fresnel zones, lateral dimensions of resolvable structure will also broaden with depth.

### 4.3 Results

We show the \( S \)-wave tomogram of the mantle under the South Island of New Zealand in 11 depth slices ranging from 60 - 465 km (Figure 4.9) and five cross-sections, four traversing the central portion of the South Island perpendicular to the surface expression of the Alpine fault (Figure 4.10 and Figure 4.11a) and one traversing the South Island NE – SW parallel to the surface expression of the Alpine fault (Figure 4.11b). This model tomogram accounts for \( \sim 68\% \) of the \( \sim 4.7 \text{ s}^2 \) variance of the measured residuals. With an average error in the \( S \)-wave arrival time picks of \( \sim 0.9 \) s and an RMS travel-time residual of \( \sim 2.2 \) s, the best expected variance reduction is \( \sim 83\% \). Thus, the inferred structure explains much of the variance in the observed travel-time residuals.

In the mantle underneath the northwestern South Island, a nearly vertical high-speed body reaches depths of about 400 - 450 km (Figure 4.9 and Figure 4.10a-b). Speeds associated with this feature reach a maximum of at least 9 - 10\% higher than the reference model in the top 200 km and decrease in magnitude below this depth. High-speeds are also imaged in the upper mantle extending off the central west coast to a latitude of about 43\°S, though the high-speeds in this offshore region are weaker and shallower, reaching to only about 200 km depth. Given that the resolution tests
Figure 4.9: Depth slices through the S-wave tomogram using stations shown in Figure 4.2 and earthquakes shown in Figure 4.4 at 60, 90, 125, 165, 205, 245, 285, 330, 375, 420, and 465 km.
Figure 4.10: Cross-sections through the S-wave tomogram traversing the South Island perpendicular to the surface expression of the Alpine fault (A-C). Map inserts show the location of each cross-section. Black circles in A) mark earthquake hypocenters in New Zealand since 2009 with $m_b > 4$ and between 40°S and 42°S. We did not invert for structure shallower than 60 km.
Figure 4.11: Cross-sections through the S-wave tomogram and highlighting the high-speed structure underneath the Southern Alps.
suggest a best magnitude recovery of about 90%, high-speeds from this feature could likely be 10 - 11% faster than reference speeds. Such speeds are consistent with the observed travel-time data. For example, many arrival times measured from earthquakes with incidence angles at the Moho between $15^\circ$ – $30^\circ$ at station NNZ are upwards of 6 s early (Figure 4.12a). Assuming that a 6 s residual developed from passing through a 300 km thick layer at an incidence angle of $15^\circ$ (corresponding to a typical $S$-wave travel-time of $\sim 69$ s), the average speed must be about 9% faster than reference speeds. Underneath the southwestern South Island, we image another high-speed body in the upper mantle of about 6% higher than average speeds (Figure 4.9a-f), though only to depths of about 250 km.

A region with speeds approximately 5 to 8% low lies about 150 - 200 km deep to the west and south of the city of Christchurch (Figure 4.9a-e). Farther south near the city of Dunedin, a similar low-speed region also reaches depths of about 200 km in the upper mantle. Measured travel-time residuals at station ODZ (Figure 4.12d), located between Christchurch and Dunedin, show mostly late arrivals to the station, consistent with a low wave speed region beneath the east coast of the South Island. Off the west coast under the Challenger plateau, similar low speeds also exist, though they are less reliable since they are on the edge of the array.

The tomogram shows a high-speed feature approximately 7 - 9% faster than reference speeds that is stronger at shallow depths compared to deeper in the mantle under the Southern Alps centered near $45^\circ$S, $169^\circ$E (Figure 4.9a-c). The feature is approximately 150 km in width and observed to no deeper than about 150 km, and well defined by the observed travel-times at station WKZ (Figure 4.12g). Assuming an incidence angle of $30^\circ$ and a mantle speed of 4.5 km/s, an $S$-wave passing through a 150 km thick layer of 9% fast material yields a delay time of about 3.5 s, consistent with delays observed at WKZ. To ensure that this feature is not an artifact or the result of a poor crustal correction at station WKZ, we performed an inversion with travel-time measurements made at WKZ removed (Figure 4.13b). Although the magnitudes of the high speeds under the Southern Alps decreased with the removal of arrival times at WKZ, the travel times at other stations still require high speeds in this region to depths of 100 - 150 km. Including
Figure 4.12: Lower-hemisphere stereographic projections of observed (left column) and predicted (middle column) S-wave travel-time measurements and their difference (right column) plotted using angles of incidence at the Moho for stations NNZ (A - C), ODZ (D - F), WKZ (G - I), and NZ14 (J - L). Areas of the circles are proportional to magnitudes of travel-time anomalies.
travel-time measurements from this station in the inversion, but varying the crustal correction to account for error in the $Pn$ time term ($4.44 \pm 0.13$ s [Collins and Molnar, 2014] which maps to a crustal thickness of $42.8 \pm 1.3$ km beneath WKZ) also had little affect. To force high-speeds to not exist under the Southern Alps, we had to vary the crustal thickness at WKZ, as well as at the neighboring stations EAZ, LBZ, JCZ, MLZ, and MSZ, by about 10 km (Figure 4.13c). A 10 km change in crustal thickness requires a change in the measured $Pn$ time term of over 1 s, an error approximately 10 times that reported in Collins and Molnar [2014] at those stations.

Figure 4.11a shows the high-speed structure under the Southern Alps dipping to the northwest. To investigate whether this apparent dip is the result of ray coverage, we inserted a synthetic vertical structure of 10% faster than reference speeds to a 200 km depth under the Southern Alps (Figure 4.14). Resolution of such a synthetic structure shows a dip towards the northwest as well as smearing of high-speeds along this direction, similar to the dip and vertically smeared high-speeds seen in the tomogram (Figure 4.11a). Because of the vertical smearing, we investigated the minimum depth the model requires high speeds to reach under the Southern Alps. Plotting variance reduction as a function of model depth (Figure 4.15a) suggests that variance reduction does not significantly change (i.e., change by $>10\%$) when the model depth exceeds 285 km. Increases in variance reduction for model depths greater than about 300 km are likely due to better fitting the high speeds under the northwestern South Island. We then performed a squeeze test (Figure 4.15c-e). First, we forced all the variations in the observed travel times into the top 285 km of the model. We then differenced the predicted travel times produced by the squeezed inversion from the observed travel-time measurements and inverted the differenced travel times for a model with a relaxed depth constraint. Squeezing the speed anomalies into the top 285 km (Figure 4.15c) yields a structure similar to the inferred model (Figure 4.11a), with high speeds (i.e., $>5\%$ of the reference speed) under the Southern Alps reaching 200 - 250 km. Relaxing the squeezed condition (Figure 4.15e) reveals that these data do not require high-speed structure much deeper than about 200 km under the Southern Alps.

Observed travel-time residuals at stations in the offshore region of the central west coast
Figure 4.13: Cross-sections perpendicular to the surface expression of the Alpine fault illustrating the effect that station WKZ has on the high-speed structure under the Southern Alps. A) Cross-section as in Figure 4.11a; B) inversion with travel-time measurements at station WKZ removed; C) inversion with the crustal thickness at WKZ and surrounding stations decreased by 10km; D) map insert showing location of cross-sections.
Figure 4.14: A synthetic spike test of +10% the reference speeds at each depth to 200 km depth illustrating the resolution of high-speed structure under the Southern Alps. A) Input structure; B) output structure; C) map showing location of the cross-sections.
Figure 4.15: A squeeze test illustrating the minimum depth to which high speeds reach under the Southern Alps as required by the travel-time data. A) Variance reduction as a function of model depth; B) map showing location of the cross-sections; C) speed perturbations squeezed to 285 km; D) result of inverting the residuals from C with a relaxed depth constraint; and E) tomogram produced from adding the squeezed (C) and relaxed (D) travel-time residuals.
(e.g., NZ14 and NZ15) suggest that high speeds in the upper mantle do extend offshore. Almost all measured residuals at NZ14 (Figure 4.12j) and NZ15 are more than ~ 2.5 s early. Zietlow et al. [2014] (Chapter 2) found large SKS split times (> 2 s) in this region, with orientations of fast quasi-S waves trending approximately parallel to N60°E, suggesting anisotropy may enhance the high speeds observed extending off the central west coast of the South Island. To test if anisotropy possibly enhances the southwestward extent of this high-speed body, we removed stations NZ14 and NZ15, and therefore rays traveling directly through this anisotropic region, from the inversion. With (Figure 4.9) or without (Figure G.6) these stations, the nearly identical inferred structures show high speeds in the mantle under the northwestern South Island to depths below 350 km, as well as in the mantle offshore the central west coast.

We also investigated the effect that structure outside of the model region, such as subducted Pacific plate under the North Island, may have on the high-speed region in the northwest of the South Island. Rays passing through lithospheric slabs should arrive earlier at NZ14, NZ15, QRZ, NNZ, and other stations nearby. To do so, we removed travel-time residuals measured from earthquakes occurring near Tonga (backazimuths from 10° – 30° in Figure 4.4). Even after eliminating travel-time residuals from the Tonga events, high speeds remained in the upper mantle underneath the northwestern part of the South Island as well as offshore the central west coast, but the magnitude of the high-speed anomalies, especially below depths of about 250 km, decreased (Figure 4.16). These two tests (Figures 4.16 and G.6) suggest that although anisotropy and the passage of waves through high-speed regions outside the model space may contribute to the high-speed anomalies observed in the northwest, they can account for only a small fraction of the observed travel-time residuals.

4.4 Discussion

The 150 - 200 km deep approximately 5 to 8% low-speed features along the east coast near Christchurch and Dunedin (Figure 4.9a-e) coincide with regions that experienced Miocene and younger surface volcanism [Godfrey et al., 2001, and references therein]. The Akaroa and Lyttel-
Figure 4.16: Depth slices at 60, 90, 125, 165, 205, 245, 285, 330, 375, 420, and 465 km through the S-wave tomogram obtained without travel-time measurements made on earthquakes from near Tonga (backazimuth $\sim 30^\circ$). Eliminating these events tests the influence of ray paths that pass through high-speed subducted lithosphere beneath the North Island of New Zealand. Paths to NZ14, NZ15, QRZ, NNZ, and others nearby would pass through subducted lithospheric slab, which would contribute to early arrivals, but those to other offshore stations and those in the southern part of the South Island would avoid the high-speed subducted slab.
ton volcanoes near Christchurch last erupted $\sim$ 6 Ma, and the Otago volcano near Dunedin last erupted $\sim$ 10 Ma. Both regions also correspond to elevated heat flux, approximately 70 - 74 mW m$^{-2}$ around Christchurch and approximately 90 - 92 mW m$^{-2}$ around Dunedin [Godfrey et al., 2001]. The elevated heat flux and recent surface volcanism imply a thinned lithosphere underneath the east coast of the South Island, suggesting that the observed low-speed regions in this area are anomalously warm. Other studies also propose a thinned lithosphere under the east coast, namely those based on geochemical investigations that invoke lithospheric removal to explain the composition and origin of the Cenozoic intraplate volcanism [Hoernle et al., 2006; Timm et al., 2009].

In the southwestern South Island, observed high speeds result from subduction of Australian lithosphere beneath the Pacific plate at the Puysegur Trench. The nearly vertical high-speed zone extends to depths of $\sim$ 200 km (Figure 4.9a-e), consistent with the tomogram of Eberhart-Phillips and Regniers [2001], as well as with studies of seismicity that show a nearly vertical seismic zone [e.g., Anderson and Webb, 1994]. This is also consistent with the $P$-wave tomogram of Zietlow et al. [2016a] (Figure 4.17a-b versus Figure 4.17f-g; Chapter 3), though the region is not as well defined in width and depth in this study possibly because of poor ray coverage in this southwestern area compared to the $P$-wave study.

### 4.4.1 Subduction of Pacific Lithosphere

A significant finding from Zietlow et al. [2016a] (Chapter 3) was the depth to which high speeds, interpreted as marking subducted Pacific lithosphere, reached underneath the northwestern South Island to around 400 - 450 km. In this $S$-wave study, we again see evidence for near-vertical high-speeds to depths of at least 400 km in the mantle under the northwestern South Island (Figures 4.9 and 4.10a-b, Figures 4.17 to 4.19), even when ray paths passing through Pacific lithosphere beneath the North Island are removed. This again supports the inference of subducted Pacific lithosphere to depths of about 400 - 450 km beneath the northwestern portion of the South Island. The enhanced high speeds under stations CROE, NZ14, NZ15, and WVZ in the $S$-wave
Figure 4.17: Comparison of the $P$-wave tomogram of Zietlow et al. [2016a] (left column, A-E; Chapter 3) with the $S$-wave tomogram (right column, F-J).
Figure 4.18: Comparative cross-sections through the $P$-wave tomogram of Zietlow et al. [2016a] (left column, A-C; Chapter 3) and $S$-wave tomogram (right column, D-F). Cross-sections in A and D are in the same location as in Figure 4.10a, cross-sections in B and E are in the same location as in Figure 4.10b, and cross-sections in C and F are in the same location as in Figure 4.10c.
tomogram compared to the $P$-wave tomogram of Zietlow et al. [2016a] (Figure 4.17a-c versus Figure 4.17f-h) may be due to the inclusion of additional travel-time measurements on earthquakes not present in the $P$-wave study. For instance, SKS phases (incidence angles $< 15^\circ$ in Figure 4.12j) show travel-time residuals more than about 2.5 s fast. Ray paths from an additional earthquake occurring near Tonga (backazimuth of approximately 30$^\circ$ in Figure 4.12j) with travel-time residuals more than 3 s fast may also contribute.

Figure 4.19: 3D plot of the $P$-wave and $S$-wave tomograms showing the $+4\%$ $V_p$ isosurface (green) and $+10\%$ $V_s$ isosurface (teal). The bounding box has a depth extent of 350 km. Seismicity $> 40$ km from the USGS catalog of $M > 3$ since 1990 highlights the earthquakes related to each subducting slab. Plot generated using the UNAVCO IDV.

Pacific lithosphere to depths of about 400 km is consistent with estimates of finite rotations that show 850 - 1000 km of Pacific lithosphere has been obliquely subducted beneath the northwestern South Island since 45 Ma [Sutherland, 1995]. We estimate about 500 - 600 km of gently dipping subducted slab lies between the Hikurangi trench and the observed high speeds, and reach a depth of about 100 km. For oblique convergence of 850 - 1000 km, the vertical extent of the subducted slab should reach 400 - 600 km depth. Furthermore, as proposed by Zietlow et al. [2016a] (Chapter 3), the depth that high speeds reach in the $S$-wave tomogram below the location of intermediate depth (subcrustal to 200 km) seismicity suggests that about 200 - 300 km of slab
is needed below the deepest earthquakes to generate enough weight to produce the intermediate depth seismicity.

