S-wave splitting in the transpressional zone of New Zealand plate-boundary: implications for deformation and dynamics

by
Sapthala Karalliyadda

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Abstract

Seismic anisotropy in the transpressional plate-boundary zone in New Zealand is investigated with shear-wave splitting to gain insights into lithospheric deformation and mantle flow. Constraints on plate-boundary deformation in the lithosphere of the oblique-collision and subduction regimes in South Island have been estimated from the local and regional shear-wave splitting parameters that are made at both inland and offshore seismographs. Mantle and lithospheric anisotropy of the southernmost Hikurangi subduction zone in the southern North Island is examined from SKS, ScS and teleseismic S-phases. The splitting of these phases measured on a recent transect crossing the Wellington region is analyzed to understand the lateral anisotropic structure of the fore-arc Hikurangi subduction zone.

Local and regional splitting reveal both laterally and depth varying anisotropy in South Island. The scatter in splitting parameters at individual stations suggests the splitting of high-frequency S-phases is mainly controlled by heterogeneous anisotropic structure and S-wave propagation direction within those heterogeneities. When the average results are examined as a whole through 2-D delay time tomographic inversion and spatial averaging, consistent patterns in delay times and fast azimuths exist. Spatially averaged fast azimuths indicate a localized high strain zone in the southern central region of the South Island. Based on fast azimuths observed above 100 km depth, we suggest that the plate-boundary sub-parallel anisotropy that is produced by pervasive shear is mainly distributed within a zone extending ~ 130 km SE of the Alpine fault in the southern South Island and is widely distributed (at least 200 km wide) in the northern South Island. Average station delay times (δt) of $\sim 0.1 - 0.4$ s compared to 1.7 s SKS δt from previous studies in South Island further suggest a deep seated anisotropic zone or sensitivity of S-wave splitting to the layered and/or heterogeneous anisotropic structure of the plate-boundary zone in the inland South Island. The heterogeneous anisotropic structure further suggests that the lithosphere is not only characterized by the plate-boundary parallel shear related to Cenozoic deformation, but is also affected by anisotropic imprints from the other tectonic episodes and anisotropy that is governed by the contemporary stress.

A shear-wave splitting anisotropy investigation in the offshore South Island regions is an extended study of the inland experiment and aims to provide a broad-scale understanding of the plate-boundary deformation. Individual splitting of local and regional S-phases yield a range of δt that varies between very small δt (~ 0.02 s), which may represent a nearly isotropic medium, and large δt (~ 0.6 s), which corresponds to lithospheric anisotropy. The average station δt of ~ 0.25 s and variable delays of the individual splitting measurements imply that the observed splitting is most likely controlled by the geometry of the ray paths. Long ray paths that are detected at the stations further away from the plate-boundary appear to penetrate to deeper lithosphere and capture a significant portion of the upper-mantle anisotropy to produce fast azimuths parallel to the plate-boundary shear (NE-SW). Thus, the long and deep ray paths respond to the deeper structure, but may not be

re-split by the upper-most crustal structures. However, the observed variable delays suggest that changes in ray propagation direction with respect to the orientation of symmetry axes of the anisotropic media may have an effect on the measured anisotropy. Offshore measurements that are close to the land are consistent with the inland measurements and appear to be controlled by the regional stress field. This implies that short and shallow ray paths are mostly sensitive to the crustal anisotropy. The uneven distribution of ray paths from the shallow and deep events, therefore, plays a dominant role in controlling the observed splitting depending on their depth sensitivity and/or extent of anisotropy. Consequently, when fast directions are spatially averaged along with the inland measurements consistent patterns appear to correlate with the possible depth contribution of anisotropy in the region. We are unable to provide accurate constraints on the offshore extent of plate-boundary parallel shear because of the shallow stress-controlled anisotropy that likely overlies the mantle-shear zone. However, the splitting parameters from long and deep ray paths suggest a deep-seated, plate-boundary sub-parallel shear in a broad zone at least in the northern and upper-central South Island.

Mantle anisotropy detected from teleseismic earthquakes recorded across the southern North Island displays NE-SW fast axis alignment, consistent with the strike of the Hikurangi trench and the predominant upper-plate faulting trends, with a range of δt ($\sim 0.5 - 3.0$ s) and small-scale variation in NE-SW fast azimuths. When combined with the previous measurements in the western side of the array, δt from long period (>7 s) S-phases indicate an abrupt lateral variation across the fore-arc Hikurangi subduction zone. This lateral variation together with frequency dependence suggest that the shear wave splitting in the fore-arc of the Hikurangi subduction zone in the southern North Island is governed in part by the laterally varying crustal contribution of anisotropy or isotropic velocity variations within the shallow crust. Frequency dependent splitting also suggests that the anisotropic structure is governed by either multilayer or more complex anisotropy perhaps due to the combined effects of laterally varying multilayer structure. If the variations are due to lateral changes in crustal anisotropy, then mantle and crustal deformation are most likely coupled in the east of the Wairarapa fault where there is a possibility of strong crustal contribution.

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Chapter 1

Overview

1.1 Introduction to study

This thesis reports an investigation of seismic anisotropy measured by shear-wave splitting in the transpressional domain of the plate-boundary zone in New Zealand to understand the deformation and dynamics of the system. This domain includes a continental oblique-collisional setting and two subduction systems. The first phase of study is designed to investigate anisotropic structure in the transpression-subduction domain in South Island from local/regional shear-waves. This is carried out as two sub-studies, namely an on-land experiment and an off-shore experiment. These two experiments provide a unique splitting parameter data-base and an extensive splitting analysis that allow us to seek answers to address the following questions:

1. How does anisotropy distribute laterally and vertically in and around the plate boundary zone at upper-lithospheric depths in South Island? Where does the majority of anisotropy reside?
2. What are the different sources of observed anisotropy? Is this zone governed only by plate-boundary shear?
3. If there is a plate-boundary shear induced deformation, can we resolve the depth extent of that anisotropy from local/regional S-splitting?
4. How do S-wave splitting parameters vary in off-shore South Island?
5. Does the mantle-shear zone beneath the transpressional Alpine fault extend over a wide region? Or is it localized in the lithospheric upper-mantle and distributed in the asthenosphere?

6. Can we delineate the possible deformation kinematics (e.g. dislocation creep, diffusion creep, and dynamic recrystallization etc.) from the observed anisotropy to understand the rheology of the system?

Local splitting parameters on land have been previously investigated by small-scale studies in some localized areas, especially in the northern South Island. Our attempt is to carry out a full-scale splitting analysis in the entire South Island including off-shore regions using a new seismic data set from an enhanced station network in South Island.

The second phase of the thesis investigates the subduction dynamics and lithospheric deformation in the south-east Hikurangi subduction zone of the North Island from teleseismic S-wave (i.e. SKS, ScS and regional S) splitting. Earthquake records from the broad-band transect (SAHKE transect) across the previously unexplored region of the south-east Hikurangi subduction zone enable us to examine:

1. Variations in anisotropic structure across the fore-arc region
2. Lithospheric-mantle wedge complexities and upper-plate deformation in response to oblique subduction

1.1.1 Thesis structure

The thesis consists of seven chapters. The first three chapters present background information and methodology. The next three chapters include results and interpretations. The final chapter summarizes the key findings of this study. The summary of the content in each chapter is given in the next page:

Chapter 1: Discusses the tectonic character and deformation at transpressional and subduction settings, and introduces New Zealand's tectonic setting, geology and geophysical structure. It also summarizes previous work that is related to this study.

Chapter 2: Seismic anisotropy and theories behind shear-wave splitting are explained. Different sources of anisotropic fabrics, their origin and tectonic significances are discussed to explain how deformation/dynamics can be investigated from splitting parameters.

Chapter 3: Contains information about the data utilized, experiment set-up and methodology. It also includes information on other different phenomena that can be used to characterize the observed splitting parameters.

Chapter 4: is a published journal article on splitting parameter analysis from land stations. Note that some parts of the introduction and methodology of this chapter are repetitive. Work related to this section, but completed after this manuscript publication, has been included appendix 5.3.1.

Chapter 5: This chapter includes results from the offshore experiment and this is also in the process of preparation for journal publication. Therefore, some introduction parts can be repetitive from previous chapters.

Chapter 6: Presents results from the study of deep anisotropy in the south-east Hikurangi subduction zone.

Chapter 7: Concludes the key findings and discusses limitations and future directions.