### 4.4.2 Lithospheric Accommodation of Convergence

Other studies document the high-speed structure under the Southern Alps (Figure 4.9a-c and Figure 4.11). *Stern et al.* [2000] suggested an approximately 100 km wide structure reaching to depths of \(~ 170\) km beneath the thickest crust, but not highest topography, of the Southern Alps from measurements of teleseismic *P*-wave delays. *Kohler and Eberhart-Phillips* [2002] inferred a similar high-speed structure 60 - 100 km wide to no deeper than 200 km beneath the central South Island. A combined analysis of body and surface waves suggests high speeds about 100 km in width reach no deeper than \(~ 125\) km in the lithospheric mantle [*Fry et al.,* 2014]. These are similar to the structure observed in this study, which appears no wider than approximately 150 km to about 150 km into the mantle. With about \(~ 70\) km of shortening across the South Island [*Cande and Stock*, 2004] and with a lithospheric mantle 100 km thick, the high-speed body should reach a depth of 150 - 200 km, as seen here and other studies. The squeeze test (Figure 4.15) indicates that high-speed structure under the Southern Alps deeper than about 200 km is not required by these travel-time data, a suggestion against an unstable drip sinking deep into the mantle (Figure 4.1c). Furthermore, travel-time residuals at station WKZ (Figure 4.12g) are consistent with a high-speed structure under the Southern Alps to no deeper than about 200 km. Measurements from earthquakes occurring to the northwest with incidence angles at the Moho of about 30° show residuals of about 3 s faster than reference speeds. Assuming an *S*-wave speed of 4.5 km/s, a 140 km thick 9% fast structure produces about a 3 s fast *S*-wave travel-time residual. Likewise, a 180 km thick 7% fast structure also produces a 3 s fast residual.

The shape and dip of the high-speed structure, as well as the observed travel-time measurements could help distinguish between the other modes of accommodation (Figure 4.1a versus Figure 4.1b). From synthetic tests we found that a vertical, high-speed structure will appear dipping to the northwest due to our ray geometry (Figure 4.14). Considering only the shallow core
(i.e., < 100 km deep) of at least 7% faster than reference speeds, the high-speed feature under the Southern Alps appears near-vertical. Furthermore, if a northwestward shallow dipping slab (i.e., dipping < 20°) existed beneath the South Island, stations along and off the west coast would record early arrivals from earthquakes arriving from the northwest. From such earthquakes (Figure 4.5e and Figure G.2c-d), OBS stations deployed off the west coast record late arrivals. On land, stations such as JCZ and MSZ also record late arrivals for earthquakes arriving from the northwest less than 60° away. Since stations like JCZ and MSZ record late arrivals, this suggests high speeds in the central portion of the South Island do not reach farther west than these stations and further indicates that this high-speed region is near-vertical, consistent with mantle accommodation of convergence via lithospheric thickening; however, 3D geodynamic models [e.g., Pysklywec et al., 2010] that model the strain rate field of mantle lithosphere with a non-linear viscous flow law show that bounding subduction of opposite polarity can bifurcate a thickening mantle lithosphere. Thus, a near-vertical high-speed region under the Southern Alps could also be consistent with an intra-continental subduction zone.

4.4.3 Mechanisms for Seismic Heterogeneity

The ratio between relative perturbations in \( V_p \) and \( V_s \), defined as

\[
\nu = \frac{\partial \ln V_s}{\partial \ln V_p},
\]

provides an estimate of the relative sensitivity of \( V_p \) and \( V_s \) to variations in temperature, composition, or the presence of partial melt. In general, higher values of \( \nu \) indicate a higher sensitivity in the shear modulus than the bulk modulus. Origins of seismic heterogeneity include variations in temperature, bulk composition, or partial melt [e.g., Goes et al., 2000]. High-temperature measurements of the elastic constants of olivine show \( \nu \sim 1.2 \) [Anderson et al., 1992]. Considering anelastic effects (when the relationship between stress and strain is time dependent, though such solids will return to their original shape given enough time), \( \nu \) increases, with 1.8 - 2.0 indicative that variations in temperature alone explain variations in seismic speed [Goes et al., 2000, and
From a seismic tomography study, Kennett et al. [1998] found a value of $\nu$ from approximately 1.9 - 2.2 in the top 500 km of the mantle. Thus, from both laboratory experiments and tomographic studies, $\nu$ could range from $\sim$ 1.2 - 2.2 when considering effects due to temperature alone. Values greater than this require an additional mechanism such as partial melt to explain variations in seismic speed.

A linear regression of the $P$-wave travel-time residuals of Zietlow et al. [2016a] (Chapter 3) and the $S$-wave travel-time residuals presented here produces a path integrated estimated of $\nu$ (Figure 4.20) [e.g., Schmandt and Humphreys, 2010]. Considering travel-time residuals from earthquakes and stations with both a $P$- and $S$-wave measurement, a regression that takes into account uncertainty in $P$-wave residuals and $S$-wave residuals yields a slope of $2.86 \pm 0.11$. Assuming an average $V_p/V_s$ ratio of 1.7 for the upper mantle under the South Island [Eberhart-Philips et al., 2014; Reyners et al., 2006], a slope of 2.86 suggests $\nu \sim 1.68$. This is consistent with seismic heterogeneity generated by only thermal effects; however, because residuals are a path integrated value and because $S$-waves tend to have longer wavelengths than $P$-waves, regression analysis of $P$- and $S$-wave residuals likely yields an underestimate of $\nu$ [Schmandt and Humphreys, 2010]. For instance, low speeds along the east coast of the South Island associated with volcanism and therefore some degree of partial melt could increase the value of $\nu$. Pervasive anisotropy observed in the mantle under the South Island [Collins and Molnar, 2014; Zietlow et al., 2014] may also alter the value of $\nu$.

4.5 Summary

We inverted teleseismic $S$-wave travel times at seismographs on and offshore the South Island of New Zealand to yield a 3D $S$-wave speed structure. Similar to the teleseismic $P$-wave tomogram of Zietlow et al. [2016a] (Chapter 3), a low $S$-wave speed feature along the east coast in the mantle to depths of 150 - 200 km indicates thinned lithosphere underlying Miocene and younger volcanism. Subduction at the Puysegur trench is imaged to about 200 km beneath the southwestern South Island. Beneath the Southern Alps, a high-speed region of about 7 – 10\% faster than average speeds
Figure 4.20: Plot of a regression of $P$-wave and $S$-wave travel-time residuals. The solid blue line has a slope of $2.86 \pm 0.11$. 
no wider than approximately 150 km reaches to about 150 km into the mantle. This suggests there is no unstable drip due to convergence across the South Island, though the tomogram is consistent with accommodation of convergence via either lithospheric thickening or intracontinental subduction. In the mantle underneath the northwestern South Island, high speeds of upwards of 10% higher than average continue to depths of about 400 km, and apparently mark Pacific lithosphere subducted obliquely westward since 45 Ma.

4.6 Acknowledgements

We thank Justin Ball, Cailey Condit, Danny Feucht, Jenny Nakai, Colin O’Rourke, Martha Savage, Vera Schulte-Pelkum, Steven Plescia Josh Stachnik, Laura Wallace, Stuart Wier, and especially Craig Jones for helpful discussions, Brandon Schmandt for sharing his tomography code and advising us on how to use it, and John Collins for leading the MOANA OBS experiment. We thank the captains and crews of the R/V Thomas G. Thompson (cruise TN229) in 2009 and the R/V Roger Revelle (cruise RR1002) in 2010. The Ocean Bottom Seismic data used in this research were provided by instruments from the Ocean Bottom Seismograph Instrument Pool (www.obsip.org), which is funded by the National Science Foundation. The IRIS PASSCAL program provided three of the land broadband stations. The OBSIP and PASSCAL data are available for download from the IRIS Data Management Center. The New Zealand GeoNet project and its sponsors EQC, GNS Science, and LINZ provided data. The figures were made with the Generic Mapping Tools (GMT) software by [Wessel and Smith, 1998]. This study was supported by the National Science Foundation under grant no. EAR-0409835, and in part by a CIRES Graduate Research Fellowship Award and by a gift from the Crafoord Foundation.
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Appendix  A

Orienting the Horizontal Components of Ocean Bottom Seismometers using $P$-Waves

During installation of a seismic sensor, it is common practice to align one of the horizontal components with true north. The remaining two components are thus aligned in the orthogonal horizontal direction and the vertical; however, such a practice is nearly impossible during the deployment of ocean bottom seismometers. Thus, before analysis of data collected on OBSs may begin, it is necessary to determine the orientation of the horizontal components.

For a correctly oriented seismic sensor and assuming no anisotropic media along the ray path, the amplitude of a direct $P$-wave arrival on the transverse ($T$) component should be near zero. A nonzero $P$-wave amplitude on the $T$ component is an indication of an improperly oriented sensor. To determine sensor orientation, we performed a grid search (Figure A.1) over the orientation angle (ranging from 0 – 180° clockwise from north) to identify the maximum amplitude ratio of the $P$-wave on the radial ($R$) component to that on the $T$ component. We also cross correlate the resulting $R$ component with the vertical component to resolve the 180° ambiguity from the grid search since only a half-space (0 – 180°) is used.

In this body-wave analysis, a total of 11 teleseismic events with epicentral distances ranging from 30 – 90° and from various backazimuths were used (Table A.1). The utilized time windows were selected by hand. The waveforms are band-pass filtered from 0.04 – 0.1 Hz after removal of the mean and trend. The orientations determined from this body-wave analysis compare well to those determined by a more robust method using Rayleigh waves (Figure A.2) [Stachnik et al.,
2012]. Final location and orientation information for the ocean bottom seismometers used in this study are summarized (Table A.2).

Figure A.1: Horizontal components of a seismic sensor (NZ16) before and after finding the orientation showing event 272. A) Horizontal component rotated to north. B) Orthogonal horizontal component rotated to east. C) Radial component. D) Transverse component.
The polarization analysis was also performed in the frequency band 0.02–0.04 Hz.

Mismeasurement of Rayleigh-wave arrival azimuth can possibly be caused by the presence of Rayleigh-polarized microseismic noise (ambient noise) of similar amplitude and greater cross correlation than the earthquake signal. Surface-wave magnitude $M_s$ is typically measured in the 18

It may be possible to use noise at stations to calibrate measurements.

ACKNOWLEDGMENTS

We thank Peter Molnar, Craig Jones, and Gabi Laske for helpful discussions. We thank captain and crew of the R/V Thomas G. Thompson (cruise TN229) in 2009, of the R/V Roger Revelle (cruise RR1002) in 2010, and the Scripps Institution of Oceanography Ocean Bottom Seismic Instrumentation Pool (OBSIP) in this field program were provided by the U.S. National Ocean Bottom Seismic Instrumentation Pool (http://www.obsip.org). The collection of seismic data was funded by the National Science Foundation under grants EAR-0409564, EAR-0409609, and EAR-0409835. J. Stachnik was funded by the Cooperative Institute for Research in Environmental Sciences (CIRES) program. The authors also thank editor Jonathan Lees and reviewer Neil Selby for constructive comments.

REFERENCES

Figure A.2: Comparison of sensor misorientation values determined by $P$-wave amplitude ratio with those determined by Rayleigh wave polarization analysis. Figure from Stachnik et al. [2012].
Table A.1: Event information for earthquakes used in $P$-wave orientation analysis.

<table>
<thead>
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<th>Name</th>
<th>Date</th>
<th>Lat (°)</th>
<th>Lon (°)</th>
<th>Depth (km)</th>
<th>Mw</th>
<th>BAZ (°)</th>
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Table A.2: Location and orientation information for the ocean bottom seismometers used in this study. “PW Orient” represents the horizontal orientation as found by the $P$-wave analysis and “RW Orient” represents the value of the horizontal orientation found via the Rayleigh wave analysis. Both values represent the azimuth to which the BH2 component is pointing instead of true north.

<table>
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<th>Station</th>
<th>Lat (°)</th>
<th>Lon (°)</th>
<th>Depth (m)</th>
<th>PW Orient (°)</th>
<th>Error (°)</th>
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</table>
Appendix B

Diagnostic Plots of Shear Wave Splitting from SplitLab

The following figures show diagnostic plots of individual shear wave splitting measurements made on each seismic station using the method of Silver and Chan [1991].

The top panel shows the horizontal components of the filtered waveforms (0.03 – 0.08 Hz) with analysis window depicted by grey shading (left) and a stereoplot of the results (right). Header information gives specifics of the event used in the analysis, as well as a summary of the shear wave splitting results from three techniques. Only the method of Silver and Chan [1991] is used in this study (called “minimum energy” in these plots). Assignment of quality was subjective, though was primarily based on signal-to-noise ratios. Waveforms with signal-to-noise ratios less than 5 were not considered.

The bottom panel shows the waveforms corrected first for delay time (left) and then for both delay time and orientation of the fast quasi-shear wave (center left), the particle motion before (dashed blue line) and after (solid red line) the effects due to splitting are removed (center right), and finally the resulting contours of transverse energy (right).

Events are labeled as year (Y), Julian day (J), and hour (H) in the form YYYY.JJJ.HH.
B.1 APZ

![Diagnostic plot of splitting measurement made on APZ for event 2006.130.02.]

Figure B.1: Diagnostic plot of splitting measurement made on APZ for event 2006.130.02.

![Diagnostic plot of splitting measurement made on APZ for event 2006.236.21.]

Figure B.2: Diagnostic plot of splitting measurement made on APZ for event 2006.236.21.

![Diagnostic plot of splitting measurement made on APZ for event 2006.274.09.]

Figure B.3: Diagnostic plot of splitting measurement made on APZ for event 2006.274.09.
Figure B.4: Diagnostic plot of splitting measurement made on APZ for event 2006.317.01.

Figure B.5: Diagnostic plot of splitting measurement made on APZ for event 2007.055.02.

Figure B.6: Diagnostic plot of splitting measurement made on APZ for event 2007.119.12.
Corrected Fast (··) & Slow (-)

Figure B.7: Diagnostic plot of splitting measurement made on APZ for event 2007.230.02.

Corrected Fast (··) & Slow (-)

Figure B.8: Diagnostic plot of splitting measurement made on APZ for event 2007.253.01.

Corrected Fast (··) & Slow (-)

Figure B.9: Diagnostic plot of splitting measurement made on APZ for event 2007.320.03.
Figure B.10: Diagnostic plot of splitting measurement made on APZ for event 2007.353.09.

Figure B.11: Diagnostic plot of splitting measurement made on APZ for event 2007.355.07.

Figure B.12: Diagnostic plot of splitting measurement made on APZ for event 2007.360.22.
Figure B.13: Diagnostic plot of splitting measurement made on APZ for event 2008.063.09.

Figure B.14: Diagnostic plot of splitting measurement made on APZ for event 2008.107.05.

Figure B.15: Diagnostic plot of splitting measurement made on APZ for event 2008.187.02.
Event: 24-Nov-2008 (329) 09:02   54.20N 154.32E  492km  Mw=7.3
       Station: APZ   Backazimuth: 351.9º   Distance: 111.7º
Rotation Correlation: 22º  34º <  60  0.200.444º   0.800.844º   0.80
Minimum Energy: -1º  44º <  80  0.200.444º   0.800.844º   0.80
Incipience: 46º  49º <  80  0.200.444º   0.800.844º   0.80
Filter: 0.030Hz - 0.08Hz
Init. Pol.:
Eigenvalue: 87º  <  87º   0.030Hz - 0.08Hz   0.80
Energy Map of T
Rotation-Correlation
Particle motion before (··) & after (-)
Corrected Fast (··) & Slow(-)
Corrected Q(··) & T(-)

Figure B.16: Diagnostic plot of splitting measurement made on APZ for event 2008.329.09.

Event: 17-Apr-2009 (107) 02:08  -19.58N -70.48E  25km  Mw=6.1
       Station: APZ   Backazimuth: 277.6º   Distance: 111.7º
Rotation Correlation: -80º  -90º <  80  0.300.404º   0.80
Minimum Energy: -1º  44º <  80  0.200.444º   0.800.844º   0.80
Incipience: -7º  46º <  80  0.200.444º   0.800.844º   0.80
Filter: 0.030Hz - 0.08Hz
Init. Pol.:
Eigenvalue: 119º  <  87º   0.030Hz - 0.08Hz   0.80
Energy Map of T
Rotation-Correlation
Particle motion before (··) & after (-)
Corrected Fast (··) & Slow(-)
Corrected Q(··) & T(-)

Figure B.17: Diagnostic plot of splitting measurement made on APZ for event 2009.040.14.

Event: 24-Nov-2008 (329) 09:02   54.20N 154.32E  492km  Mw=7.3
       Station: APZ   Backazimuth: 351.9º   Distance: 111.7º
Rotation Correlation: 22º  34º <  60  0.200.444º   0.800.844º   0.80
Minimum Energy: -1º  44º <  80  0.200.444º   0.800.844º   0.80
Incipience: 46º  49º <  80  0.200.444º   0.800.844º   0.80
Filter: 0.030Hz - 0.08Hz
Init. Pol.:
Eigenvalue: 87º  <  87º   0.030Hz - 0.08Hz   0.80
Energy Map of T
Rotation-Correlation
Particle motion before (··) & after (-)
Corrected Fast (··) & Slow(-)
Corrected Q(··) & T(-)

Figure B.18: Diagnostic plot of splitting measurement made on APZ for event 2009.107.02.
Figure B.19: Diagnostic plot of splitting measurement made on APZ for event 2009.111.05.

Figure B.20: Diagnostic plot of splitting measurement made on APZ for event 2009.187.14.

Figure B.21: Diagnostic plot of splitting measurement made on APZ for event 2009.264.08.
Figure B.22: Diagnostic plot of splitting measurement made on APZ for event 2009.286.05.

Figure B.23: Diagnostic plot of splitting measurement made on APZ for event 2009.317.03.

Figure B.24: Diagnostic plot of splitting measurement made on APZ for event 2010.144.16.
Figure B.25: Diagnostic plot of splitting measurement made on APZ for event 2010.199.05.

Figure B.26: Diagnostic plot of splitting measurement made on APZ for event 2010.216.12.

Figure B.27: Diagnostic plot of splitting measurement made on APZ for event 2010.224.11.
Event: 08-Oct-2010 (281) 03:26   51.37N -175.36E  19km  Mw=6.4
       Station: APZ   Backazimuth:  10.4º   Distance: ... 1 2 3 4sec
-90
-60
-30
0
30
60
90
-20 0 20 40 60 80
-1500
-1000
-500
0
500
1000
N
E
  Inc = 8.7º

Event: 01-Jan-2011 (001) 09:56  -26.79N -63.09E  577km  Mw=7.0
       Station: APZ   Backazimuth: 135.9º   Distance: ... 1 2 3 4sec
-90
-60
-30
0
30
60
90
-20 0 20 40 60 80
-1500
-1000
-500
0
500
1000
N
E
  Inc = 9.4º

Figure B.28: Diagnostic plot of splitting measurement made on APZ for event 2010.281.03.

Event: 23-Dec-2010 (357) 14:00   53.13N 171.16E  18km  Mw=6.4
       Station: APZ   Backazimuth:   1.9º   Distance: ... of T
0 1 2 3 4sec
-90
-60
-30
0
30
60
90
-20 0 20 40 60 80
-1500
-1000
-500
0
500
1000
N
E
  Inc = 8.6º

Figure B.29: Diagnostic plot of splitting measurement made on APZ for event 2010.357.14.

Figure B.30: Diagnostic plot of splitting measurement made on APZ for event 2011.001.09.
Figure B.31: Diagnostic plot of splitting measurement made on APZ for event 2011.171.16.

Figure B.32: Diagnostic plot of splitting measurement made on APZ for event 2011.301.18.

Figure B.33: Diagnostic plot of splitting measurement made on APZ for event 2011.326.18.
Figure B.34: Diagnostic plot of splitting measurement made on APZ for event 2012.080.18.

Figure B.35: Diagnostic plot of splitting measurement made on APZ for event 2012.102.22.

B.2 CRLZ

Figure B.36: Diagnostic plot of splitting measurement made on CRLZ for event 2003.076.16.
Figure B.37: Diagnostic plot of splitting measurement made on CRLZ for event 2003.166.19.

Figure B.38: Diagnostic plot of splitting measurement made on CRLZ for event 2003.167.22.

Figure B.39: Diagnostic plot of splitting measurement made on CRLZ for event 2003.174.12.
Figure B.40: Diagnostic plot of splitting measurement made on CRLZ for event 2004.162.15.

Figure B.41: Diagnostic plot of splitting measurement made on CRLZ for event 2004.283.21.

Figure B.42: Diagnostic plot of splitting measurement made on CRLZ for event 2004.320.09.
Figure B.43: Diagnostic plot of splitting measurement made on CRLZ for event 2005.080.12.

Figure B.44: Diagnostic plot of splitting measurement made on CRLZ for event 2005.269.01.

Figure B.45: Diagnostic plot of splitting measurement made on CRLZ for event 2006.023.20.
Figure B.46: Diagnostic plot of splitting measurement made on CRLZ for event 2008.187.02.

Figure B.47: Diagnostic plot of splitting measurement made on CRLZ for event 2008.329.09.

Figure B.48: Diagnostic plot of splitting measurement made on CRLZ for event 2009.097.04.
Figure B.49: Diagnostic plot of splitting measurement made on CRLZ for event 2009.123.16.

Figure B.50: Diagnostic plot of splitting measurement made on CRLZ for event 2009.187.14.

Figure B.51: Diagnostic plot of splitting measurement made on CRLZ for event 2009.222.19.
Figure B.52: Diagnostic plot of splitting measurement made on CRLZ for event 2009.286.05.

Figure B.53: Diagnostic plot of splitting measurement made on CRLZ for event 2011.001.09.

Figure B.54: Diagnostic plot of splitting measurement made on CRLZ for event 2011.175.03.
Figure B.55: Diagnostic plot of splitting measurement made on CRLZ for event 2012.080.18.

Figure B.56: Diagnostic plot of splitting measurement made on CRLZ for event 2012.102.22.

B.3 CTZ

Figure B.57: Diagnostic plot of splitting measurement made on CTZ for event 2007.353.09.
Figure B.58: Diagnostic plot of splitting measurement made on CTZ for event 2008.063.09.

Figure B.59: Diagnostic plot of splitting measurement made on CTZ for event 2008.141.13.

Figure B.60: Diagnostic plot of splitting measurement made on CTZ for event 2008.179.11.
Figure B.61: Diagnostic plot of splitting measurement made on CTZ for event 2008.180.12.

Figure B.62: Diagnostic plot of splitting measurement made on CTZ for event 2008.187.02.

Figure B.63: Diagnostic plot of splitting measurement made on CTZ for event 2008.329.09.
Figure B.64: Diagnostic plot of splitting measurement made on CTZ for event 2009.089.07.

Figure B.65: Diagnostic plot of splitting measurement made on CTZ for event 2009.097.04.

Figure B.66: Diagnostic plot of splitting measurement made on CTZ for event 2009.108.19.
Figure B.67: Diagnostic plot of splitting measurement made on CTZ for event 2009.222.19.

Figure B.68: Diagnostic plot of splitting measurement made on CTZ for event 2009.264.08.

Figure B.69: Diagnostic plot of splitting measurement made on CTZ for event 2009.286.05.
Figure B.70: Diagnostic plot of splitting measurement made on CTZ for event 2009.344.02.

Figure B.71: Diagnostic plot of splitting measurement made on CTZ for event 2010.049.01.

Figure B.72: Diagnostic plot of splitting measurement made on CTZ for event 2010.089.16.
Figure B.73: Diagnostic plot of splitting measurement made on CTZ for event 2010.144.16.

Figure B.74: Diagnostic plot of splitting measurement made on CTZ for event 2010.169.02.

Figure B.75: Diagnostic plot of splitting measurement made on CTZ for event 2010.216.12.
Figure B.76: Diagnostic plot of splitting measurement made on CTZ for event 2010.24611.

Figure B.77: Diagnostic plot of splitting measurement made on CTZ for event 2010.35714.

Figure B.78: Diagnostic plot of splitting measurement made on CTZ for event 2011.32618.
**Event:** 20-Mar-2012 (080) 18:02  16.50N -98.22E  20km  Mw=7.5
**Station:** CTZ   Backazimuth:  70.2º   Distance: 82.24º
**Inc:** 9.6º

**Event:** 30-Apr-2006 (120) 21:40  -27.21N -71.06E  12km  Mw=6.5
**Station:** DCZ   Backazimuth: 130.9º   Distance: 58.38º
**Inc:** 10.2º

**Event:** 22-Jun-2006 (173) 10:53   45.42N 149.34E  95km  Mw=6.0
**Station:** DCZ   Backazimuth: 347.6º   Distance: 92.24º
**Inc:** 9.8º

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Figure B.79: Diagnostic plot of splitting measurement made on CTZ for event 2012.080.18.

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**B.4 DCZ**

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Figure B.80: Diagnostic plot of splitting measurement made on DCZ for event 2006.120.21.

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Figure B.81: Diagnostic plot of splitting measurement made on DCZ for event 2006.173.10.
Figure B.82: Diagnostic plot of splitting measurement made on DCZ for event 2006.265.02.

Figure B.83: Diagnostic plot of splitting measurement made on DCZ for event 2006.293.10.

Figure B.84: Diagnostic plot of splitting measurement made on DCZ for event 2006.317.01.
Figure B.85: Diagnostic plot of splitting measurement made on DCZ for event 2007.055.02.

Figure B.86: Diagnostic plot of splitting measurement made on DCZ for event 2007.119.12.

Figure B.87: Diagnostic plot of splitting measurement made on DCZ for event 2007.227.23.
Figure B.88: Diagnostic plot of splitting measurement made on DCZ for event 2007.228.05.

Figure B.89: Diagnostic plot of splitting measurement made on DCZ for event 2008.063.09.

Figure B.90: Diagnostic plot of splitting measurement made on DCZ for event 2008.187.02.
Event: 17-Apr-2009 (107) 02:08 -19.58N -70.48E  25km  Mw=6.1
       Station: DCZ   Backazimuth: 126.8º   Distance: ...

Figure B.91: Diagnostic plot of splitting measurement made on DCZ for event 2009.107.02.

Event: 18-Apr-2009 (108) 19:17   46.01N 151.43E  35km  Mw=6.6
       Station: DCZ   Backazimuth: 349.1º   Distance: ...

Figure B.92: Diagnostic plot of splitting measurement made on DCZ for event 2009.108.19.

Event: 18-Apr-2009 (108) 19:17   46.01N 151.43E  35km  Mw=6.6
       Station: DCZ   Backazimuth: 349.1º   Distance: ...

Figure B.93: Diagnostic plot of splitting measurement made on DCZ for event 2009.264.08.
Figure B.94: Diagnostic plot of splitting measurement made on DCZ for event 2009.317.03.

Figure B.95: Diagnostic plot of splitting measurement made on DCZ for event 2010.144.16.

Figure B.96: Diagnostic plot of splitting measurement made on DCZ for event 2011.001.09.
Figure B.97: Diagnostic plot of splitting measurement made on DCZ for event 2011.171.16.

Figure B.98: Diagnostic plot of splitting measurement made on DCZ for event 2011.245.13.

Figure B.99: Diagnostic plot of splitting measurement made on DCZ for event 2011.301.18.
Figure B.100: Diagnostic plot of splitting measurement made on DCZ for event 2011.326.18.

B.5  DSZ

Figure B.101: Diagnostic plot of splitting measurement made on DSZ for event 2003.022.02.

Figure B.102: Diagnostic plot of splitting measurement made on DSZ for event 2004.162.15.
Figure B.103: Diagnostic plot of splitting measurement made on DSZ for event 2004.325.22.

Figure B.104: Diagnostic plot of splitting measurement made on DSZ for event 2005.165.17.

Figure B.105: Diagnostic plot of splitting measurement made on DSZ for event 2008.187.02.
Figure B.106: Diagnostic plot of splitting measurement made on DSZ for event 2009.286.20.

B.6 EAZ

Figure B.107: Diagnostic plot of splitting measurement made on EAZ for event 2004.320.09.

Figure B.108: Diagnostic plot of splitting measurement made on EAZ for event 2004.325.08.
Figure B.109: Diagnostic plot of splitting measurement made on EAZ for event 2004.333.18.

Figure B.110: Diagnostic plot of splitting measurement made on EAZ for event 2004.353.06.

Figure B.111: Diagnostic plot of splitting measurement made on EAZ for event 2005.080.12.
Figure B.112: Diagnostic plot of splitting measurement made on EAZ for event 2005.166.02.

Figure B.113: Diagnostic plot of splitting measurement made on EAZ for event 2006.053.22.

Figure B.114: Diagnostic plot of splitting measurement made on EAZ for event 2006.165.04.
Figure B.115: Diagnostic plot of splitting measurement made on EAZ for event 2006.236.21.

Figure B.116: Diagnostic plot of splitting measurement made on EAZ for event 2006.317.01.

Figure B.117: Diagnostic plot of splitting measurement made on EAZ for event 2007.119.12.
Figure B.118: Diagnostic plot of splitting measurement made on EAZ for event 2007.210.04.

Figure B.119: Diagnostic plot of splitting measurement made on EAZ for event 2007.214.03.

Figure B.120: Diagnostic plot of splitting measurement made on EAZ for event 2008.063.09.
Figure B.121: Diagnostic plot of splitting measurement made on EAZ for event 2008.187.02.

Figure B.122: Diagnostic plot of splitting measurement made on EAZ for event 2008.329.09.

Figure B.123: Diagnostic plot of splitting measurement made on EAZ for event 2009.123.16.
Figure B.124: Diagnostic plot of splitting measurement made on EAZ for event 2009.267.07.

Figure B.125: Diagnostic plot of splitting measurement made on EAZ for event 2009.286.05.

Figure B.126: Diagnostic plot of splitting measurement made on EAZ for event 2009.317.03.
Figure B.127: Diagnostic plot of splitting measurement made on EAZ for event 2009.321.15.

Figure B.128: Diagnostic plot of splitting measurement made on EAZ for event 2010.181.07.

Figure B.129: Diagnostic plot of splitting measurement made on EAZ for event 2010.199.05.
Figure B.130: Diagnostic plot of splitting measurement made on EAZ for event 2010.216.12.

Figure B.131: Diagnostic plot of splitting measurement made on EAZ for event 2011.001.09.

Figure B.132: Diagnostic plot of splitting measurement made on EAZ for event 2011.301.18.
Figure B.133: Diagnostic plot of splitting measurement made on EAZ for event 2011.326.18.

Figure B.134: Diagnostic plot of splitting measurement made on EAZ for event 2012.080.18.

Figure B.135: Diagnostic plot of splitting measurement made on EAZ for event 2012.102.22.
Figure B.136: Diagnostic plot of splitting measurement made on FOZ for event 2004.320.09.

Figure B.137: Diagnostic plot of splitting measurement made on FOZ for event 2004.353.06.

Figure B.138: Diagnostic plot of splitting measurement made on FOZ for event 2005.080.12.
Figure B.139: Diagnostic plot of splitting measurement made on FOZ for event 2005.165.17.

Figure B.140: Diagnostic plot of splitting measurement made on FOZ for event 2005.269.01.