1.2 Deformation and dynamics at convergent margins

Convergent plate boundaries are the locus for subduction and/or mountain-building. Oceanic-continental plate convergence results in the subduction of the denser oceanic lithosphere underneath the continental lithosphere, whereas regions of continental-continental convergence, which are also known as ‘collisional zones’, can result in two lithospheric blocks sliding past each other, resulting in transform faulting and/or orogeny. Formation of these end products at continental convergent zones greatly depends on the obliquity of convergence. Both plate-boundary parallel and orthogonal plate movements occur at oblique-convergent zones; therefore, such zones are called ‘transpressional zones’. Transpressional boundaries can be characterized by both oblique-transform faulting and orogenesis as in the case of South Island, New Zealand. A significant portion of the plate-boundary that traverses New Zealand is considered to have a transpressional character and this transpressional zone contains to two subduction systems at the northern (Hikurangi) and southern (Puyseger) ends, in which the Pacific plate subducts beneath the Australian plate and visa versa (Fig. 1.1). The plate-boundary in New Zealand is a unique example of where subduction, transform faulting, and orogenesis can be observed along a single boundary-zone and provides an opportunity to seek solutions for the collision and subduction related questions.

One of the main concerns in both collision and subduction systems is how lithospheric blocks deform when they interact at plate-boundaries. The rheology of the system controls the distribution of the deformation that is produced by the plate-boundary forces [Thatcher, 2009]. Rheology primarily depends on the lithology (i.e. mineralogy and rocks), temperature, fluid involvement and tectonic environment [Burgmann and Dresen, 2008]. Unlike rigid oceanic plates, weak continental lithosphere is more susceptible to deform intensely at the plate-boundaries and is often associated with large-scale faulting [Molnar and Tapponnier,

Figure 1.1 Tectonic setting of New Zealand (MFZ: Marlborough Fault Zone, CVR: Central Volcanic Region). The relative plate motion (vector) between Australian and Pacific plates is denoted by the arrow with 38 mm/yr.

1975, England and Jackson, 1989]. The study of deformation distribution in such large-scale plate-boundary faults helps in understanding the strength of the fault at depth, delineating possible deformation mechanisms, and perhaps in forecasting the locations of mega-scale plate-boundary earthquakes. An architecture of such a deformation distribution around plate-boundaries can be explored via structural-geological, geodetic, petrophysical and geophysical investigations.

1.2.1 Deformation at continental zones

Subduction, collision and mountain-building

In certain tectonic settings (e.g. plate-boundary zone in South Island), subduction of the oceanic lithosphere at convergent margins leads to the collision of the two continental lithospheric blocks soon after consuming the oceanic lithosphere. Sediments extracted from the ocean basin are pushed up as two continental blocks converge, initiating mountain belts. In the long term, continental-collision results in the accumulation of deformation/strain that is possibly distributed in wide regions of the adjacent plate-boundary zone as a result of crustal shortening with folding, thrusting and continuous growth of the mountain belts (e.g. Asia-Eurasia collisional zone; *Molnar and Tapponnier* [1975]). As argued in several tectonic models [*Toussaint et al.*, 2004], structural evolution during transition from subduction to collision may occur either as continuous subduction of the lithosphere or lithospheric delamination, which leaves the whole crust or part of the crust at the surface. In the latter case, crust at the surface can be subjected to intense deformation (e.g. folding, faulting, crustal thickening) as the plates collide. Then the detached mantle-lithosphere may [*Toussaint et al.*, 2004]:

1. continue its subduction
2. break off and sink into the convecting mantle
3. thicken until it becomes unstable and drips down
4. shorten by folding

Which process is favored is controlled by factors such as rheology, thermal structure, composition of the lithosphere, rate of convergence, and the thickness of the lithospheric blocks [*van den Beukel and Wortel*, 1987].

Plate-boundary faults

At continental-collisional zones the deformation of the adjacent lithospheric blocks via large-scale faulting plays a dominant role in the dynamics of the entire system. In the suture zone, where two blocks of continental lithosphere collide, large-scale plate-boundary faults evolve depending on the relative motion of the two plates [Vauchez and Nicolas, 1991]. Forces acting on such faults are large enough to develop a strong structural discontinuity that extends from the surface through the brittle/semi-brittle crust and into the ductile upper-mantle or asthenosphere. Fault properties and behavior change from top to bottom, depending on the stress conditions and rheology of the system [Otsuki, 1978, Thatcher, 2009]. The depth extent of such large-scale faults is poorly constrained, since seismic slip is mostly confined to the first 10-20 km of the crust [Bourne *et al.*, 1998]. Two end member scenarios explain the discontinuous deformation and character of the faults in the brittle crust and the ductile continental lithosphere. One end member scenario is mainly focused on the character of the faults in the brittle upper crust [Bourne *et al.*, 1998]. This model emphasizes the block-like behavior of the regions between faults in the brittle crust and follows the plate-kinematic theory. Here, faulting areas are considered to be relatively weak compared to the host rock, dividing crust into blocks that are smaller than tectonic-plates. In contrast, the other end member scenario explains the continuous deformation of the entire lithosphere, assuming that deformation occurs on thin viscous sheets (e.g. vertical shear planes) with little or no depth variation in strain rates [England *et al.*, 1985, England and Jackson, 1989]. Thus, it does not consider the discontinuous slip behavior on faults as in the block model [Savage and Burford, 1970].

It is important to consider how large-scale faults extend horizontally and in depth by means of the above two scenarios. Is the deformation at faults biased towards one model? Or can faults accommodate the combined effects of both models? These

are some of the important questions yet to be answered from case studies. How plate boundary faults evolve in the ductile lithosphere and/or in the asthenosphere has been demonstrated by two models that incorporate the above two end-member scenarios (Fig. 1.2). As described in one model (figure 1.2-ii), narrow faults that exist in the brittle crust can extend into the lower-crust and to the upper-mantle as a narrow shear zone until they find a mechanically weak domain where they could attain distributed shear deformation [*Savage and Burford, 1970*]. In this case typically the semi-brittle lower-crust and ductile upper-mantle is strong enough to be cut by the narrow localized shear zones and distributed deformation only occurs in the asthenosphere, where dislocation creep deformation dominates (section 2.5). Although the preferred deformation mechanism in the lower-crust and upper-mantle is dislocation creep, other mechanisms (e.g. dynamic recrystallization, diffusion creep) can become prominent depending on the stress/strain, temperature conditions, mineralogy and grain size of the fault zone (section 2.5). For instance, grain size reduction in highly strained (shear strain 1-10) rock would produce fine grained mylonite/ultramylonite layers (high-strain shear zones) that indicate strain localization. In such regions the width of the shear zone can be as small as a few millimeters [*Burgmann and Dresen, 2008*]. Narrow (1-2 km) mylonite zones that were formed in the mid-crust and exposed at the exhumed sections of the Alpine fault are evidence for such narrow localized shear-zones at depth [*Little et al., 2002a*].

The other model (model (i) in figure 1.2) consists of a weak lower crust and upper mantle that deform via dislocation creep (section 2.5), allowing distributed deformation beneath the brittle faults [*McKenzie and Jackson, 1983*] similar to thin-viscous sheet type deformation. Intrinsically anisotropic minerals in the lower-crust (e.g. mica and amphiboles) and the upper-mantle (e.g. olivine) will produce an anisotropic medium when they deform via dislocation creep in such shear zones [*Nicolas, 1993*]. As inferred by seismic anisotropy studies (SKS shear-wave splitting, petrophysics and Pn anisotropy) in plate boundary fault zones (e.g. Alpine fault

Figure 1.2 The two end-member models of distributed shear (i) and localized (ii) deformation that explain the deformation in the plate-boundary zone of central South Island. Gray-shaded region indicates the shear zone and dark to light gray shading marks the distribution of strain (high to low shear strain) around the Alpine fault. The red arrow denotes the predicted depth distribution of shear zone below the fault.

zone, San Andreas fault zone), those shear deformation zones can be distributed in 200-400 km wide areas around the fault [Klosko *et al.*, 1999, Vauchez and Tommasi, 2003]. Nevertheless, in some cases such observations contradict the geodetic strain estimations and structural geology observations, suggesting either variations in deformation conditions (or the deformation mechanisms) within the shear zone or temporal variations in stress-strain conditions that may effect the state of deformation resulting in complex structure or that there is evolution from model(i) to model(ii) (see Fig. 1.2) [Houlié and Stern, 2012, Burgmann and Dresen, 2008].