Figure B.141: Diagnostic plot of splitting measurement made on FOZ for event 2006.053.22.
Figure B.142: Diagnostic plot of splitting measurement made on FOZ for event 2006.165.04.

Figure B.143: Diagnostic plot of splitting measurement made on FOZ for event 2006.317.01.

Figure B.144: Diagnostic plot of splitting measurement made on FOZ for event 2007.119.12.
Figure B.145: Diagnostic plot of splitting measurement made on FOZ for event 2007.196.13.

Figure B.146: Diagnostic plot of splitting measurement made on FOZ for event 2007.353.09.

Figure B.147: Diagnostic plot of splitting measurement made on FOZ for event 2007.360.22.
Figure B.148: Diagnostic plot of splitting measurement made on FOZ for event 2008.141.13.

Figure B.149: Diagnostic plot of splitting measurement made on FOZ for event 2008.187.02.

Figure B.150: Diagnostic plot of splitting measurement made on FOZ for event 2008.329.09.
Figure B.151: Diagnostic plot of splitting measurement made on FOZ for event 2009.286.05.

Figure B.152: Diagnostic plot of splitting measurement made on FOZ for event 2009.344.02.

Figure B.153: Diagnostic plot of splitting measurement made on FOZ for event 2010.103.23.
Figure B.154: Diagnostic plot of splitting measurement made on FOZ for event 2010.144.16.

Figure B.155: Diagnostic plot of splitting measurement made on FOZ for event 2010.181.07.

Figure B.156: Diagnostic plot of splitting measurement made on FOZ for event 2010.216.12.
Figure B.157: Diagnostic plot of splitting measurement made on FOZ for event 2011.097.13.

Figure B.158: Diagnostic plot of splitting measurement made on FOZ for event 2011.279.11.

Figure B.159: Diagnostic plot of splitting measurement made on FOZ for event 2012.080.18.
Figure B.160: Diagnostic plot of splitting measurement made on FOZ for event 2012.102.22.

Figure B.161: Diagnostic plot of splitting measurement made on JCZ for event 2004.320.09.

Figure B.162: Diagnostic plot of splitting measurement made on JCZ for event 2004.325.22.
Figure B.163: Diagnostic plot of splitting measurement made on JCZ for event 2004.353.06.

Figure B.164: Diagnostic plot of splitting measurement made on JCZ for event 2005.165.17.

Figure B.165: Diagnostic plot of splitting measurement made on JCZ for event 2005.288.10.
Figure B.166: Diagnostic plot of splitting measurement made on JCZ for event 2006.130.02.

Figure B.167: Diagnostic plot of splitting measurement made on JCZ for event 2006.165.04.

Figure B.168: Diagnostic plot of splitting measurement made on JCZ for event 2006.317.01.
Figure B.169: Diagnostic plot of splitting measurement made on JCZ for event 2007.196.13.

Figure B.170: Diagnostic plot of splitting measurement made on JCZ for event 2007.246.16.

Figure B.171: Diagnostic plot of splitting measurement made on JCZ for event 2007.320.03.
Figure B.172: Diagnostic plot of splitting measurement made on JCZ for event 2007.353.09.

Figure B.173: Diagnostic plot of splitting measurement made on JCZ for event 2007.355.07.

Figure B.174: Diagnostic plot of splitting measurement made on JCZ for event 2008.187.02.
Figure B.175: Diagnostic plot of splitting measurement made on JCZ for event 2009.264.08.

Figure B.176: Diagnostic plot of splitting measurement made on JCZ for event 2009.286.05.

Figure B.177: Diagnostic plot of splitting measurement made on JCZ for event 2009.317.03.
Figure B.178: Diagnostic plot of splitting measurement made on JCZ for event 2010.181.07.

Figure B.179: Diagnostic plot of splitting measurement made on JCZ for event 2011.001.09.

Figure B.180: Diagnostic plot of splitting measurement made on JCZ for event 2011.326.18.
Figure B.181: Diagnostic plot of splitting measurement made on JCZ for event 2012.080.18.

Figure B.182: Diagnostic plot of splitting measurement made on JCZ for event 2012.102.22.

B.9 KHZ

Figure B.183: Diagnostic plot of splitting measurement made on KHZ for event 2003.265.04.
Figure B.184: Diagnostic plot of splitting measurement made on KHZ for event 2004.120.00.

Figure B.185: Diagnostic plot of splitting measurement made on KHZ for event 2004.162.15.

Figure B.186: Diagnostic plot of splitting measurement made on KHZ for event 2004.283.21.
Figure B.187: Diagnostic plot of splitting measurement made on KHZ for event 2004.325.08.

Figure B.188: Diagnostic plot of splitting measurement made on KHZ for event 2005.080.12.

Figure B.189: Diagnostic plot of splitting measurement made on KHZ for event 2005.164.22.
Figure B.190: Diagnostic plot of splitting measurement made on KHZ for event 2005.269.01.

Figure B.191: Diagnostic plot of splitting measurement made on KHZ for event 2005.339.12.

Figure B.192: Diagnostic plot of splitting measurement made on KHZ for event 2006.265.02.
Figure B.193: Diagnostic plot of splitting measurement made on KHZ for event 2006.317.01.

Figure B.194: Diagnostic plot of splitting measurement made on KHZ for event 2007.320.03.

Figure B.195: Diagnostic plot of splitting measurement made on KHZ for event 2007.360.22.
Figure B.196: Diagnostic plot of splitting measurement made on KHZ for event 2008.035.17.

Figure B.197: Diagnostic plot of splitting measurement made on KHZ for event 2008.239.21.

Figure B.198: Diagnostic plot of splitting measurement made on KHZ for event 2009.222.19.
Figure B.199: Diagnostic plot of splitting measurement made on KHZ for event 2009.286.20.

Figure B.200: Diagnostic plot of splitting measurement made on KHZ for event 2009.317.03.

Figure B.201: Diagnostic plot of splitting measurement made on KHZ for event 2009.318.19.
Figure B.202: Diagnostic plot of splitting measurement made on KHZ for event 2010.089.16.

Figure B.203: Diagnostic plot of splitting measurement made on KHZ for event 2010.144.16.

Figure B.204: Diagnostic plot of splitting measurement made on KHZ for event 2010.181.07.
Figure B.205: Diagnostic plot of splitting measurement made on KHZ for event 2011.001.09.

Figure B.206: Diagnostic plot of splitting measurement made on KHZ for event 2011.097.13.

Figure B.207: Diagnostic plot of splitting measurement made on KHZ for event 2011.326.18.
Figure B.208: Diagnostic plot of splitting measurement made on KHZ for event 2012.065.07.

Figure B.209: Diagnostic plot of splitting measurement made on KHZ for event 2012.080.18.

Figure B.210: Diagnostic plot of splitting measurement made on KHZ for event 2012.102.22.
Figure B.211: Diagnostic plot of splitting measurement made on LBZ for event 2004.283.21.

Figure B.212: Diagnostic plot of splitting measurement made on LBZ for event 2004.320.09.

Figure B.213: Diagnostic plot of splitting measurement made on LBZ for event 2004.325.22.
Figure B.214: Diagnostic plot of splitting measurement made on LBZ for event 2004.353.06.

Figure B.215: Diagnostic plot of splitting measurement made on LBZ for event 2005.165.17.

Figure B.216: Diagnostic plot of splitting measurement made on LBZ for event 2005.264.02.
Figure B.217: Diagnostic plot of splitting measurement made on LBZ for event 2005.288.10.

Figure B.218: Diagnostic plot of splitting measurement made on LBZ for event 2006.130.02.

Figure B.219: Diagnostic plot of splitting measurement made on LBZ for event 2006.173.10.
Figure B.220: Diagnostic plot of splitting measurement made on LBZ for event 2006.236.21.

Figure B.221: Diagnostic plot of splitting measurement made on LBZ for event 2006.274.09.

Figure B.222: Diagnostic plot of splitting measurement made on LBZ for event 2007.068.03.
Figure B.223: Diagnostic plot of splitting measurement made on LBZ for event 2007.150.20.

Figure B.224: Diagnostic plot of splitting measurement made on LBZ for event 2007.210.04.

Figure B.225: Diagnostic plot of splitting measurement made on LBZ for event 2007.246.16.
Figure B.226: Diagnostic plot of splitting measurement made on LBZ for event 2007.320.03.

Figure B.227: Diagnostic plot of splitting measurement made on LBZ for event 2007.355.07.

Figure B.228: Diagnostic plot of splitting measurement made on LBZ for event 2007.360.22.
Figure B.229: Diagnostic plot of splitting measurement made on LBZ for event 2008.063.09.

Figure B.230: Diagnostic plot of splitting measurement made on LBZ for event 2008.141.13.

Figure B.231: Diagnostic plot of splitting measurement made on LBZ for event 2008.187.02.
Figure B.232: Diagnostic plot of splitting measurement made on LBZ for event 2008.329.09.

Figure B.233: Diagnostic plot of splitting measurement made on LBZ for event 2009.097.04.

Figure B.234: Diagnostic plot of splitting measurement made on LBZ for event 2009.108.19.
Figure B.235: Diagnostic plot of splitting measurement made on LBZ for event 2009.111.05.

Figure B.236: Diagnostic plot of splitting measurement made on LBZ for event 2009.286.05.

Figure B.237: Diagnostic plot of splitting measurement made on LBZ for event 2009.344.02.
Figure B.238: Diagnostic plot of splitting measurement made on LBZ for event 2010.037.04.

Figure B.239: Diagnostic plot of splitting measurement made on LBZ for event 2010.103.23.

Figure B.240: Diagnostic plot of splitting measurement made on LBZ for event 2010.181.07.
Figure B.241: Diagnostic plot of splitting measurement made on LBZ for event 2010.216.12.

Figure B.242: Diagnostic plot of splitting measurement made on LBZ for event 2010.357.14.

Figure B.243: Diagnostic plot of splitting measurement made on LBZ for event 2011.001.09.
Figure B.244: Diagnostic plot of splitting measurement made on LBZ for event 2012.080.18.

Figure B.245: Diagnostic plot of splitting measurement made on LBZ for event 2012.102.22.

B.11 LTZ

Figure B.246: Diagnostic plot of splitting measurement made on LTZ for event 2004.162.15.
Figure B.247: Diagnostic plot of splitting measurement made on LTZ for event 2004.323.21.

Figure B.248: Diagnostic plot of splitting measurement made on LTZ for event 2004.320.09.

Figure B.249: Diagnostic plot of splitting measurement made on LTZ for event 2004.325.22.
Figure B.250: Diagnostic plot of splitting measurement made on LTZ for event 2006.053.22.

Figure B.251: Diagnostic plot of splitting measurement made on LTZ for event 2006.317.01.

Figure B.252: Diagnostic plot of splitting measurement made on LTZ for event 2007.119.12.
Figure B.253: Diagnostic plot of splitting measurement made on LTZ for event 2007.196.13.

Figure B.254: Diagnostic plot of splitting measurement made on LTZ for event 2007.320.03.

Figure B.255: Diagnostic plot of splitting measurement made on LTZ for event 2007.360.22.
Figure B.256: Diagnostic plot of splitting measurement made on LTZ for event 2008.107.05.

Figure B.257: Diagnostic plot of splitting measurement made on LTZ for event 2008.187.02.

Figure B.258: Diagnostic plot of splitting measurement made on LTZ for event 2008.329.09.
Figure B.259: Diagnostic plot of splitting measurement made on LTZ for event 2009.107.02.

Figure B.260: Diagnostic plot of splitting measurement made on LTZ for event 2009.123.16.

Figure B.261: Diagnostic plot of splitting measurement made on LTZ for event 2010.181.07.
Figure B.262: Diagnostic plot of splitting measurement made on LTZ for event 2010.224.11.

Figure B.263: Diagnostic plot of splitting measurement made on LTZ for event 2010.281.03.

Figure B.264: Diagnostic plot of splitting measurement made on LTZ for event 2011.097.13.
Figure B.265: Diagnostic plot of splitting measurement made on LTZ for event 2012.080.18.

Figure B.266: Diagnostic plot of splitting measurement made on LTZ for event 2012.102.22.

B.12 MLZ

Figure B.267: Diagnostic plot of splitting measurement made on MLZ for event 2004.162.15.
Figure B.268: Diagnostic plot of splitting measurement made on MLZ for event 2004.333.18.

Figure B.269: Diagnostic plot of splitting measurement made on MLZ for event 2004.353.06.

Figure B.270: Diagnostic plot of splitting measurement made on MLZ for event 2005.018.14.
Figure B.271: Diagnostic plot of splitting measurement made on MLZ for event 2005.080.12.

Figure B.272: Diagnostic plot of splitting measurement made on MLZ for event 2005.164.22.

Figure B.273: Diagnostic plot of splitting measurement made on MLZ for event 2005.165.17.
Figure B.274: Diagnostic plot of splitting measurement made on MLZ for event 2005.166.02.

Figure B.275: Diagnostic plot of splitting measurement made on MLZ for event 2005.339.12.

Figure B.276: Diagnostic plot of splitting measurement made on MLZ for event 2006.293.10.
Event: 13-Nov-2006 (317) 01:26  -26.04N -63.22E  552km  Mw=6.8
       Station: MLZ   Backazimuth: 135.3º   Distance: 9150

Event: 15-Aug-2007 (227) 23:40  -13.39N -76.60E  39km  Mw=8.0
       Station: MLZ   Backazimuth: 117.5º   Distance: 9795

Figure B.277: Diagnostic plot of splitting measurement made on MLZ for event 2006.317.01.

Event: 20-May-2007 (150) 20:22   52.14N 157.29E  116km  Mw=8.0

Figure B.278: Diagnostic plot of splitting measurement made on MLZ for event 2007.150.20.

Event: 15-Aug-2007 (227) 23:40  -13.39N -76.60E  39km  Mw=8.0

Figure B.279: Diagnostic plot of splitting measurement made on MLZ for event 2007.227.23.
Figure B.280: Diagnostic plot of splitting measurement made on MLZ for event 2008.187.02.

Figure B.281: Diagnostic plot of splitting measurement made on MLZ for event 2008.329.09.

Figure B.282: Diagnostic plot of splitting measurement made on MLZ for event 2010.144.16.
Figure B.283: Diagnostic plot of splitting measurement made on MLZ for event 2010.181.07.

Figure B.284: Diagnostic plot of splitting measurement made on MLZ for event 2010.357.14.

Figure B.285: Diagnostic plot of splitting measurement made on MLZ for event 2011.245.10.
Figure B.286: Diagnostic plot of splitting measurement made on MLZ for event 2011.301.18.

Figure B.287: Diagnostic plot of splitting measurement made on MLZ for event 2011.326.18.

Figure B.288: Diagnostic plot of splitting measurement made on MLZ for event 2012.065.07.
Figure B.289: Diagnostic plot of splitting measurement made on MLZ for event 2012.080.18.

Figure B.290: Diagnostic plot of splitting measurement made on MLZ for event 2012.102.22.

B.13 MQZ

Figure B.291: Diagnostic plot of splitting measurement made on MQZ for event 2003.022.02.
Figure B.292: Diagnostic plot of splitting measurement made on MQZ for event 2007.013.04.

Figure B.293: Diagnostic plot of splitting measurement made on MQZ for event 2008.187.02.

Figure B.294: Diagnostic plot of splitting measurement made on MQZ for event 2009.317.03.
Figure B.295: Diagnostic plot of splitting measurement made on MQZ for event 2010.144.16.