1.3 Background: tectonic setting of New Zealand

New Zealand's land mass consist of two Islands (North and South), which are emergent segments (10 % of the total land mass) of the mostly submerged continental micro-plate 'Zealandia'. It was detached from the Gondwana supercontinent by sea floor spreading and oceanic-crust formation during the late Cretaceous (Fig. 1.1; Cox and Sutherland [2007]). Since then the geometry of the land mass has been reshaped through major tectonic episodes (section 1.3.1). The present-day setting of

New Zealand is tectonically significant as it lies along the active plate-boundary zone of the Australia-Pacific (AUS-PAC) plates. Tectonic activity along this boundary is marked by the two oblique subduction systems (Pysugur and Hikurangi subduction zones) with opposite convergence polarity, the central transpressional/collisional zone with major oblique-strike slip fault (i.e. the Alpine Fault) and Southern Alps mountain range, and Marlborough fault zone (MFZ) that indicates the transition from subduction to collision (Fig. 1.1).

1.3.1 Tectonic synthesis

New Zealand has evolved since the late Cretaceous (80 Ma) and undergone various tectonic processes. In the early Cretaceous, New Zealand was a part of the eastern Gondwana subduction system [Mortimer, 2004]. The crustal component of New Zealand is derived from the accretionary tectonic processes that took place during the Paleozoic and Mesozoic (Fig. 1.3-A; Cox and Sutherland [2007]). Late Cretaceous rifting that took place in the Tasman and Pacific sea isolated the continental fragment with New Zealand's landmass from the Gondwana super-continent, resulting in basin and oceanic crust formation in the southern ocean (Fig. 1.3-B). The plate-boundary through New Zealand started to develop during Eocene (~ 40 -45 Ma) approximately during the same time that the Resolution rift boundary and passive margin formed to the south and southwest of New Zealand respectively [Sutherland *et al.*, 2000].

Until the mid Oligocene, the deformation within the New Zealand plate-boundary zone was not that prominent, and the relative motion of AUS-PAC plates was relatively slow due to close proximity of the pole of rotation between the AUS-PAC plates and New Zealand [Cox and Sutherland, 2007]. However, in the late Oligocene (~ 25 Ma) the rate of relative motion of the AUS-PAC plate increased and the plate-boundary attained a dominant strike-slip behavior in the central collisional zone as

a result of the southward movement of the pole of rotation [*Cande and Stock*, 2004, *Cox and Sutherland*, 2007]. This change enhanced the strike-slip movement and caused the Australian passive-margin, which was formed during Eocene rifting, to migrate towards the central collisional zone (Fig. 1.3-C). The Eocene rift margin that intruded in to the New Zealand plate boundary zone in the late Oligocene had a direct impact on the late Cenozoic evolution of the plate-boundary and the deformation associated with the Alpine Fault especially in the South Island [*Sutherland et al.*, 2000]. Subduction along the northern part of the plate-boundary was initiated during 20 -25 Ma at the Hikurangi margin in the northern part of the strike-slip plate-boundary zone (Fig. 1.3-D). Subduction at the Puysegur margin, which is in the south of the strike-slip zone, has a relatively young origin history (~ 7 -10 Ma) compared to the Hikurangi subduction, however both systems started to evolve during the time when the central continental-continental margin was mostly strike-slip.

In central South Island, the pure strike-slip nature of the plate-boundary changed to oblique convergence around ~ 6 Ma, resulting in significant compression [*Walcott*, 1998]. According to *Cande and Stock* [2004], this convergence initiated ~ 11 Ma in the southern South Island (lat:44.5°). The present-day relative motion vector can be decomposed into a ~ 36 mma^{-1} plate-boundary parallel component and ~ 11 mm a^{-1} fault-normal components [*Norris et al.*, 1990]. Present-day GPS inversion suggests that $\sim 70\%$ of the total plate displacement is taken up by the Alpine fault [*Beavan et al.*, 1999, *Sutherland et al.*, 2000], and most of the rest is accommodated in the lithosphere as distributed deformation [*Norris et al.*, 1990]. The fault-normal component is responsible for ~ 100 km of crustal shortening since the origin of the oblique-convergence. The plate-parallel component governs the strike-slip movement of the Alpine Fault. Because of both strike-slip motion and the compressional character of the central South Island, this zone is well-described as a transpressional zone. The transpressional zone is linked with the Hikurangi subduction system

Figure 1.3 The tectonic evolution of New Zealand plate-boundary (after *Cox and Sutherland* [2007]). A dashed line marks the position of the Junction Magnetic Anomaly. Western, Eastern provinces and eastern North Island are denoted by **WNZ**, **ENZ**, and **ENI** respectively. Red and yellow stars show the positions of the two points on either side of the Alpine fault that became juxtaposed at the central Alpine fault as result of the plate-boundary evolution. A) ancient Gondwana subduction. B) rifting in the Tasman sea/south of New Zealand and separation from Gondwanaland. C) migration of the Australian Passive margin towards the strike-slip fault zone in Central South Island. D) a change in fault character from pure strike-slip to oblique strike-slip movement and subduction initiation.

(which is located in the northeast of the South Island and the east of North Island) by the transitional fault zone (Marlborough fault system) in the north/northeast of the South Island (Fig. 1.1). In the North Island, the Hikurangi subduction and the associated structures dominate over a ~ 250 km broad zone of plate margin [Lamb, 2011]. This subduction zone is characterized by the back-arc spreading in central-north, Central Volcanic Region (**CVR**), and the dextral-faulting system in the overriding plate, accommodating transpression (Fig.1.1).

1.4 Geology and structure of the South Island

The geology and geophysical structure of the area have been explored thoroughly in some localized areas (e.g. SIGHT and SAPSE experiments). In map view, the geology and geomorphology of New Zealand simply reflect the oblique-transform nature of the present-day plate-boundary zone (Fig. 1.4a).

Figure 1.4 (a) The geology map of New Zealand (data source: GNS science - www.gns.cri.nz.)

1.4.1 Geology

According to *Mortimer* [2004], New Zealand basement geology is mainly characterized by nine major volcanic-sedimentary terranes (from west to east: Buller, Takaka,

Figure 1.4 (b) The basement terrane map of New Zealand (after *Mortimer* [2004]).

Brook Street, Murihiku, Maitai, Caples, Bay of Islands, Rakaia, and Pahau), and minor igneous intrusions (three batholiths units: Median, Karamea-Paparoa and Hohonu) and overprinted metamorphic belts (Haast schist, gneiss and melange) in the volcanic-sedimentary terranes (Fig. 1.4b). Exposed basement rocks do not contain any Precambrian rock units and rock age varies from middle-Cambrian to late Early Cretaceous. These rocks are weakly metamorphosed and foliations are not intense. The parent source of the basement rock is the eroded sediments from the mountain belts on the subduction margin of eastern Gondwanaland (section

1.3.1; *Mortimer* [2004]). About 100 km away from the present-day plate-boundary, the alignment of volcanic-sedimentary belts and the lithologic boundaries that were built eastward represent the orientation of the ancient Gondwana margin, but close to the plate margin this orientation changes from fault-oblique to fault-subparallel with an apparent oroclinal bend that is likely associated with the dextral sense of motion during Cenozoic deformation [*Cox and Sutherland*, 2007]. Sedimentary terranes are divided into Western and Eastern provinces taking into account the position of Median Batholith that was formed ~ 375 -110 Ma (long-lived composite batholith) from subduction process, and considering the age of the rock units [*Cox and Sutherland*, 2007]. Eastern province rocks are generally younger than the western province. The following section summarizes the important information about those geologic units (Fig. 1.4b) and most of the details are adapted from *Mortimer* [2004] and *Cox and Sutherland* [2007].

Volcanic-sedimentary terranes

Volcanic-sedimentary terranes display weak metamorphism. Buller terrane, which is the most prominent lithology in the west of the South Island, has metamorphosed siliciclastic sandstone and mud stones. The rocks in this terrane generally display upright folds and steep cleavage, and belong to lower Paleozoic Gondwana sedimentary units. In northwestern South Island, the rock unit consists of granitic intrusions. Takaka terrane, which spreads across the western North Island, contains siliciclastic, carbonate and volcanic rocks of the Cambrian to Early Devonian age. This is a lithologically stratified complex rock unit, which is separated from the Brook Street terrane by the intrusion of median Batholith. The main rock type in the eastern province (~ 60 % of the New Zealand rock) is quartzofeldspathic greywacke. Pyroxene-rich basaltic volcanic rocks dominate in the Brook Street terrane. The contact between Brook Street terrane and Murihiku terrane in the northwestern North Island is marked by the Letham Ridge Thrust. Murihiku terrane consists

of volcanoclastic rock rich with marine sandstone. Maitai terrane contains the eastern Dun Mountain Ophiolite Belt that is used to constrain the plate-boundary deformation due to strike-slip motion and the Alpine fault offset [Wellman, 1979]. It is a well-stratified volcanoclastic sedimentary rock unit with spatially correlated magnetic properties (e.g. Junction magnetic anomaly).

Weakly metamorphosed volcanoclastic greywacke and argillite are stratified in the Caples terrane. The lithologic boundary between the Caples and Rakaia terranes is considered to be overprinted by the Otago (Haast) schist. The Rakaia terrane, which is in the eastern South Island and central North Island, is the largest terrane by area. This unit consists of submarine quartzofeldspathic sandstone-mudstone. In most of the places, steeply dipping greywacke (Torlesse greywacke) bedding planes (>60 - 90°) with rapid changes in strike of the bedding is significant. In some locations, shallowly dipping bedding planes ($<30^\circ$) are overturned to form folds with steeply dipping fold axes. Within the rock unit, metamorphic grade increases from east towards the overprinted Haast schist in west. The Bay of Island terrane also has a complex internal structure. Sandstone fabric in this terrane is similar to the petrofabric of the Rakaia terrane. Lithology of the Pahau terrane is consistent with the Rakaia terrane, but it is a younger Torlesse terrane compared to Rakaia terrane that consists of clastic detritus and trench deposits. Clastic detritus in this terrane is assumed to be recycled material from the Rakaia terrane [Wandres and Bradshaw, 2005, Mortimer, 2004].

High-grade metamorphic rocks

Geologically significant rock units in New Zealand are the three types of overprinted high-grade rocks (gneisses, Haast schist and melanges) with regional metamorphism. These metamorphosed rock units are linked with different deformational processes (e.g. subduction, oblique-convergence related) associated with the plate-

boundary evolution. Gneisses are considered to be the product of an amphibolite-granulite facies recrystallization during the Devonian to Cretaceous period [Cox and Sutherland, 2007], and are commonly seen in the western province. Older Torlesse quartzofeldspathic greywacke (Rakaia terrane) is considered to be the protolith of the Haast schist. Haast schist in South Island represents the distinct schist units (Otago, Marlborough and Alpine schists) with different history of origin and metamorphic characteristics. Otago schist (garnet-biotite-albite zone) is a north-west trending anticlinorium with sub-horizontal foliations, which spreads ~ 150 km in area. The schist is dominated by fold, shear zones and clear stretching lineations [Cox and Sutherland, 2007]. Although Marlborough schist is located further north, structures within that unit are similar to Otago schist in the south. Marlborough and Otago schists are interconnected by elongated, narrow schist bands of Alpine schist in the central South Island. The Alpine schist had been subjected to both amphibolite-facies peak metamorphism and ductile-brittle deformation during exhumation [Little *et al.*, 2005, Cox and Sutherland, 2007]. Depending on the metamorphic grade, Haast schist can also be sub-grouped into several deformation zones. Petrophysical studies have shown that these zones also exhibit a high degree of anisotropy [Godfrey *et al.*, 2000, Okaya *et al.*, 1995]. The outer-most zone is a chlorite zone with low grade metamorphism. Biotite and garnet-oligoclase zones reflect the increase in metamorphic grade towards the Alpine fault (and the plate boundary). These zones are also characterized by well developed foliations. Unlike Haast schist, melanges do not show pronounced foliations and they represent the deformation in the shallower zone of the Mesozoic accretionary prism.

1.4.2 Geophysical implications of the present-day structure

This section summarizes the important geophysical observations from previous experiments. The key questions that those studies aimed to address in this region are: How do plate-boundary faults (e.g. Alpine fault) evolve at depth? What is the crust-mantle character in response to transpression? What type of crust-mantle structure would best explain the observed geophysical, structural geology and geodetic observations? An on-land plate-boundary configuration, a young orogen with an active deformation and rapid exhumation, and ongoing tectonic processes with enhanced seismicity are some of the advantages that make this plate-boundary domain an exceptional place to seek answers for the above questions. Present-day structures in the study region can be divided into three main domains from north to south: 1) Hikurangi subduction zone, 2) Central transpressional zone, and 3) Fiordland subduction zone. Most of the geophysical experiments to date were focused on the structure beneath the continental collision zone with transpressional character.

Transpressional zone

The central transpressional zone is characterized by structures resulting from both collision and oblique strike-slip motion between the plates. The Alpine fault, which is the main surface manifestation of the plate-boundary, extends for about 800 km, striking 055° in the South Island [Sutherland, 1999]. It accommodates 70% of the current plate motion [Norris *et al.*, 1990] and the rate of relative plate motion is ~ 38 mm/yr with ~ 36 mm/yr plate-parallel and ~ 11 mm/yr plate-perpendicular components [DeMets *et al.*, 1994a]. Because of the right-lateral strike-slip motion, geologic terranes in South Island display a clear ~ 400 - 460 km dextral offset. According to [Little *et al.*, 2002b], the bending related deformation of geologic units and other structures in the central South Island (e.g. fold and faults) is mainly

governed by the dextral-transpression and the shortening in the crust (~ 100 km shortening across the plate-boundary [Norris *et al.*, 1990]) during the last 10 My.

The transpressional character of the central Alpine fault has resulted in crustal thickening. Consequently, the Southern Alps (peak height of 3700 m) is being uplifted at a rate of approximately 7-10 mm/yr and the shallow-mid crust of the Pacific plate is being exhumed along the Alpine fault [Davey *et al.*, 2007, Little *et al.*, 2005]. P_n travel times estimated from the waves refracted along the root of the Southern Alps in the central South Island suggest the crustal thickness of the South Island is highest (~ 48 km) about 80 km SE of the Alpine fault [Bourguignon *et al.*, 2007a]. As inferred by the SIGHT experiment, crustal thickening (~ 17 km) due to shortening is asymmetrically distributed and mostly concentrated in the central-eastern South Island beneath the Southern Alps. The thickness of the lithospheric root is about 100 km and this zone is also characterized by a mantle high-speed (and high density) body at depths of ~ 80 -100 km [Stern *et al.*, 2000]. Scherwath *et al.* [2003] noticed a negative Bouguer gravity anomaly that fits well with the location of the thick crustal root, which is isostatically imbalanced with negative buoyancy. This negatively-buoyant crustal root is assumed to be supported by the high density mantle body (i.e. the mantle high-speed zone discussed above) [Stern *et al.*, 2000]. Based on these findings along with SKS and P_n anisotropy estimates, Stern *et al.* [2000] proposed a lithospheric-thickening model with distributed deformation to explain the crust-mantle dynamics beneath the transpressional zone. Lithospheric deformation in this region has also been explained using the intracontinental subduction model [Wellman, 1979] with localized deformation. This model is also supported by numerical modeling in the central South Island [Beavan *et al.*, 2004, Ellis *et al.*, 2006].

Fiordland subduction

The central transpressional zone is linked to Puysegur subduction in the southwest offshore. Oceanic lithosphere that was formed at the AUS-PAC boundary about 40-25 Ma from rifting processes is being subducted at the Puysegur Trench [Davey *et al.*, 2007]. This subduction migrates into the Fiordland region (Fig. 1.1) due to transcurrent plate-motion of the inland plate-boundary [Lebrun *et al.*, 2003]. In Fiordland, the plate motion is partitioned between subduction and strike-slip motion. The Alpine fault in this region tends to be sub-vertical and accommodates $\sim 55\text{--}85\%$ of the total plate motion [Sutherland *et al.*, 2000]. Significant crustal deformation is assumed to occur in the region's eastern part because of the strong strike-slip behavior of the fault in the southern South Island, extending offshore. Studies also suggested an uplifted lower-crust during the Cretaceous in western Fiordland [Davey and Smith, 1983] and a high-velocity zone in the upper-mantle of the Pacific plate [Eberhart-Phillips and Reyners, 2001]. This high-velocity zone is attributed to a mechanically strong region in the upper-mantle, which may cause the subducting slab to change its strike (from $\sim 70^\circ$ to 15°) and dip from moderately to steeply dipping at the north end [Reyners *et al.*, 2002, Eberhart-Phillips and Reyners, 2001]. Yet, this subduction is relatively young (~ 10 My) with a maximum slab extension of ~ 150 km [Davey and Smith, 1983]. High SKS delays (2.5 s) recorded in this region have been explained as a coupling of trench-parallel flow in the subduction settings and the plate-boundary parallel shear deformation associated with the transpressional zone [Duclos *et al.*, 2005].