Figure B.296: Diagnostic plot of splitting measurement made on MQZ for event 2011.001.09.

Figure B.297: Diagnostic plot of splitting measurement made on MQZ for event 2011.171.16.
Figure B.298: Diagnostic plot of splitting measurement made on MQZ for event 2011.301.18.

Figure B.299: Diagnostic plot of splitting measurement made on MQZ for event 2012.080.18.

Figure B.300: Diagnostic plot of splitting measurement made on MQZ for event 2012.102.22.
Figure B.301: Diagnostic plot of splitting measurement made on MSZ for event 2006.274.09.

Figure B.302: Diagnostic plot of splitting measurement made on MSZ for event 2006.317.01.

Figure B.303: Diagnostic plot of splitting measurement made on MSZ for event 2007.150.20.
Figure B.304: Diagnostic plot of splitting measurement made on MSZ for event 2007.355.07.

Figure B.305: Diagnostic plot of splitting measurement made on MSZ for event 2008.187.02.

Figure B.306: Diagnostic plot of splitting measurement made on MSZ for event 2008.329.09.
Figure B.307: Diagnostic plot of splitting measurement made on MSZ for event 2009.123.16.

Figure B.308: Diagnostic plot of splitting measurement made on MSZ for event 2010.103.23.

Figure B.309: Diagnostic plot of splitting measurement made on MSZ for event 2010.181.07.
Figure B.310: Diagnostic plot of splitting measurement made on MSZ for event 2011.001.09.

Figure B.311: Diagnostic plot of splitting measurement made on MSZ for event 2012.102.22.

B.15 NNZ

Figure B.312: Diagnostic plot of splitting measurement made on NNZ for event 2003.265.04.
Figure B.313: Diagnostic plot of splitting measurement made on NNZ for event 2004.120.00.

Figure B.314: Diagnostic plot of splitting measurement made on NNZ for event 2004.162.15.

Figure B.315: Diagnostic plot of splitting measurement made on NNZ for event 2004.283.21.
Figure B.316: Diagnostic plot of splitting measurement made on NNZ for event 2004.320.09.

Figure B.317: Diagnostic plot of splitting measurement made on NNZ for event 2004.325.22.

Figure B.318: Diagnostic plot of splitting measurement made on NNZ for event 2005.165.17.
Figure B.319: Diagnostic plot of splitting measurement made on NNZ for event 2005.269.01.

Figure B.320: Diagnostic plot of splitting measurement made on NNZ for event 2007.360.22.

Figure B.321: Diagnostic plot of splitting measurement made on NNZ for event 2008.141.13.
Figure B.322: Diagnostic plot of splitting measurement made on NNZ for event 2008.187.02.

Figure B.323: Diagnostic plot of splitting measurement made on NNZ for event 2009.329.09.

Figure B.324: Diagnostic plot of splitting measurement made on NNZ for event 2009.040.14.
Figure B.325: Diagnostic plot of splitting measurement made on NNZ for event 2009.123.16.

Figure B.326: Diagnostic plot of splitting measurement made on NNZ for event 2009.264.08.

Figure B.327: Diagnostic plot of splitting measurement made on NNZ for event 2009.286.05.
Figure B.328: Diagnostic plot of splitting measurement made on NNZ for event 2010.103.23.

Figure B.329: Diagnostic plot of splitting measurement made on NNZ for event 2010.281.03.

Figure B.330: Diagnostic plot of splitting measurement made on NNZ for event 2010.357.14.
Figure B.331: Diagnostic plot of splitting measurement made on NNZ for event 2011.097.13.

Figure B.332: Diagnostic plot of splitting measurement made on NNZ for event 2011.345.01.

Figure B.333: Diagnostic plot of splitting measurement made on NNZ for event 2012.080.18.
Figure B.334: Diagnostic plot of splitting measurement made on NNZ for event 2012.102.22.

Figure B.335: Diagnostic plot of splitting measurement made on NZ02 for event 2009.286.05.

Figure B.336: Diagnostic plot of splitting measurement made on NZ02 for event 2009.317.03.
Figure B.337: Diagnostic plot of splitting measurement made on NZ02 for event 2009.318.19.

Figure B.338: Diagnostic plot of splitting measurement made on NZ02 for event 2010.010.00.

B.17 NZ03

Figure B.339: Diagnostic plot of splitting measurement made on NZ03 for event 2009.108.19.
Figure B.340: Diagnostic plot of splitting measurement made on NZ03 for event 2009.264.08.

Figure B.341: Diagnostic plot of splitting measurement made on NZ03 for event 2009.286.05.

Figure B.342: Diagnostic plot of splitting measurement made on NZ03 for event 2009.317.03.
Figure B.343: Diagnostic plot of splitting measurement made on NZ03 for event 2009.318.19.

Figure B.344: Diagnostic plot of splitting measurement made on NZ03 for event 2009.321.15.

Figure B.345: Diagnostic plot of splitting measurement made on NZ03 for event 2009.344.02.
Figure B.346: Diagnostic plot of splitting measurement made on NZ03 for event 2010.010.00.

B.18 NZ04

Figure B.347: Diagnostic plot of splitting measurement made on NZ04 for event 2009.123.16.

Figure B.348: Diagnostic plot of splitting measurement made on NZ04 for event 2009.264.08.
Figure B.349: Diagnostic plot of splitting measurement made on NZ04 for event 2009.286.05.

Figure B.350: Diagnostic plot of splitting measurement made on NZ04 for event 2009.317.03.

Figure B.351: Diagnostic plot of splitting measurement made on NZ04 for event 2009.318.19.
Figure B.352: Diagnostic plot of splitting measurement made on NZ04 for event 2010.010.00.

Figure B.353: Diagnostic plot of splitting measurement made on NZ05 for event 2009.108.19.

Figure B.354: Diagnostic plot of splitting measurement made on NZ05 for event 2009.286.05.
Figure B.355: Diagnostic plot of splitting measurement made on NZ05 for event 2009.317.03.

B.20 NZ08

Figure B.356: Diagnostic plot of splitting measurement made on NZ08 for event 2010.010.00.
Figure B.357: Diagnostic plot of splitting measurement made on NZ09 for event 2009.286.05.

Figure B.358: Diagnostic plot of splitting measurement made on NZ09 for event 2009.317.03.

Figure B.359: Diagnostic plot of splitting measurement made on NZ09 for event 2009.321.15.
Figure B.360: Diagnostic plot of splitting measurement made on NZ09 for event 2009.344.02.

B.22 NZ10

Figure B.361: Diagnostic plot of splitting measurement made on NZ10 for event 2009.097.04.

Figure B.362: Diagnostic plot of splitting measurement made on NZ10 for event 2009.267.07.
Figure B.363: Diagnostic plot of splitting measurement made on NZ10 for event 2009.318.19.

Figure B.364: Diagnostic plot of splitting measurement made on NZ10 for event 2009.344.02.

Figure B.365: Diagnostic plot of splitting measurement made on NZ10 for event 2010.010.00.
Figure B.366: Diagnostic plot of splitting measurement made on NZ11 for event 2009.097.04.

Figure B.367: Diagnostic plot of splitting measurement made on NZ11 for event 2009.317.03.

Figure B.368: Diagnostic plot of splitting measurement made on NZ11 for event 2009.344.02.
B.24 NZ12

Figure B.369: Diagnostic plot of splitting measurement made on NZ12 for event 2009.097.04.

Figure B.370: Diagnostic plot of splitting measurement made on NZ12 for event 2009.222.19.
Figure B.371: Diagnostic plot of splitting measurement made on NZ13 for event 2009.097.04.

Figure B.372: Diagnostic plot of splitting measurement made on NZ13 for event 2009.111.05.

Figure B.373: Diagnostic plot of splitting measurement made on NZ13 for event 2009.222.19.
Figure B.374: Diagnostic plot of splitting measurement made on NZ13 for event 2009.286.05.

B.26 NZ15

Figure B.375: Diagnostic plot of splitting measurement made on NZ15 for event 2009.097.04.

Figure B.376: Diagnostic plot of splitting measurement made on NZ15 for event 2009.108.19.
**Figure B.377:** Diagnostic plot of splitting measurement made on NZ16 for event 2009.123.16.

**Figure B.378:** Diagnostic plot of splitting measurement made on NZ16 for event 2009.222.19.

**Figure B.379:** Diagnostic plot of splitting measurement made on NZ16 for event 2009.286.05.
Figure B.380: Diagnostic plot of splitting measurement made on NZ16 for event 2009.321.15.

Figure B.381: Diagnostic plot of splitting measurement made on NZ18 for event 2009.123.16.

Figure B.382: Diagnostic plot of splitting measurement made on NZ18 for event 2009.222.19.
Figure B.383: Diagnostic plot of splitting measurement made on NZ18 for event 2009.264.08.

B.29 NZ20

Figure B.384: Diagnostic plot of splitting measurement made on NZ20 for event 2009.097.04.

Figure B.385: Diagnostic plot of splitting measurement made on NZ20 for event 2009.108.19.
Figure B.386: Diagnostic plot of splitting measurement made on NZ20 for event 2009.222.19.

Figure B.387: Diagnostic plot of splitting measurement made on NZ20 for event 2009.317.03.

B.30  NZ22

Figure B.388: Diagnostic plot of splitting measurement made on NZ22 for event 2009.222.19.
Figure B.389: Diagnostic plot of splitting measurement made on NZ23 for event 2009.222.19.

Figure B.390: Diagnostic plot of splitting measurement made on NZ24 for event 2009.097.04.
Figure B.391: Diagnostic plot of splitting measurement made on NZ24 for event 2009.107.02.

Figure B.392: Diagnostic plot of splitting measurement made on NZ24 for event 2009.123.16.

Figure B.393: Diagnostic plot of splitting measurement made on NZ24 for event 2009.286.05.
Figure B.394: Diagnostic plot of splitting measurement made on NZ25 for event 2009.089.07.

Figure B.395: Diagnostic plot of splitting measurement made on NZ25 for event 2009.097.04.

Figure B.396: Diagnostic plot of splitting measurement made on NZ25 for event 2009.108.19.
Figure B.397: Diagnostic plot of splitting measurement made on NZ25 for event 2009.111.05.

Figure B.398: Diagnostic plot of splitting measurement made on NZ25 for event 2009.222.19.

Figure B.399: Diagnostic plot of splitting measurement made on NZ25 for event 2009.317.03.
Figure B.400: Diagnostic plot of splitting measurement made on NZ25 for event 2009.344.02.

Figure B.401: Diagnostic plot of splitting measurement made on NZ25 for event 2009.358.00.

B.34 NZ26

Figure B.402: Diagnostic plot of splitting measurement made on NZ26 for event 2009.089.07.
Figure B.403: Diagnostic plot of splitting measurement made on NZ26 for event 2009.097.04.

Figure B.404: Diagnostic plot of splitting measurement made on NZ26 for event 2009.222.19.

Figure B.405: Diagnostic plot of splitting measurement made on NZ26 for event 2009.286.05.
Figure B.406: Diagnostic plot of splitting measurement made on NZ26 for event 2009.317.03.

B.35 NZ27

Figure B.407: Diagnostic plot of splitting measurement made on NZ27 for event 2009.097.04.

Figure B.408: Diagnostic plot of splitting measurement made on NZ27 for event 2009.108.19.
Event: 10-Aug-2009 (222) 19:55 14.10N 92.89E 5km Mw=7.5
Station: NZ27 Backazimuth: 286.6º Distance: 32.7º

Station: NZ27 Backazimuth: 122.3º Distance: 31.7º

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Figure B.409: Diagnostic plot of splitting measurement made on NZ27 for event 2009.222.19.

Event: 10-Aug-2009 (222) 19:55 14.10N 92.89E 5km Mw=7.5
Station: NZ27 Backazimuth: 286.6º Distance: 32.7º

Station: NZ27 Backazimuth: 122.3º Distance: 31.7º

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Figure B.410: Diagnostic plot of splitting measurement made on NZ27 for event 2009.317.03.

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B.36 NZ28

Event: 07-Apr-2009 (097) 04:23 46.05N 151.55E 31km Mw=6.9
Station: NZ28 Backazimuth: 346.1º Distance: 29.7º

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Figure B.411: Diagnostic plot of splitting measurement made on NZ28 for event 2009.097.04.
Figure B.412: Diagnostic plot of splitting measurement made on NZ28 for event 2009.286.05.

Figure B.413: Diagnostic plot of splitting measurement made on NZ28 for event 2009.317.03.

B.37 NZ29

Figure B.414: Diagnostic plot of splitting measurement made on NZ29 for event 2009.097.04.
Figure B.415: Diagnostic plot of splitting measurement made on NZ29 for event 2009.222.19.

Figure B.416: Diagnostic plot of splitting measurement made on NZ29 for event 2009.264.08.

Figure B.417: Diagnostic plot of splitting measurement made on NZ29 for event 2009.267.07.
Figure B.418: Diagnostic plot of splitting measurement made on NZ29 for event 2010.010.00.

B.38 NZ30

Figure B.419: Diagnostic plot of splitting measurement made on NZ30 for event 2009.123.16.

Figure B.420: Diagnostic plot of splitting measurement made on NZ30 for event 2009.222.19.
Figure B.421: Diagnostic plot of splitting measurement made on NZ30 for event 2009.317.03.

Figure B.422: Diagnostic plot of splitting measurement made on NZ30 for event 2009.318.19.

Figure B.423: Diagnostic plot of splitting measurement made on NZ30 for event 2009.358.00.
Figure B.424: Diagnostic plot of splitting measurement made on ODZ for event 2003.301.21.

Figure B.425: Diagnostic plot of splitting measurement made on ODZ for event 2004.180.09.

Figure B.426: Diagnostic plot of splitting measurement made on ODZ for event 2004.283.21.
Figure B.427: Diagnostic plot of splitting measurement made on ODZ for event 2004.320.09.

Figure B.428: Diagnostic plot of splitting measurement made on ODZ for event 2004.333.18.

Figure B.429: Diagnostic plot of splitting measurement made on ODZ for event 2004.353.06.
Figure B.430: Diagnostic plot of splitting measurement made on ODZ for event 2005.269.01.

Figure B.431: Diagnostic plot of splitting measurement made on ODZ for event 2005.288.10.

Figure B.432: Diagnostic plot of splitting measurement made on ODZ for event 2006.173.10.
Figure B.433: Diagnostic plot of splitting measurement made on ODZ for event 2006.293.10.

Figure B.434: Diagnostic plot of splitting measurement made on ODZ for event 2007.055.02.

Figure B.435: Diagnostic plot of splitting measurement made on ODZ for event 2007.068.03.
Figure B.436: Diagnostic plot of splitting measurement made on ODZ for event 2007.246.16.

Figure B.437: Diagnostic plot of splitting measurement made on ODZ for event 2007.253.01.

Figure B.438: Diagnostic plot of splitting measurement made on ODZ for event 2007.320.03.
Figure B.439: Diagnostic plot of splitting measurement made on ODZ for event 2008.063.09.

Figure B.440: Diagnostic plot of splitting measurement made on ODZ for event 2009.040.14.

Figure B.441: Diagnostic plot of splitting measurement made on ODZ for event 2009.046.10.
Figure B.442: Diagnostic plot of splitting measurement made on ODZ for event 2009.108.19.

Figure B.443: Diagnostic plot of splitting measurement made on ODZ for event 2009.264.08.