Hikurangi subduction and Marlborough faults

The northern termination of the transpressional zone is marked by the transitional fault zone (i.e. Marlborough fault zone-MFZ) in north/northeast South Island (Fig. 1.1). This zone then extends into the Hikurangi subduction zone with an

oblique subduction. The northern South Island is, therefore, subjected to both subduction related deformation and deformation associated with transform faults. The MFZ consists of four main strike-slip faults (Hope, Clarence, Awatere, and Wairau faults), and the northernmost Wairau fault is considered to be the northward continuation of the Alpine fault (section 4.1). As suggested by receiver function studies, unbroken Moho (depth $\gtrsim 15$ km) beneath this zone is compatible with pervasive shear deformation in the lithosphere below the crustal faults [Wilson *et al.*, 2004]. This shear deformation appears to control the rotational-type deformation of the crustal fault blocks that are presumably aligned with the shear fabric below [Hall *et al.*, 2004]. Oblique-subduction dominates in the southern North Island. Northwest-dipping subducting Pacific crust first encounters the overlying submerged accretionary prism and then the overriding Australian plate. The relative plate motion of ~ 42 mm/yr in this region is partitioned into plate-boundary parallel and orthogonal components, resulting in offshore thrust faulting, dextral faulting in the upper-plate and fault-block rotation [Henry *et al.*, 2013].

1.5 Previous work: seismic anisotropy and deformation in South Island

Various techniques (e.g. Pn anisotropy, shear-wave splitting, P-wave delays and GPS/structural studies) have been implemented to understand the deformation and dynamic structure of the plate-boundary zone in the South Island. When consider the plate motion history (section 1.3.1), this plate-boundary was mainly pure strike-slip during $\sim 30 - 6$ Ma and it attained transpressional character only in the last ~ 6 Ma. An SKS splitting study by Klosko *et al.* [1999] pointed out the possibility of broad-scale deformation that results from the plate-boundary shear, however the spatial and depth extent of this pervasive shear zone was not accurately resolved by this study. Plate-boundary parallel to sub-parallel SKS-splitting azimuths through

out the entire South Island have been interpreted as a large-scale strike-slip straining in the mantle that is possibly associated with the shear zone [Klosko *et al.*, 1999, Molnar *et al.*, 1999]. Such distributed shear deformation had also been predicted by geodetic studies [Walcott, 1998, Beavan *et al.*, 1999].

Finite strain governs the preferred alignment of intrinsically anisotropic minerals and creates an anisotropic medium in which S-waves can split into two mutually perpendicular quasi S-waves with different speeds [Savage, 1999]. For large strains in olivine dominant systems, the polarization direction of the fast S-wave reflects the maximum extension (or shear plane), whereas the polarization direction of the slow wave corresponds to the axis of maximum shortening [Ribe, 1992, Molnar *et al.*, 1999]. Taking into account the mantle strain that is constrained by SKS splitting measurements [Klosko *et al.*, 1999], geodetic and structural constraints [Walcott, 1998, Norris *et al.*, 1990] and the distributed lateral heterogeneity as detected from P-wave delays across the central plate-boundary zone, Molnar *et al.* [1999] suggested that the mantle lithosphere of the South Island is most likely characterized by distributed deformation as demonstrated in a thin-viscous sheet model and does not agree with the localized deformation model (Fig. 1.2-ii and section 1.2.1). Constraints from mantle rheology studies and strain modeling further suggest that pervasive shear at depths is most likely accommodated as vertical shear planes (fan shape) that reside within the lower-crust and the uppermost 100 km of lithospheric mantle [Collins *et al.*, 2010]. In the transition fault zone (Marlborough faults), interpretation of anisotropic receiver functions suggests a continuous Moho with spatial variations in crustal thickness that indicates the presence of distributed strain in the middle to lower crust just beneath the strain-localized segment of the fault [Wilson *et al.*, 2004]. This study suggested a large-scale pervasive shear zone at deeper parts (deeper than ~ 15 km) possibly extending from the lower-crust [Wilson *et al.*, 2004]. In agreement with these observations, local shear-wave splitting studies in the same region found fault-parallel (NE-SW) fast direction, possibly resulting

from this pervasive shear [Balfour *et al.*, 2005, Audoine *et al.*, 2000]. An anisotropic velocity inversion by Eberhart-Phillips and Henderson [2004] found a high anisotropy zone between the Awatere and Wairau faults (north-west of the Marlborough faults) that was interpreted as distributed aligned fractures, which resulted from the long-term shear strain in the brittle crust.

The following subsections summarize some of the other geophysical constraints on the plate-boundary deformation especially in the central South Island to test the competing models (Fig. 1.2) that with whether the deformation in this plate-boundary zone either localized at the plate-boundary fault (i.e Alpine Fault), extending into deep lithosphere, or distributed as a large-scale shear zone at depths.

Figure 1.5 Summary of the previous seismic anisotropy investigations in South Island (after Bourguignon *et al.* [2007b]). Red (T), blue (B_1 , B_2 and S) and green (E_1 and E_2) bars indicate the P_n anisotropy measurements. SKS measurements are denoted by the orange bars. T_1 , T_2 , 3W and 4E show the SIGHT transects across the South Island.

Constraints from P_n anisotropy studies

P_n phases that are refracted in the upper-mantle have been used to study crustal and upper-mantle anisotropy structure. Two parallel studies of Pn anisotropy across the South Island were carried out as a part of the South Island Geophysical Transect (SIGHT) experiment [Stern *et al.*, 2007]. P_n anisotropy investigated at the intersection between transect one (T_1) and line 3W west (~ 30 km) of the Alpine fault ranges between 11 - 13 % (Scherwath *et al.* [2002]; Fig. 1.5).

Assuming that faster Pn orientation is caused by lattice preferred orientation and reflects the same anisotropy detected by SKS splitting, the angle between maximum extension with respect to the shear plane (Φ) can be related to width of shear deformation (W) and the amount of displacement due to shear (d) via:

$$\tan(2\Phi) = 2W/d \quad [\text{Molnar } et al., 1999] \quad (1.1)$$

(also see section 3.4). Pn velocities estimated on transect 2 (T_2) and T_1 across the central South Island by Scherwath *et al.* [2002] and Van Avendonk *et al.* [2000] respectively found low Pn velocities compared to typical mantle velocities of 8 km/s within a zone distributed over 100 km west of the Alpine fault, suggesting shear-strain is distributed in the west offshore, producing mantle anisotropy. Assuming the ratio of P-wave to S-wave anisotropy is equal to ~ 1.4 (from mantle xenoliths, Ismail and Mainprice [1998]), the percent S-wave anisotropy associated with this distributed shear zone ranges from $\sim 7 \pm 3\%$ to $9.5 \pm 5\%$. If 1.8 s shear-wave splitting is assumed [Klosko *et al.*, 1999], then the thickness (see equation 3.17, Vinnik *et al.* [1989a]) of this anisotropic layer may range either from 85 to 212 km or from 58 to 188 km. This implies that the majority of the anisotropy could plausibly reside in the mantle lithosphere. Usually dislocation creep governs development of lattice preferred orientation in the mantle material. However dynamic recrystallization can be activated under favorable conditions (see section 2.5.1), and fabrics produced via

dynamic recrystallization is highly anisotropic ($>10\%$). Resulting fast alignment is more consistent with the shear plane compared to anisotropy produced via dislocation creep [see *Zhang and Karato*, 1995, Fig. 4]. Thus, inferred higher Pn anisotropy (11-13 %) and higher shear strains (2.5 - 4.6) in the west coast suggest that the anisotropy in the mantle lithosphere in the central plate-boundary zone may have been influenced by temperature-dependent dynamic recrystallization.

Pn anisotropy estimation at the intersections of 4E with T₁ and T₂ (*Baldock and Stern* [2005]; Fig. 1.5) suggested apparent changes in Pn velocities. On T₂ velocities change from $\sim 7.95 \pm 0.05$ km/s to typical mantle Pn velocities of 8.45 ± 0.05 km/s about 235 km east of the plate-boundary. This change has been interpreted as the limit of the plate-boundary induced lithospheric shear. Notably, the intersection of 4E with T₁ and T₂ yield null anisotropy and the most likely explanation for such a change is the absence of strain-induced anisotropy in the mantle lithosphere in the east. However, null anisotropy is also possible if the upper-mantle is laterally heterogeneous or if the source of anisotropy is deeper than the P_n paths. The width of the shear deformation in the central South Island (region north of T₂) is, therefore, most likely to be within ~ 200 km east of the plate-boundary.

Both *Scherwath et al.* [2002] and *Baldock and Stern* [2005] concluded that some degree of dynamic recrystallization in the mantle lithosphere of the central South Island occurs because of the narrower deformation zone (<200 km) in the central South Island, higher Pn anisotropy 30 km west offshore [*Scherwath et al.*, 2002] and changes in SKS fast orientations from the Alpine fault subparallel (30° counter clockwise from the Alpine fault) to more fault-parallel orientation in the central South Island. *Bourguignon et al.* [2007b] found maximum 7-13% Pn anisotropy ~ 80 km east of the Alpine fault on the Fiordland-Cheviot profile (Fig. 1.5) compared to higher (11-13 %) Pn and null measurement 30 km west of the Alpine fault [*Scherwath et al.*, 2002] and ~ 235 km east of the fault respectively [*Baldock and Stern*, 2005]. As constrained by *Bourguignon et al.* [2007b], if anisotropy is homogeneous and is

7-13% P_n anisotropy, a ~ 100 km thick anisotropic layer is needed to explain the SKS splitting in South Island.

Petrophysical constraints

Because of the high seismic anisotropy inferred from in-situ studies, several laboratory experiments were carried out to study petrophysics of the crust and mantle rocks in South Island. Compressional and shear-wave anisotropy measurements in the Alpine Fault Zone, the Haast schist terrane, and the Torlesse greywacke terrane (Figures 1.4a) determined an increase in anisotropy towards the Alpine fault, implying the rocks with stronger metamorphic foliations could produce a significant anisotropy at crustal depths [Okaya *et al.*, 1995]. This study also noticed that anisotropy in the Alpine fault mylonites and adjacent Haast schist can be as high as 17% (compressional wave) at mid-crustal pressures and concluded that structural orientation (e.g. changes in orientation of the foliations from vertical to sub-horizontal from the Alpine Fault towards Canterbury plains) with respect to ray propagation direction may have a considerable effect on the amount of anisotropy observed.

A more focused petrophysical analysis of the Haast schist was done by Godfrey *et al.* [2000], comparing the Haast schist measurements with other global schist measurements. The study suggested that significant crustal anisotropy can be contributing to SKS splitting, if near-vertical ray paths of SKS phases propagate through the anisotropic material with a sub-horizontal/horizontal symmetry axis and sample the whole thickness. This study clearly demonstrated how split delays can vary as a function of angle (ψ) between ray path (in this case vertical ray paths) and the axis of symmetry in 5 km, 10 km and 20 km anisotropic layer thicknesses (Fig. 1.6).

Curve A-1 in figure 1.6 represents the predicted delay time variation in the Haast

Figure 1.6 Expected delay times from the split SKS waves that travel vertically through schist samples with varying orientations of symmetry axis from the vertical [Godfrey *et al.*, 2000]. The lines with square symbols (A-1) show the estimated delay times in the Haast schist in South Island. The other two lines represent the expected delay times in similar schist terranes in Alaska (TA-Coldfoot Schist) and California (LA-Pelona Schist). Three measurements at each site indicate the estimations at depths 5 km, 10 km and 20 km. Note that vertical ray paths (Haast schist: A-1) yield the highest delay times when the axis of symmetry is subhorizontal/horizontal (i.e. 90° from the vertical).

schist. The rays traveling vertically through 20 km anisotropic layer thickness can produce $\sim 0.5\text{--}0.8$ s splitting delay time if the symmetry axis is horizontal to sub-horizontal ($\psi > 70^\circ$; i.e. sub-/vertical foliation), whereas rays can yield little ($< \sim 0.2$ s) or no splitting if the symmetry axis is sub-vertical to vertical [Godfrey *et al.*, 2000]. These observations emphasize the importance of considering ray propagation direction with respect to the symmetry axis when evaluating splitting measurements, especially in the case of local and regional shear-wave splitting that involve wide angle ray paths.

S-wave anisotropy determined from the petrophysical analysis of mantle xenoliths in South Island is about 5% for waves traveling in the foliation plane perpendicular to the flow direction [Duclos *et al.*, 2005]. Lack of splitting is also possible if the S-wave propagation is at 45° to the foliation plane [Duclos *et al.*, 2005]. The inferred layer thickness of anisotropy is about 150 km if we assume 1.5 s average S-wave splitting for

South Island [Duclos *et al.*, 2005]. Petrological information acquired from peridotites indicated high-temperature deformation of olivine with pronounced foliation and lineation. Lineations are aligned with the $[100]$ axis and foliations are normal to $[010]$ axis of the olivine crystals, suggesting the dominant slip system is $(010)[100]$ at high temperatures. However, at high temperatures $(100)[001]$ slip systems have also been observed from orthopyroxene in some samples. Duclos *et al.* [2005] noticed stronger lithospheric anisotropy in South Island ($\sim 5\%$ S-wave anisotropy with ~ 1.5 s splitting) compared to $\sim 3.5\%$ S-wave anisotropy (0.8 s splitting) in North Island. A possible explanation given for this discrepancy is that intense deformation and vertical foliations around the Alpine fault result from dextral-shear between the AU and PAC lithospheric blocks.

Figure 1.7(a,b,c) shows the dynamic model proposed for the central South Island that Duclos *et al.* [2005] used to explain the petrophysical measurements of mantle Xneoliths, P_n anisotropy and SKS splitting observations. The model consists of a narrow zone of shear to Moho depths that explains the dextral shearing of the Alpine fault, but a broad zone (as wide as the entire South Island) of shear deformation in the deep mantle-lithosphere as in the case of ‘wrench-fault type’ deformation [Vauchez and Tommasi, 2003]. Present day structure consists of an asymmetric anisotropic body as a result of the change in fault movement (pure strike-slip to oblique strike-slip) and collision in the central part. During strike-slip motion of the Alpine fault, this body was symmetric (1.7-a). But now it has a shallower upper-limit on the Australian side, and deeper upper-surface (deeper than Moho depth) and more distributed configuration on the Pacific side (1.7-c). This present-day model explains the higher P_n anisotropy on Australian side [Scherwath *et al.*, 2002], null P_n anisotropy [Baldock and Stern, 2005] about 235 km east from the plate-boundary, and the distributed zone of plate-boundary parallel shear as determined from SKS splitting [Klosko *et al.*, 1999]. The model further suggests that the plate-boundary (the Alpine fault) parallel shear is mostly in the upper-mantle lithosphere,

but can extend into the asthenosphere.

Figure 1.7 Dynamic model proposed by *Duclos et al.* [2005] to explain the seismic anisotropy measurements in central South Island. (A) Symmetric anisotropic body due to wrench-fault type deformation when the fault was pure strike-slip (before ~ 7 Myr). (B) Accumulation of asymmetric nature of the anisotropic body as a result of the oblique collision and subduction processes. (C) Present-day asymmetric and distributed anisotropic body. Note that the upper-surface of this body is shallower on the Australian side than in the Pacific plate. 1, 2, and 3 indicate the different stages of the Alpine fault evolution.

Constraints from crustal finite strain estimations and geodetic data

How the shallow-crust and surface deform in response to plate-boundary deformation in the mantle-lithosphere has been investigated quantitatively by geodetic studies (e.g. velocity field determination from GPS) and structural analysis (e.g. finite strain estimation from geological markers). Unlike the GPS-based geodetic studies that consider the present-day deformation patterns on the Earth's surface, bending related geological markers may have recorded multiple deformation histories and therefore provide more accurate and detailed constraints on the strain history and associated mantle fabrics [Little *et al.*, 2002b]. Bending-related geological markers in four domains that are defined with an increase in distance from the Alpine fault [see **D**, **C**, **B**, and **A** domains in Little *et al.*, 2002b, Fig. 4] indicated an increase in finite strain towards the fault. According to this study, little or no bending-related finite strain is found in the farthest domain from the Alpine fault (domain A), where there is still a considerable shear-wave splitting, suggesting weak coupling of shallow-crustal and upper-mantle deformation. Maximum horizontal stretch calculated from the domains that are close to the Alpine fault are consistent with the SKS-splitting directions (i.e. $\sim 30^\circ$ anticlockwise from the Fault). This suggests that crust and mantle deform coherently towards the fault in this region, where SKS splitting is assumed to follow finite strain due to shear at depth. According to structural geology estimates, a low shear strain value of ~ 1 -2 [Little *et al.*, 2002b] is too low to promote deformation via dynamic recrystallization and, thus, whether dynamic recrystallization itself explains the high P_n anisotropy and the deflection of SKS fast azimuth towards the fault in the central South Island is unresolved.