Figure B.444: Diagnostic plot of splitting measurement made on ODZ for event 2009.267.07.
Figure B.445: Diagnostic plot of splitting measurement made on ODZ for event 2009.286.05.

Figure B.446: Diagnostic plot of splitting measurement made on ODZ for event 2009.344.02.

Figure B.447: Diagnostic plot of splitting measurement made on ODZ for event 2010.037.04.
Figure B.448: Diagnostic plot of splitting measurement made on ODZ for event 2010.103.23.

Figure B.449: Diagnostic plot of splitting measurement made on ODZ for event 2010.126.02.

Figure B.450: Diagnostic plot of splitting measurement made on ODZ for event 2010.169.02.
Figure B.451: Diagnostic plot of splitting measurement made on ODZ for event 2010.181.07.

Figure B.452: Diagnostic plot of splitting measurement made on ODZ for event 2011.245.10.

Figure B.453: Diagnostic plot of splitting measurement made on ODZ for event 2011.301.18.
Figure B.454: Diagnostic plot of splitting measurement made on ODZ for event 2012.080.18.

Figure B.455: Diagnostic plot of splitting measurement made on ODZ for event 2012.102.22.

B.40 OPZ

Figure B.456: Diagnostic plot of splitting measurement made on OPZ for event 2009.097.04.
Figure B.457: Diagnostic plot of splitting measurement made on OPZ for event 2009.264.08.

Figure B.458: Diagnostic plot of splitting measurement made on OPZ for event 2009.286.05.

Figure B.459: Diagnostic plot of splitting measurement made on OPZ for event 2009.317.03.
Figure B.460: Diagnostic plot of splitting measurement made on OPZ for event 2009.321.15.

Figure B.461: Diagnostic plot of splitting measurement made on OPZ for event 2010.037.04.

Figure B.462: Diagnostic plot of splitting measurement made on OPZ for event 2010.089.16.
Figure B.463: Diagnostic plot of splitting measurement made on OPZ for event 2010.181.07.

Figure B.464: Diagnostic plot of splitting measurement made on OPZ for event 2010.199.05.

Figure B.465: Diagnostic plot of splitting measurement made on OPZ for event 2010.224.11.
Figure B.466: Diagnostic plot of splitting measurement made on OPZ for event 2010.246.11.

Figure B.467: Diagnostic plot of splitting measurement made on OPZ for event 2011.001.09.

Figure B.468: Diagnostic plot of splitting measurement made on OPZ for event 2012.102.22.
B.41 OXZ

Figure B.469: Diagnostic plot of splitting measurement made on OXZ for event 2009.097.04.

Figure B.470: Diagnostic plot of splitting measurement made on OXZ for event 2009.107.02.

Figure B.471: Diagnostic plot of splitting measurement made on OXZ for event 2009.108.19.
Event: 10-Aug-2009 (222) 19:55  14.10N  92.89E  5km  Mw=7.5
Station: OXZ   Backazimuth: 287.6º   Distance:...

Figure B.472: Diagnostic plot of splitting measurement made on OXZ for event 2009.222.19.

Event: 13-Oct-2009 (286) 05:37   52.75N -167.00E  24km  Mw=6.4
Station: OXZ   Backazimuth: 12.6º   Distance:...

Figure B.473: Diagnostic plot of splitting measurement made on OXZ for event 2009.286.05.

Station: OXZ   Backazimuth: 123.0º   Distance:...

Figure B.474: Diagnostic plot of splitting measurement made on OXZ for event 2009.317.03.
Figure B.475: Diagnostic plot of splitting measurement made on OXZ for event 2010.012.21.

Figure B.476: Diagnostic plot of splitting measurement made on OXZ for event 2010.181.07.

Figure B.477: Diagnostic plot of splitting measurement made on OXZ for event 2010.216.12.
Figure B.478: Diagnostic plot of splitting measurement made on OXZ for event 2010.246.11.

Figure B.479: Diagnostic plot of splitting measurement made on OXZ for event 2010.281.03.

Figure B.480: Diagnostic plot of splitting measurement made on OXZ for event 2011.001.09.
Figure B.481: Diagnostic plot of splitting measurement made on OXZ for event 2012.080.18.

Figure B.482: Diagnostic plot of splitting measurement made on OXZ for event 2012.102.22.

B.42 PYZ

Figure B.483: Diagnostic plot of splitting measurement made on PYZ for event 2007.196.13.
Figure B.484: Diagnostic plot of splitting measurement made on PYZ for event 2007.320.03.

Figure B.485: Diagnostic plot of splitting measurement made on PYZ for event 2007.353.09.

Figure B.486: Diagnostic plot of splitting measurement made on PYZ for event 2007.355.07.
Figure B.487: Diagnostic plot of splitting measurement made on PYZ for event 2009.107.02.

Figure B.488: Diagnostic plot of splitting measurement made on PYZ for event 2009.286.20.

Figure B.489: Diagnostic plot of splitting measurement made on PYZ for event 2010.216.12.
Event: 12-Aug-2010 (224) 11:54   -1.27N -77.31E  207km  Mw=7.1
Station: PYZ   Backazimuth: 110.3º   Distance: ... 0 20 40 60 80
-4000
-3000
-2000
-1000
0
1000
2000
3000
N
E
Inc = 7.6º

Event: 20-Jun-2011 (171) 16:36  -21.70N -68.23E  128km  Mw=6.5
Station: PYZ   Backazimuth: 130.2º   Distance: ... Map of T
0 1 2 3 4sec
-90
-60
-30
0
30
60
90
-20 0 20 40 60 80
-1000
-500
0
500
1000
N
E
Inc = 9.2º

Figure B.490: Diagnostic plot of splitting measurement made on PYZ for event 2010.224.11.

Event: 20-Jun-2011 (171) 16:36  -21.70N -68.23E  128km  Mw=6.5
Station: PYZ   Backazimuth: 130.2º   Distance: ... Map of T
0 1 2 3 4sec
-90
-60
-30
0
30
60
90
-20 0 20 40 60 80
-1000
-500
0
500
1000
N
E
Inc = 9.2º

Figure B.491: Diagnostic plot of splitting measurement made on PYZ for event 2010.246.11.

Event: 20-Jun-2011 (171) 16:36  -21.70N -68.23E  128km  Mw=6.5
Station: PYZ   Backazimuth: 130.2º   Distance: ... Map of T
0 1 2 3 4sec
-90
-60
-30
0
30
60
90
-20 0 20 40 60 80
-1000
-500
0
500
1000
N
E
Inc = 9.2º

Figure B.492: Diagnostic plot of splitting measurement made on PYZ for event 2011.171.16.
Event: 28-Oct-2011 (301) 18:54  -14.44N -75.97E  24km  Mw=7.0
       Station: PYZ   Backazimuth: 119.9º   Distance: ...

Figure B.493: Diagnostic plot of splitting measurement made on PYZ for event 2011.301.18.

Event: 20-Mar-2012 (080) 18:02   16.50N -98.22E  20km  Mw=7.5
       Station: PYZ   Backazimuth:  81.9º   Distance: ...

Figure B.494: Diagnostic plot of splitting measurement made on PYZ for event 2012.080.18.

Event: 11-Apr-2012 (102) 22:55   18.23N -102.69E  20km  Mw=6.7
       Station: PYZ   Backazimuth:  77.6º   Distance: ...

Figure B.495: Diagnostic plot of splitting measurement made on PYZ for event 2012.102.22.
Figure B.496: Diagnostic plot of splitting measurement made on QRZ for event 2003.339.21.

Figure B.497: Diagnostic plot of splitting measurement made on QRZ for event 2004.162.15.

Figure B.498: Diagnostic plot of splitting measurement made on QRZ for event 2004.283.21.
Figure B.499: Diagnostic plot of splitting measurement made on QRZ for event 2004.320.09.

Figure B.500: Diagnostic plot of splitting measurement made on QRZ for event 2005.165.17.

Figure B.501: Diagnostic plot of splitting measurement made on QRZ for event 2005.269.01.
Figure B.502: Diagnostic plot of splitting measurement made on QRZ for event 2006.236.21.

Figure B.503: Diagnostic plot of splitting measurement made on QRZ for event 2007.055.02.

Figure B.504: Diagnostic plot of splitting measurement made on QRZ for event 2007.196.13.
Figure B.505: Diagnostic plot of splitting measurement made on QRZ for event 2007.210.04.

Figure B.506: Diagnostic plot of splitting measurement made on QRZ for event 2007.318.15.

Figure B.507: Diagnostic plot of splitting measurement made on QRZ for event 2008.106.22.
Figure B.508: Diagnostic plot of splitting measurement made on QRZ for event 2008.141.13.

Figure B.509: Diagnostic plot of splitting measurement made on QRZ for event 2008.187.02.

Figure B.510: Diagnostic plot of splitting measurement made on QRZ for event 2008.329.09.
Event: 21-Sep-2009 (264) 08:53   27.33N  91.44E  14km  Mw=6.1
       Station: QRZ   Backazimuth: 296.5º   Distance: 101.30º
Rotation Correlation: 66: 74º < 86   0.647.SN44.3
Minimum Energy: 34º 40º < 63   1.622.SN44.3
Supershear: 26º 63º < 69   0.647.SN44.3
Quality: fair   SNR 0.030Hz - 0.08Hz... Map of T

Event: 30-Jun-2010 (181) 07:22   16.40N -97.78E  20km  Mw=6.3
       Station: QRZ   Backazimuth: 77.3º   Distance: 100.41º
Rotation Correlation: 30<  40      1.4<1.8s<2.4
Minimum Energy: 35<  36      1.8<2.0s<2.4
Supershear: No Quality: SKS fair... Map of T

Event: 08-Oct-2010 (281) 03:26   51.37N -175.36E  19km  Mw=6.4
       Station: QRZ   Backazimuth: 7.5º   Distance: 100.69º
Rotation Correlation: 46: 69º < 66   1.622.SN44.3
Minimum Energy: 35º 40º < 66   1.622.SN44.3
Supershear: 46º 67º < 66   1.622.SN44.3
Quality: fair   SNR 0.030Hz - 0.08Hz... Map of T

Figure B.511: Diagnostic plot of splitting measurement made on QRZ for event 2009.264.08.

Figure B.512: Diagnostic plot of splitting measurement made on QRZ for event 2010.181.07.

Figure B.513: Diagnostic plot of splitting measurement made on QRZ for event 2010.281.03.
Figure B.514: Diagnostic plot of splitting measurement made on QRZ for event 2011.001.09.

Figure B.515: Diagnostic plot of splitting measurement made on QRZ for event 2012.080.18.

Figure B.516: Diagnostic plot of splitting measurement made on QRZ for event 2012.102.22.
Figure B.517: Diagnostic plot of splitting measurement made on RPZ for event 2003.050.03.

Figure B.518: Diagnostic plot of splitting measurement made on RPZ for event 2003.076.16.

Figure B.519: Diagnostic plot of splitting measurement made on RPZ for event 2003.166.19.
Figure B.520: Diagnostic plot of splitting measurement made on RPZ for event 2003.174.12.

Figure B.521: Diagnostic plot of splitting measurement made on RPZ for event 2003.265.04.

Figure B.522: Diagnostic plot of splitting measurement made on RPZ for event 2004.162.15.
Event: 09-Oct-2004 (283) 21:26   11.42N -86.67E  35km  Mw=6.9
Station: RPZ   Backazimuth: 90.1º   Distance: ...

Station: RPZ   Backazimuth: 132.2º   Distance: ...

Figure B.523: Diagnostic plot of splitting measurement made on RPZ for event 2004.283.21.

Figure B.524: Diagnostic plot of splitting measurement made on RPZ for event 2004.325.22.

Figure B.525: Diagnostic plot of splitting measurement made on RPZ for event 2005.080.12.
Figure B.526: Diagnostic plot of splitting measurement made on RPZ for event 2005.165.17.

Figure B.527: Diagnostic plot of splitting measurement made on RPZ for event 2006.053.22.

Figure B.528: Diagnostic plot of splitting measurement made on RPZ for event 2006.274.09.
Figure B.529: Diagnostic plot of splitting measurement made on RPZ for event 2006.317.01.

Figure B.530: Diagnostic plot of splitting measurement made on RPZ for event 2007.119.12.

Figure B.531: Diagnostic plot of splitting measurement made on RPZ for event 2007.150.20.
Figure B.532: Diagnostic plot of splitting measurement made on RPZ for event 2008.063.09.

Figure B.533: Diagnostic plot of splitting measurement made on RPZ for event 2008.187.02.

Figure B.534: Diagnostic plot of splitting measurement made on RPZ for event 2008.329.09.
Figure B.535: Diagnostic plot of splitting measurement made on RPZ for event 2009.040.14.

Figure B.536: Diagnostic plot of splitting measurement made on RPZ for event 2009.123.16.

Figure B.537: Diagnostic plot of splitting measurement made on RPZ for event 2009.286.20.
Figure B.538: Diagnostic plot of splitting measurement made on RPZ for event 2010.181.07.

Figure B.539: Diagnostic plot of splitting measurement made on RPZ for event 2010.216.12.

Figure B.540: Diagnostic plot of splitting measurement made on RPZ for event 2010.357.14.
Figure B.541: Diagnostic plot of splitting measurement made on RPZ for event 2011.001.09.

Figure B.542: Diagnostic plot of splitting measurement made on RPZ for event 2011.097.13.

Figure B.543: Diagnostic plot of splitting measurement made on RPZ for event 2011.175.03.
Figure B.544: Diagnostic plot of splitting measurement made on RPZ for event 2011.326.18.

Figure B.545: Diagnostic plot of splitting measurement made on RPZ for event 2012.080.18.

Figure B.546: Diagnostic plot of splitting measurement made on RPZ for event 2012.102.22.
Figure B.547: Diagnostic plot of splitting measurement made on SYZ for event 2007.055.02.

Figure B.548: Diagnostic plot of splitting measurement made on SYZ for event 2007.119.12.

Figure B.549: Diagnostic plot of splitting measurement made on SYZ for event 2007.196.13.
Figure B.550: Diagnostic plot of splitting measurement made on SYZ for event 2007.214.02.

Figure B.551: Diagnostic plot of splitting measurement made on SYZ for event 2007.320.03.

Figure B.552: Diagnostic plot of splitting measurement made on SYZ for event 2007.353.09.
Figure B.553: Diagnostic plot of splitting measurement made on SYZ for event 2007.355.07.

Figure B.554: Diagnostic plot of splitting measurement made on SYZ for event 2008.286.20.

Figure B.555: Diagnostic plot of splitting measurement made on SYZ for event 2009.286.05.
Figure B.556: Diagnostic plot of splitting measurement made on SYZ for event 2010.144.16.

Figure B.557: Diagnostic plot of splitting measurement made on SYZ for event 2011.245.10.

Figure B.558: Diagnostic plot of splitting measurement made on SYZ for event 2012.080.18.
Figure B.559: Diagnostic plot of splitting measurement made on SYZ for event 2012.102.22.

B.46 THZ

Figure B.560: Diagnostic plot of splitting measurement made on THZ for event 2003.265.04.

Figure B.561: Diagnostic plot of splitting measurement made on THZ for event 2004.162.15.
Figure B.562: Diagnostic plot of splitting measurement made on THZ for event 2004.283.21.

Figure B.563: Diagnostic plot of splitting measurement made on THZ for event 2004.320.09.

Figure B.564: Diagnostic plot of splitting measurement made on THZ for event 2004.325.08.
Figure B.565: Diagnostic plot of splitting measurement made on THZ for event 2005.141.05.