Strain estimations from geodetic velocities at the surface are assumed to provide inconsistent constraints on the deformation distribution at depths as they reflect the elastic strain accumulation due to continuous slip at deeper part of the fault [Ellis *et al.*, 2006, Moore *et al.*, 2002]. Most of these studies are based on the assumption

Figure 1.8 (a) Comparison of GPS velocity data (gray symbols) and the predicted horizontal surface velocities using the analytical solution by *Beavan et al.* [2004] and models with boundary conditions that satisfy the elastic half-space (HS) and mantle-shortening (MS). (b) The same comparison using a model with boundary conditions of mantle-thickening (MT) model [*Stern et al.*, 2000]. Velocity predictions from MT model with a width of distributed deformation (L) = 100 km fit well with the observed data compared to a model with L = 200 km. Figure after *Ellis et al.* [2006].

that strain estimations from the present day velocity field is representative of the crustal deformation during **My** time-scales [*Beavan et al.*, 1999, *Houlié and Stern*, 2012]. Strain estimations from GPS velocity measurements in the central South Island indicate surface deformation due to relative plate motion of AUS-PAC plates is distributed in a zone of ~ 80 km width [*Moore et al.*, 2002]. Campaign GPS surveys in the central South Island also suggested a narrow zone of shear-strain parallel to

the Alpine fault in the upper-mid crust and a distributed zone (~ 100 km east) of fault-perpendicular deformation [Beavan *et al.*, 2004, 1999]. Both Moore *et al.* [2002] and Beavan *et al.* [2004] predicted the possibility of distributed fault-parallel shear deformation over a 36 km wide region at deeper depths ($\gtrsim 18$ km), however the oppositely dipping slip patch, which has been modelled in the deeper section, is attributed to either subducted Pacific mantle-lithosphere or an artifact of the distributed shear [Beavan *et al.*, 2004, 1999]. Horizontal surface velocities predicted from the finite-element modeling agree well with the interseismic velocities estimated from the GPS data regardless of the model assumption (i.e. distributed shear or localized models-Fig. 1.8). The present-day deformation is limited to 100 km distance from the Alpine fault as implied by both GPS estimations and finite element modeling. Those predictions do not agree with the model with ~ 200 km width of distributed lithospheric shear that was constrained by the shear-wave splitting and Pn anisotropy studies [Klosko *et al.*, 1999, Scherwath *et al.*, 2002]. Nevertheless, the recent study by Houlié and Stern [2012] used the direction of extension axis derived from GPS data (i.e. direction perpendicular to the compression direction) as a proxy for the direction of shear to measure the angle between SKS splitting directions and the shear plane (Φ), and to determine the width of deformation using the relationship given in equation 1.1. This study divided deformation in South Island into two distinct domains according to Φ . In the northern and upper central South Island, Φ is estimated to be 28° and interpreted as a product of finite strain during simple shear. Here, shear deformation is estimated to distribute in at least in a ~ 200 km wide zone. Φ is found to be smaller ($\Phi = 3^\circ$) in the southern and lower-central South Island than the northern South Island and is attributed to either much higher strain or the combined effects of pure shear and simple shear deformation [Houlié and Stern, 2012].

1.6 Previous anisotropy studies in the southern Hikurangi subduction zone

Previous studies by *Gledhill and Stuart* [1996] and *Gledhill and Gubbins* [1996] have investigated shallow and deep anisotropy respectively in the eastern part of the southernmost North Island using the data recorded on the Tararua array (Fig. 3.6). *Gledhill and Stuart* [1996] found little crustal anisotropy (~ 0.2 s) in the overriding Australian plate, but a clear increase in delay time with depth (~ 1.4 % S-wave velocity anisotropy). The 1.5 ± 0.4 s δt and $28^\circ \pm 4^\circ$ ϕ of SKS splitting anisotropy calculated from the same station network [*Gledhill and Gubbins*, 1996] suggested trench parallel flow governs the anisotropy in this region, and backazimuthal dependence suggested a dipping axis of symmetry or multilayer anisotropy. Follow-up studies by *Marson-Pidgeon and Savage* [1997], *Marson-Pidgeon and Savage* [2004a], *Marson-Pidgeon et al.* [1999], and *Marson-Pidgeon and Savage* [2004b] provide an expanded set of splitting measurements from various S-phases (SKS, ScS and teleseismic S) recorded on the SNZO station in Wellington, the POMSII-array that was distributed in the lower-central North Island and the Tararua array (Fig 6.2). The POMSII-array only had three stations (LBLU, LKIR and LMAU) in our region of interest. Splitting anisotropy studies at the SNZO found frequency dependent splitting from SKS, teleseismic S and ScS. Small delays observed from ScS phases from both the SNZO [*Marson-Pidgeon and Savage*, 1997] and the Tararua array [*Marson-Pidgeon et al.*, 1999] were attributed to small-scale heterogeneities. Consistent orientation of ϕ (NE-SW) through out the eastern part of the lower North Island suggest that a similar process controls the anisotropy in the mantle wedge and sub-slab region. It is possibly caused by trench-parallel flow.

Comparing S-wave splitting with the maximum P-velocity directions of NE-SW ($\sim 30^\circ$) in the southwest Pacific [*Galea*, 1993, *Shearer and Orcutt*, 1985], *Marson-Pidgeon et al.* [1999] argued for the possibility of observed splitting to be occurring

within the subducting oceanic lithosphere as a result of the fossil anisotropy at the ancient spreading margins in the Pacific ocean. However, modeling of splitting parameters with respect to anisotropic thicknesses in the slab (or the length of the ray path within the slab) in different regions found that the model predictions do not match the observed splitting, suggesting anisotropy does not entirely reside in the slab, but may contribute significant amounts to the total anisotropy [Faccenda *et al.*, 2008]. Modeling of P-wave dispersion in Wellington region [Gubbins and Snieder, 1991] suggested a thin (6-15 km) high-velocity lid on top of the subducting lithosphere that possibly resulted from the eclogitization of the subducting gabbroic crust. Deformed eclogite rocks are considered to be strongly anisotropic [Zhang *et al.*, 2008]. According to Gledhill and Gubbins [1996], stress-induced fluid filled cracks may also play a dominant role in the Hikurangi subducting slab. Matcham *et al.* [2000] found $4.4 \pm 0.99\%$ velocity anisotropy with ϕ of 29° from the subducting slab earthquakes between 20 - 70 km depth range in the Wellington region, supporting the idea of crack-induced anisotropy in the slab or implying high anisotropy in the slab. This observation also supports the previous findings of increase in δt with depth in the same region by Gledhill and Stuart [1996] and Brisbane *et al.* [1999].

In general, the splitting observations in this region are assumed to agree with the trench parallel flow model (see section 2.7.2). Consistent NE-SW ϕ from both shallow and deep anisotropy investigations parallel to structural and geological trends and strike of the subduction zone further imply that this region most likely follows the vertical coherent deformation model [Silver, 1996]. Nevertheless, the observed small-scale variations in δt and ϕ [Marson-Pidgeon *et al.*, 1999, Marson-Pidgeon and Savage, 2004a] with azimuth have been attributed to a structure with more than a single layer of anisotropy or plunging symmetry axis [Gledhill and Gubbins, 1996]. Modeling of seismic anisotropy measurements from shear wave splitting (both local and SKS) and anomalous surface wave polarizations

(fundamental modes of Love waves) observed at the Tararua array, POMS II array and at SNZO in the south of the North Island found a best-fit model with plunging hexagonal symmetry axis ($\sim 30^\circ$ SE to fit surface wave measurement and $\sim 15^\circ$ SE to fit local S-wave splitting measurements) in the overriding Australian plate [Brisbourne *et al.*, 1999]. The discrepancy between surface wave anisotropy and S-wave splitting has been explained in terms of frequency dependent splitting that was determined by *Marson-Pidgeon and Savage* [1997] at SNZO. Slab and sub-slab anisotropy in the proposed model are assumed to have horizontal symmetry axes to explain the increase in δt with depth. A receiver function study by *Savage et al.* [2007], however, used an upper anisotropic layer with horizontal symmetry axis and a metasediment layer with vertical symmetry axis in the lithospheric mantle-wedge to explain the delay in Ps arrival with a polarity change.