Figure B.566: Diagnostic plot of splitting measurement made on THZ for event 2005.168.06.

Figure B.567: Diagnostic plot of splitting measurement made on THZ for event 2005.269.01.
Figure B.568: Diagnostic plot of splitting measurement made on THZ for event 2007.230.02.

Figure B.569: Diagnostic plot of splitting measurement made on THZ for event 2007.320.03.

Figure B.570: Diagnostic plot of splitting measurement made on THZ for event 2008.187.02.
Event: 02-Feb-2009 (033) 17:53 -13.58N -76.56E  21km  Mw=6.0
Station: THZ   Backazimuth: 113.8º   Distance: ... 1 2 3 4sec
-90
-60
-30
0
30
60
90
-20 0 20 40 60 80
-200
-150
-100
-50
0
50
100
150
N
E
Inc = 9.3º

Event: 21-Apr-2009 (111) 05:26   50.83N 155.01E  152km  Mw=6.2
Station: THZ   Backazimuth: 348.8º   Distance: ... Map of T
0 1 2 3 4sec
-90
-60
-30
0
30
60
90
-20 0 20 40 60 80
-200
-100
0
100
200
N
E
Inc = 9.5º

Figure B.571: Diagnostic plot of splitting measurement made on THZ for event 2009.033.17.

Figure B.572: Diagnostic plot of splitting measurement made on THZ for event 2009.107.02.

Figure B.573: Diagnostic plot of splitting measurement made on THZ for event 2009.111.05.
Figure B.574: Diagnostic plot of splitting measurement made on THZ for event 2009.222.19.

Figure B.575: Diagnostic plot of splitting measurement made on THZ for event 2009.264.08.

Figure B.576: Diagnostic plot of splitting measurement made on THZ for event 2010.144.16.
Figure B.577: Diagnostic plot of splitting measurement made on THZ for event 2010.181.07.

Figure B.578: Diagnostic plot of splitting measurement made on THZ for event 2010.224.11.

Figure B.579: Diagnostic plot of splitting measurement made on THZ for event 2011.171.16.
**Figure B.580:** Diagnostic plot of splitting measurement made on THZ for event 2011.326.18.

**Figure B.581:** Diagnostic plot of splitting measurement made on THZ for event 2012.080.18.

**Figure B.582:** Diagnostic plot of splitting measurement made on THZ for event 2012.102.22.
Figure B.583: Diagnostic plot of splitting measurement made on TUZ for event 2003.270.11.

Figure B.584: Diagnostic plot of splitting measurement made on TUZ for event 2004.162.15.

Figure B.585: Diagnostic plot of splitting measurement made on TUZ for event 2004.325.22.
Figure B.586: Diagnostic plot of splitting measurement made on TUZ for event 2004.333.18.

Figure B.587: Diagnostic plot of splitting measurement made on TUZ for event 2004.353.06.

Figure B.588: Diagnostic plot of splitting measurement made on TUZ for event 2005.165.17.
Figure B.589: Diagnostic plot of splitting measurement made on TUZ for event 2005.166.02.

Figure B.590: Diagnostic plot of splitting measurement made on TUZ for event 2005.269.01.

Figure B.591: Diagnostic plot of splitting measurement made on TUZ for event 2005.288.10.
Figure B.592: Diagnostic plot of splitting measurement made on TUZ for event 2006.236.21.

Figure B.593: Diagnostic plot of splitting measurement made on TUZ for event 2007.068.03.

Figure B.594: Diagnostic plot of splitting measurement made on TUZ for event 2007.196.13.
Figure B.595: Diagnostic plot of splitting measurement made on TUZ for event 2007.214.03.

Figure B.596: Diagnostic plot of splitting measurement made on TUZ for event 2007.320.03.

Figure B.597: Diagnostic plot of splitting measurement made on TUZ for event 2008.063.09.
Figure B.598: Diagnostic plot of splitting measurement made on TUZ for event 2008.187.02.

Figure B.599: Diagnostic plot of splitting measurement made on TUZ for event 2008.329.09.

Figure B.600: Diagnostic plot of splitting measurement made on TUZ for event 2009.108.19.
Figure B.601: Diagnostic plot of splitting measurement made on TUZ for event 2009.264.08.

Figure B.602: Diagnostic plot of splitting measurement made on TUZ for event 2009.286.20.

Figure B.603: Diagnostic plot of splitting measurement made on TUZ for event 2010.199.05.
Figure B.604: Diagnostic plot of splitting measurement made on TUZ for event 2010.216.12.

Figure B.605: Diagnostic plot of splitting measurement made on TUZ for event 2010.281.03.

B.48 WHZ

Figure B.606: Diagnostic plot of splitting measurement made on WHZ for event 2004.333.18.
Figure B.607: Diagnostic plot of splitting measurement made on WHZ for event 2005.164.22.

Figure B.608: Diagnostic plot of splitting measurement made on WHZ for event 2005.166.02.

Figure B.609: Diagnostic plot of splitting measurement made on WHZ for event 2005.269.01.
Figure B.610: Diagnostic plot of splitting measurement made on WHZ for event 2006.274.09.

Figure B.611: Diagnostic plot of splitting measurement made on WHZ for event 2006.293.10.

Figure B.612: Diagnostic plot of splitting measurement made on WHZ for event 2007.119.12.
Figure B.613: Diagnostic plot of splitting measurement made on WHZ for event 2007.150.20.

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Figure B.615: Diagnostic plot of splitting measurement made on WHZ for event 2008.187.02.
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Figure B.617: Diagnostic plot of splitting measurement made on WHZ for event 2009.015.17.

Figure B.618: Diagnostic plot of splitting measurement made on WHZ for event 2009.097.04.
Figure B.619: Diagnostic plot of splitting measurement made on WHZ for event 2009.286.20.

Figure B.620: Diagnostic plot of splitting measurement made on WHZ for event 2010.144.16.

Figure B.621: Diagnostic plot of splitting measurement made on WHZ for event 2010.281.03.
Figure B.622: Diagnostic plot of splitting measurement made on WHZ for event 2010.357.14.

Figure B.623: Diagnostic plot of splitting measurement made on WHZ for event 2011.301.18.

Figure B.624: Diagnostic plot of splitting measurement made on WHZ for event 2012.102.22.
Figure B.625: Diagnostic plot of splitting measurement made on WKZ for event 2004.162.15.

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Figure B.627: Diagnostic plot of splitting measurement made on WKZ for event 2004.325.08.
Figure B.628: Diagnostic plot of splitting measurement made on WKZ for event 2004.353.06.

Figure B.629: Diagnostic plot of splitting measurement made on WKZ for event 2005.080.12.

Figure B.630: Diagnostic plot of splitting measurement made on WKZ for event 2005.269.01.
Figure B.631: Diagnostic plot of splitting measurement made on WKZ for event 2005.339.12.

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Figure B.649: Diagnostic plot of splitting measurement made on WKZ for event 2010.181.07.

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Figure B.651: Diagnostic plot of splitting measurement made on WKZ for event 2012.102.22.
Figure B.652: Diagnostic plot of splitting measurement made on WVZ for event 22-Sep-2003.

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Figure B.674: Diagnostic plot of splitting measurement made on WVZ for event 2011.326.18.

Figure B.675: Diagnostic plot of splitting measurement made on WVZ for event 2012.080.18.
Figure B.676: Diagnostic plot of splitting measurement made on WVZ for event 2012.102.22.
Appendix C

Stacked Energy Contours from Shear Wave Splitting Measurements

Stacking individual energy contours following Wolfe and Silver [1998] helps reduce effects of acknowledged shortcomings of Silver and Chan [1991], where large amounts of noise and/or small degrees of birefringence can yield incorrect values of $\varphi$ and $\delta t$ [Levin et al., 2007; Monteiller and Chevrot, 2010; Restivo and Helffrich, 1999]. In such instances, the measured value of $\varphi$ lies parallel to the backazimuth of the earthquake used in the analysis, with the result that Silver and Chan's method can yield unrealistically large values of $\delta t$. Such measurements are typically classified as null measurements. Because the ocean is inherently a noisy environment, many splitting measurements exhibited these characteristics, which makes it difficult to distinguish null measurements from cases with low signal-to-noise ratios (defined as the ratio of the maximum amplitude on the radial component to the standard deviation of the transverse component in the analysis window). With the stacking method of Wolfe and Silver [1998], however, we were able to use such measurements to distinguish noisy waveforms and/or small delay times from nodal results.

The following figures show stacks of energy contours of the transverse component as a function of $\varphi$ versus $\delta t$. The grey contour is the 95% confidence interval. The intersection of the black lines defines the best fitting splitting parameters.
Figure C.1: Stacked energy contour for APZ.

Figure C.2: Stacked energy contour for CRLZ.

Figure C.3: Stacked energy contour for CTZ.
Figure C.4: Stacked energy contour for DCZ.

Figure C.5: Stacked energy contour for DSZ.

Figure C.6: Stacked energy contour for EAZ.
Figure C.7: Stacked energy contour for FOZ.

Figure C.8: Stacked energy contour for JCZ.

Figure C.9: Stacked energy contour for KHZ.
Figure C.10: Stacked energy contour for LBZ.

Figure C.11: Stacked energy contour for LTZ.

Figure C.12: Stacked energy contour for MLZ.
Figure C.13: Stacked energy contour for MQZ.

Figure C.14: Stacked energy contour for MSZ.

Figure C.15: Stacked energy contour for NNZ.
Figure C.16: Stacked energy contour for NZ02.

Figure C.17: Stacked energy contour for NZ03.

Figure C.18: Stacked energy contour for NZ04.
Figure C.19: Stacked energy contour for NZ05.

Figure C.20: Stacked energy contour for NZ09.

Figure C.21: Stacked energy contour for NZ10.
Figure C.22: Stacked energy contour for NZ11.

Figure C.23: Stacked energy contour for NZ12.

Figure C.24: Stacked energy contour for NZ13.
Figure C.25: Stacked energy contour for NZ15.

Figure C.26: Stacked energy contour for NZ16.

Figure C.27: Stacked energy contour for NZ18.
Figure C.28: Stacked energy contour for NZ20.

Figure C.29: Stacked energy contour for NZ24.

Figure C.30: Stacked energy contour for NZ25.
Figure C.31: Stacked energy contour for NZ26.

Figure C.32: Stacked energy contour for NZ27.

Figure C.33: Stacked energy contour for NZ28.
Figure C.34: Stacked energy contour for NZ29.

Figure C.35: Stacked energy contour for NZ30.

Figure C.36: Stacked energy contour for ODZ.
Figure C.37: Stacked energy contour for OPZ.

Figure C.38: Stacked energy contour for OXZ.

Figure C.39: Stacked energy contour for PYZ.
Figure C.40: Stacked energy contour for QRZ.

Figure C.41: Stacked energy contour for RPZ.

Figure C.42: Stacked energy contour for SYZ.
Figure C.43: Stacked energy contour for THZ.

Figure C.44: Stacked energy contour for TUZ.

Figure C.45: Stacked energy contour for WHZ.
Figure C.46: Stacked energy contour for WKZ.

Figure C.47: Stacked energy contour for WVZ.
Appendix D

Supplemental Material for Teleseismic P-wave tomography of South Island, New Zealand upper mantle: Evidence of subduction of Pacific lithosphere since 45 Ma

Table D.1: List of station locations used in this study along with estimates of crust (crust thick) and sediment (sed thick) thicknesses used to make corrections to the travel-time data. The corrections shown here (sed corr, crust corr, and total corr) are for a vertically traveling P-wave assuming a sediment speed of 2.5 km/s, a crustal speed of 6.2 km/s, an oceanic crustal speed of 6.5 km/s (stations NZ02 - NZ04), and a mantle speed of 8.1 km/s. Starred stations (*) are stations at which Pn measurements were not published [Collins and Molnar, 2014] and we instead used crustal thicknesses from Salmon et al. [2013]. This column shows the magnitude of sample station corrections, but we did not invert for station terms.

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Table D.2: List of earthquakes used in the P-wave tomography study.

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Figure D.1: Examples of the cross-correlation of waveforms. Waveforms of an earthquake A) from the Banda Sea with backazimuth of 305° ($m_b = 6.7$) in the filter band 0.05 - 0.1 Hz; and B) from offshore Japan with backazimuth of 333° ($m_b = 6.5$) in the filter band 0.08 - 0.12 Hz. Red lines denote the beam window and blue lines denote the robust window as defined in *dbxcor* [Pavlis and Vernon, 2010]. Station names are to the left of the seismograms, with MOANA OBSs beginning in NZ. The x-axis is time in seconds. Traces are ordered by correlation coefficient, with the trace with the highest correlation coefficient at the top.
Figure D.2: Measured $P$-wave travel-time residuals in the frequency band 0.08 - 0.12 Hz after application of a crustal and sediment correction for five different events from a range of backazimuths. Figure 3.4 shows residuals measured in the band 0.05 - 0.1 Hz. Earthquakes: A) near Samoa with backazimuth of $32^\circ$ ($m_b = 7.1$); B) near the South Sandwich Islands with backazimuth $177^\circ$ ($m_b = 6.2$); C) also near the South Sandwich Islands with backazimuth $171^\circ$ ($m_b = 6.2$); D) from offshore the Solomon Islands with backazimuth of $334^\circ$ ($m_b = 6.3$); and E) from offshore Japan with backazimuth of $333^\circ$ ($m_b = 6.5$). Areas of the circles are proportional to the magnitudes of the travel-time residuals. Black arrows point from the earthquake epicenter. A negative (positive) travel-time residual indicates an early (late) arrival with respect to average times for the AK135 reference model [Kennett et al., 1995].
Figure D.3: Plot depicting the trade-off analysis between model norm and variance reduction for inversion of the crustal corrected $P$-wave travel-time data. Different curves correspond to different smoothing values (labeled “S”), and each point on the curve shows a damping value (labeled “D”). The black star shows the parameters used in this study (damping = 5 and smoothing = 4).
Figure D.4: Standard checkerboard test showing depth slices at 60, 125, 285, and 375 km. The left column shows the synthetic input structure (+5% or -5% $V_p$ of the background model, AK135, with cubes 200-km-wide on a side), and the right column shows the recovered structure.
Figure D.5: A cross-section along 42.6°S for the checkerboard test. The left shows the synthetic input structure (+5% or -5% \( V_p \) of the background model, AK135), and the right shows the recovered structure.
Figure D.6: Depth slices at 60, 90, 125, 165, 205, 245, 285, 330, 375, 420, and 465 km through a P-wave tomogram obtained without travel-time measurements from NZ14 and NZ15. Removing NZ14 and NZ15 helps test a possibly dominant role played by either P-wave anisotropy or erroneous corrections for crustal thickness and sediment on the final tomographic image.
Figure D.7: Depth slices at 60, 90, 125, 165, 205, 245, 285, 330, 375, 420, and 465 km through a P-wave tomogram for which the OBS arrival times were not used.
D.1 Correlations for Sediment and Crust

We assume that the sediment and crust correction to the teleseismic travel-time measurements is the sum of the travel-time difference between a ray traveling in sediment versus mantle plus the difference between a ray traveling in crust versus mantle (Figure D.8).

\[ t_c = \frac{h_s}{v_s \cos(i)} \]  \hspace{1cm} (D.1)

where \( h_s \) is the thickness of the sediment, \( v_s \) is the \( P \)-wave speed in sediment, and \( i \) is the incidence angle in the sediment. For the equivalent path through the mantle:

\[ x_s = z \cos(k - i) \]  \hspace{1cm} (D.2)

where \( k \) is the incidence angle in the mantle and \( z = h_s / \cos(i) \). The time it takes to travel path \( x_s \) is

\[ t_m = \frac{h_s \cos(k - i)}{v_m \cos(i)} \]  \hspace{1cm} (D.3)
where $v_m$ is the $P$-wave speed in the mantle. The difference in travel-time between a ray traveling in sediment versus in the crust is

$$\Delta T_s = \frac{h_s}{\cos(i)} \left[ \frac{1}{v_s} - \frac{\cos(k-i)}{v_m} \right].$$  \hfill (D.4)

Likewise, for the travel-time difference between crust and mantle:

$$\Delta T_c = \frac{h_c}{\cos(j)} \left[ \frac{1}{v_c} - \frac{\cos(k-j)}{v_m} \right].$$  \hfill (D.5)

Thus, the total correction is

$$\Delta T = \frac{h_s}{\cos(i)} \left[ \frac{1}{v_s} - \frac{\cos(k-i)}{v_m} \right] + \frac{h_c}{\cos(j)} \left[ \frac{1}{v_c} - \frac{\cos(k-j)}{v_m} \right].$$  \hfill (D.6)

A $Pn$ time term is the difference in travel-time between a horizontally traveling $P$-wave in the mantle and one refracted to the surface. To determine crustal thickness from the $Pn$ time terms, we solve Equation (D.6) for $h_c$, with $k = 90^\circ$, $\Delta T$ equal to the $Pn$ time terms determined by Collins and Molnar [2014], $h_s$ equal to the sediment thickness given by Wood and Woodward [2002], $v_m = 8.1$ km/s, $v_c = 6.2$ km/s, and $v_s = 2.5$ km/s. From Snell’s Law, $j = \arcsin \left( \frac{v_c}{v_m} \right)$ and $i = \arcsin \left( \frac{\sin(j) v_s}{v_c} \right)$.

Once we determine the crustal thickness under each station, we can then use Equation (D.6) to calculate the corrections to the travel-time measurements on an event-by-event basis with incidence angles calculated using the TauP calculator.
Appendix E

P-Wave Travel-Time Residuals by Earthquake

The following figures show the measured $P$-wave travel-time residuals after application of a crust and sediment correction for each earthquake used in the teleseismic $P$-wave tomography. Areas of the circles are proportional to the magnitudes of the travel-time residuals. Black arrows point from the earthquake epicenter. A negative (positive) travel-time residual indicates an early (late) arrival.

Figure E.1: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 670.
Figure E.2: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 3431.

Figure E.3: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 3924.
Figure E.4: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 4670.

Figure E.5: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 6963.
Figure E.6: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 8664.

Figure E.7: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 9497.
Figure E.8: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 12368.

Figure E.9: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 12826.
Figure E.10: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 12891.

Figure E.11: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 13268.
Figure E.12: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 14079.

Figure E.13: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 14388.
Figure E.14: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 14934.

Figure E.15: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 15548.
Figure E.16: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 15587.

Figure E.17: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 15748.
Figure E.18: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 15931.

Figure E.19: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 16685.
Figure E.20: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 16977.

Figure E.21: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 17461.
Figure E.22: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 17488.

Figure E.23: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 17642.
Figure E.24: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 18092.

Figure E.25: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 19720.
Figure E.26: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 20248.

Figure E.27: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 20401.
Figure E.28: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 20855.

Figure E.29: Measured $P$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 21102.
Figure E.30: Measured $P$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 670.

Figure E.31: Measured $P$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 4670.
Figure E.32: Measured $P$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 6963.

Figure E.33: Measured $P$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 8664.
Figure E.34: Measured $P$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 12826.

Figure E.35: Measured $P$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 13268.
Figure E.36: Measured $P$-wave travel-time residuals in the frequency band $0.08 - 0.12$ Hz after application of a crust and sediment correction for event 14079.

Figure E.37: Measured $P$-wave travel-time residuals in the frequency band $0.08 - 0.12$ Hz after application of a crust and sediment correction for event 15548.
Figure E.38: Measured $P$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 15587.

Figure E.39: Measured $P$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 15748.
Figure E.40: Measured $P$-wave travel-time residuals in the frequency band $0.08 - 0.12$ Hz after application of a crust and sediment correction for event 15931.

Figure E.41: Measured $P$-wave travel-time residuals in the frequency band $0.08 - 0.12$ Hz after application of a crust and sediment correction for event 16685.
Figure E.42: Measured $P$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 16977.

Figure E.43: Measured $P$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 17461.
Figure E.44: Measured $P$-wave travel-time residuals in the frequency band $0.08$ – $0.12$ Hz after application of a crust and sediment correction for event 20401.
Appendix F

Observed and Predicted $P$-Wave Travel-Time Residuals by Station

The following figures show the observed and predicted $P$-wave travel-time residuals and their difference at each station plotted by incident angle at the Moho. Areas of circles are proportional to magnitudes of the travel-time residuals.

Figure F.1: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station APZ.
Figure F.2: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station CASS.

Figure F.3: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station CROE.
Figure F.4: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station DCZ.

Figure F.5: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station DSZ.
Figure F.6: Observed (left) and predicted (center) P-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station EAZ.

Figure F.7: Observed (left) and predicted (center) P-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station FOZ.
Figure F.8: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station FREW.

Figure F.9: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station JCZ.
Figure F.10: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station KELY.

Figure F.11: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station KHZ.
Figure F.12: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station LBZ.

Figure F.13: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station LTZ.
Figure F.14: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station MLZ.

Figure F.15: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station MQZ.
Figure F.16: Observed (left) and predicted (center) \(P\)-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station MSZ.

Figure F.17: Observed (left) and predicted (center) \(P\)-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NNZ.
Figure F.18: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ02.

Figure F.19: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ03.
Figure F.20: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ04.

Figure F.21: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ05.
Figure F.22: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ06.

Figure F.23: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ07.
Figure F.24: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ08.

Figure F.25: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ09.
Figure F.26: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ10.

Figure F.27: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ11.
Figure F.28: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ12.

Figure F.29: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ13.
Figure F.30: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ14.

Figure F.31: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ15.
Figure F.32: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ16.

Figure F.33: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ18.
Figure F.34: Observed (left) and predicted (center) \(P\)-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ19.

Figure F.35: Observed (left) and predicted (center) \(P\)-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ20.
Figure F.36: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ21.

Figure F.37: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ22.
Figure F.38: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ23.

Figure F.39: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ24.
Figure F.40: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ25.

Figure F.41: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ26.
Figure F.42: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ27.

Figure F.43: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ28.
Figure F.44: Observed (left) and predicted (center) \(P\)-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ29.

Figure F.45: Observed (left) and predicted (center) \(P\)-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ30.
Figure F.46: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station ODZ.

Figure F.47: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station OPZ.
Figure F.48: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station OXZ.

Figure F.49: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station PYZ.
Figure F.50: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station QRZ.

Figure F.51: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station RPZ.
Figure F.52: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station SYZ.

Figure F.53: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station THZ.
Figure F.54: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station TUZ.

Figure F.55: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station WHZ.
Figure F.56: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station WKZ.

Figure F.57: Observed (left) and predicted (center) $P$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station WVZ.
Appendix  G

Supplemental Material for Investigation of mantle deformation due to oblique convergence from teleseismic $S$-wave tomography of South Island, New Zealand upper mantle

Table G.1: List of station locations used in this study along with estimates of crust (crust thick) and sediment (sed thick) thicknesses used to make corrections to the travel-time data. The corrections shown here (sed corr, crust corr, and total corr) are for a vertically traveling $S$-wave assuming a sediment speed of 1.0 km/s, a crustal speed of 3.6 km/s, an oceanic crustal speed of 3.8 km/s (stations NZ02 - NZ04), and a mantle speed of 4.5 km/s. Starred stations (*) are stations at which $Pn$ measurements were not published [Collins and Molnar, 2014] and we instead used crustal thicknesses from Salmon et al. [2013].

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Table G.2: List of earthquakes used in the S-wave tomography study.

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Table G.2 – continued from previous page

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Figure G.1: Examples of the cross-correlation of waveforms. Waveforms of an earthquake A) from offshore Sumatra with backazimuth of 283° ($m_b = 7.1$) in the filter band 0.05 - 0.1 Hz; and B) from the Banda Sea with backazimuth of 297° ($m_b = 6.3$) in the filter band 0.08 - 0.12 Hz. Red lines denote the beam window and blue lines denote the robust window as defined in \textit{dbxcor} [Pavlis and Vernon, 2010]. Station names are to the left of the seismograms, with MOANA OBSs beginning in “NZ.” The x-axis is time in seconds. Traces are ordered by correlation coefficient, with the trace with the highest correlation coefficient at the top.
Figure G.2: Plot of measured travel-time residuals in the band 0.08 - 0.12 Hz after application of a crustal and sediment correction for 4 different events from different backazimuths. Figure 4.5 shows residuals measured in the band 0.05 - 0.1 Hz. Earthquakes A) near the South Sandwich Islands with backazimuth of 171° ($m_b = 6.2$); B) on the South West Indian Ridge with a backazimuth of 239° ($m_b = 5.8$); C) offshore the Solomon Islands with a backazimuth of 334° ($m_b = 6.3$); and D) north of Vanuatu with a backazimuth of 350° ($m_b = 5.7$). Black arrows point from the earthquake, with epicentral distance given on the plot. A negative (positive) travel-time residual indicates an early (late) arrival.
Figure G.3: Plot depicting the trade-off analysis between model norm and variance reduction for inversion of the crust and sediment corrected $S$-wave travel-time data. Each curve represents a single smoothing value (labeled “$S$”), and each point on the curve shows a damping value (labeled “$D$”). The black star shows the parameters used in this study (damping = 5 and smoothing = 4).
Figure G.4: Resolution for depth slices at 60, 125, 285, and 375 km based on a standard checkerboard test with cubes 200 km per side. The left column shows the synthetic input structure (+10% or -10% the reference S-wave speed of AK135) and the right column shows the recovered structure.
Figure G.5: Resolution for a cross-section W-E along 42.6°S based on a standard checkerboard test. The left shows the synthetic input structure (+10% or -10% the S-wave speed of the reference model, AK135) and the right shows the recovered structure.
Figure G.6: Depth slices at 60, 90, 125, 165, 205, 245, 285, 330, 375, 420, and 465 km through an S-wave tomogram obtained without travel-time measurements from NZ14 and NZ15. This tests the possible roll of anisotropy at these stations.
Appendix H

S-Wave Travel-Time Residuals by Earthquake

The following figures show the measured S-wave travel-time residuals after application of a crust and sediment correction for each earthquake used in the teleseismic S-wave tomography. Areas of the circles are proportional to the magnitudes of the travel-time residuals. Black arrows point from the earthquake epicenter. A negative (positive) travel-time residual indicates an early (late) arrival.

Figure H.1: Measured S-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 3431.
Figure H.2: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 3924.

Figure H.3: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 4670.
Figure H.4: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 7326.

Figure H.5: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 8664.
Figure H.6: Measured S-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 12826.

Figure H.7: Measured S-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 12891.
Figure H.8: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 13268.

Figure H.9: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 13871.
Figure H.10: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 14079.

Figure H.11: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 14934.
Figure H.12: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 15587.

Figure H.13: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 15904.
Figure H.14: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 15931.

Figure H.15: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 16247.
Figure H.16: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 16685.

Figure H.17: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 16977.
Figure H.18: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 17488.

Figure H.19: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 17642.
Figure H.20: Measured S-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 18092.

Figure H.21: Measured S-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 18202.
Figure H.22: Measured S-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 20322.

Figure H.23: Measured S-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 20855.
Figure H.24: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 21102.

Figure H.25: Measured $S$-wave travel-time residuals in the frequency band 0.05 – 0.1 Hz after application of a crust and sediment correction for event 22281.
Figure H.26: Measured $S$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 4670.

Figure H.27: Measured $S$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 8664.
Figure H.28: Measured S-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 12891.

Figure H.29: Measured S-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 13871.
Figure H.30: Measured $S$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 14079.

Figure H.31: Measured $S$-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 15587.
Figure H.32: Measured $S$-wave travel-time residuals in the frequency band $0.08 - 0.12$ Hz after application of a crust and sediment correction for event 15904.

Figure H.33: Measured $S$-wave travel-time residuals in the frequency band $0.08 - 0.12$ Hz after application of a crust and sediment correction for event 15931.
Figure H.34: Measured S-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 16685.

Figure H.35: Measured S-wave travel-time residuals in the frequency band 0.08 – 0.12 Hz after application of a crust and sediment correction for event 20322.
Figure H.36: Measured $S$-wave travel-time residuals in the frequency band $0.08 - 0.12$ Hz after application of a crust and sediment correction for event 22281.
Appendix I

Observed and Predicted $S$-Wave Travel-Time Residuals by Station

The following figures show the observed and predicted $S$-wave travel-time residuals and their difference at each station plotted by incident angle at the Moho. Areas of circles are proportional to magnitudes of the travel-time residuals.

Figure I.1: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station APZ.
Figure I.2: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station CASS.

Figure I.3: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station CROE.
Figure I.4: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station DCZ.

Figure I.5: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station DSZ.
Figure I.6: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station EAZ.

Figure I.7: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station FOZ.
Figure I.8: Observed (left) and predicted (center) S-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station FREW.

Figure I.9: Observed (left) and predicted (center) S-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station JCZ.
Figure I.10: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station KELY.

Figure I.11: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station KHZ.
Figure I.12: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station LBZ.

Figure I.13: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station LTZ.
Figure I.14: Observed (left) and predicted (center) S-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station MLZ.

Figure I.15: Observed (left) and predicted (center) S-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station MQZ.
Figure I.16: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station MSZ.

Figure I.17: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NNZ.
Figure I.18: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ02.

Figure I.19: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ03.
Figure I.20: Observed (left) and predicted (center) S-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ04.

Figure I.21: Observed (left) and predicted (center) S-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ05.
Figure I.22: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ06.

Figure I.23: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ07.
Figure I.24: Observed (left) and predicted (center) S-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ08.

Figure I.25: Observed (left) and predicted (center) S-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ09.
Figure I.26: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ10.

Figure I.27: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ11.
Figure I.28: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ12.

Figure I.29: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ13.
Figure I.30: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ14.

Figure I.31: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ15.
Figure I.32: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ16.

Figure I.33: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ18.
Figure I.34: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ19.

Figure I.35: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ20.
Figure I.36: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ21.

Figure I.37: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ22.
Figure I.38: Observed (left) and predicted (center) S-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ23.

Figure I.39: Observed (left) and predicted (center) S-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ24.
Figure I.40: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ25.

Figure I.41: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ26.
Figure I.42: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ27.

Figure I.43: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ28.
Figure I.44: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ29.

Figure I.45: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station NZ30.
Figure I.46: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station ODZ.

Figure I.47: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station OPZ.
Figure I.48: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station OXZ.

Figure I.49: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station PYZ.
Figure I.50: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station QRZ.

Figure I.51: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station RPZ.
Figure I.52: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station SYZ.

Figure I.53: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station THZ.
Figure I.54: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station TUZ.

Figure I.55: Observed (left) and predicted (center) $S$-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station WHZ.
Figure I.56: Observed (left) and predicted (center) \(S\)-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station WKZ.

Figure I.57: Observed (left) and predicted (center) \(S\)-wave travel-time residuals and their difference (right) plotted by incident angle at the Moho for station WVZ.