In contrast to consistent splitting parameters observed from previous SKS splitting investigations at Tararua array, *Marson-Pidgeon and Savage* [2004b] found noticeable lateral variations in the splitting parameters from ScS and teleseismic S waves along the profile that is aligned in NE-SW arm of the Tararua array. This study also noticed variable splitting parameters from different events recorded at the same station. These variations have been integrated with the backazimuthal variations in δt to show that lateral variations in anisotropic structure exist across the array. They suggest a distinct anisotropic boundary that runs in a $\sim$ NE-SW direction from ~ 40 km east of the Tararua array towards Kapiti Island in the western offshore region. The study also pointed out that it is important to investigate seismic anisotropy in the eastern side of the Tararua array to provide more constraints on the spatial extent of this anisotropic body.

Chapter 2

Seismic anisotropy and its relationship to tectonic deformation and dynamics

2.1 Background

Seismic anisotropy can be defined as directionally dependent elastic response of seismic waves to the anisotropic nature of the Earth's interior. The properties of seismic waves are primarily controlled by the elastic properties of the Earth's materials. *Hess* [1964] first used azimuthal variations in P_n velocities (also known as propagation anisotropy) to detect seismic anisotropy in the ocean lithosphere. Afterwards, several other methods were introduced to investigate anisotropy using shear-waves and surface waves. These approaches are sensitive to different vertical and lateral scales of anisotropy. Among these, shear-wave splitting is a well-established and commonly used technique in exploring seismic anisotropy. Compared to other methods (e.g. P-wave velocity anisotropy, surface-wave polarization anisotropy), this method is less influenced by lateral heterogeneities [*Long and Silver*, 2009].

2.2 Concepts of wave propagation in anisotropic media

This section provides some of the important aspects of elastic wave propagation in homogeneous anisotropic media and explains the shear-wave splitting phenomenon. In an anisotropic medium, behavior and properties of the seismic waves strongly depend on the direction of propagation and the elastic properties of the material [*Babuska and Cara*, 1991]. Seismic waves attain different speeds depending on the

propagation direction of the waves in such a medium. To fully understand the seismic wave propagation and polarization in anisotropic media, it is important to understand the elastic properties of the media. The annotations and formulas discussed in the subsections of section 2.2 are taken from *Babuska and Cara* [1991] and *Cerveny* [2001].

2.2.1 Elastic tensors

When materials are subjected to stress, they are strained (or deformed). For small elastic deformation, the generalized Hooke's law (equation 2.1) gives the linear relationship between stress and strain in an anisotropic medium:

$$\sigma_{ij} = c_{ijkl}\epsilon_{kl}, \quad (2.1)$$

where σ_{ij} and ϵ_{kl} are Cartesian components of stress and strain tensors. c_{ijkl} is the component of the elasticity tensor, which is also known as a stiffness tensor. The Einstein summation convention is used, in which repeated indices represent summation (e.g. $\sigma_{ij} = \sum_{k,l=1}^3 c_{ijkl}\epsilon_{kl}$).

Stress, strain and stiffness tensors

Both stress and strain tensors are symmetric (i.e. $\sigma_{ij} = \sigma_{ji}$ and $\epsilon_{kl} = \epsilon_{lk}$). For small strains, a strain tensor can be expressed using a displacement vector ($\mathbf{u}_r$) which is a function of the position of the medium (equation 2.2):

$$\epsilon_{kl} = \frac{1}{2}(\partial_l \mathbf{u}_k + \partial_k \mathbf{u}_l), \quad (2.2)$$

where $\partial_l \mathbf{u}_k$ and $\partial_k \mathbf{u}_l$ are spatial derivatives with respect to x_l and x_k coordinates, respectively.

A stiffness tensor (c_{ijkl}) is a 4th order tensor with 81 (3^4) elastic coefficients. However, it can be denoted using 36 coefficients (see c_{ijkl} matrix) due to the fact that stress and strain tensors are symmetric (i.e. $\sigma_{ij} = \sigma_{ji}$ and $\epsilon_{kl} = \epsilon_{lk}$).

$$c_{ijkl} = c_{jikl},$$

$$c_{ijkl} = c_{jilk},$$

$$c_{ijkl} = \begin{pmatrix} c_{1111} & c_{1122} & c_{1133} & c_{1123} & c_{1113} & c_{1112} \\ c_{2211} & c_{2222} & c_{2233} & c_{2223} & c_{2213} & c_{2212} \\ c_{3311} & c_{3322} & c_{3333} & c_{3323} & c_{3313} & c_{3312} \\ c_{2311} & c_{2322} & c_{2333} & c_{2323} & c_{2313} & c_{2312} \\ c_{1311} & c_{1322} & c_{1333} & c_{1323} & c_{1313} & c_{1312} \\ c_{1211} & c_{1222} & c_{1233} & c_{1223} & c_{1213} & c_{1212} \end{pmatrix}$$

Stress (σ_{ij}) can be expressed by means of a derivative of strain energy density (i.e. work done per unit volume to deform a material from a stress free state to stressed state; U) with respect to strain, and can be related to c_{ijkl} using the following relationships:

$$\sigma_{ij} = \partial U / \partial \epsilon_{ij}$$

$$c_{ijkl} = \partial \sigma_{ij} / \partial \epsilon_{kl}$$

c_{ijkl} , thus, can be expressed via partial derivatives of U and strain:

$$c_{ijkl} = \partial^2 U / \partial \epsilon_{kl} \partial \epsilon_{ij}$$

$$\partial^2 U / \partial \epsilon_{kl} \partial \epsilon_{ij} = \partial^2 U / \partial \epsilon_{ij} \partial \epsilon_{kl}$$

$$c_{ijkl} = c_{klij}$$

Considering the above symmetry condition ($c_{ijkl} = c_{klij}$), components of the stiffness tensor will be reduced further into 21 elastic coefficients. Then the resultant matrix (c_{mn}) with elastic coefficients can be written as:

$$c_{mn} = \begin{pmatrix} c_{1111} & c_{1122} & c_{1133} & c_{1123} & c_{1113} & c_{1112} \\ & c_{2222} & c_{2233} & c_{2223} & c_{2213} & c_{2212} \\ & & c_{3333} & c_{3323} & c_{3313} & c_{3312} \\ & & & c_{2323} & c_{2313} & c_{2312} \\ & & & & c_{1313} & c_{1312} \\ & & & & & c_{1212} \end{pmatrix}$$

These 21 elastic coefficients in the stiffness tensor describe the elastic behavior of the most general anisotropic solid that has the lowest degree of symmetry (i.e. triclinic).

2.2.2 Elastic waves in homogeneous anisotropic media

Spatial variations in stress ($\partial_j \sigma_{ij}$) and time variations in displacement ($\ddot{\mathbf{u}} = \partial^2 \mathbf{u}_i / \partial t^2$) are related to each other via an elastodynamic equation [Cerveny, 2001]:

$$\partial_j (\sigma_{ij}) + \mathbf{f}_i = \rho \ddot{\mathbf{u}}_i, \quad (2.3)$$

where $\mathbf{f}$ is a vector that defines the force per unit volume. If no external forces are acting on the medium, $\mathbf{f}_i = 0$. Combining equation 2.3 with equations 2.1 and 2.2, the elastodynamic equation for homogeneous anisotropic media can be expressed as:

$$c_{ijkl} \partial_j \partial_k (\mathbf{u}_l) = \rho \ddot{\mathbf{u}}_i, \quad (2.4)$$

Next, we will consider the plane wave, which is the most fundamental solution of the elastodynamic equation, to explain the properties of elastic waves in the

homogeneous anisotropic media. The plane wave solution in a homogeneous anisotropic medium can be expressed as the functions of space ($\vec{x}$) and time (t):

$$\mathbf{u}(\vec{x}, t) = \mathbf{A}f(t - \vec{N} \cdot \vec{x} / \nu), \quad (2.5)$$

where $\mathbf{A}$ is the amplitude of particle motion, $\vec{N}$ is the unit vector normal to wave propagation direction and ν is the phase velocity. Partial derivatives of time and space obtained from plane wave solution (equation 2.5) are then combined with equation 2.4 to examine the properties of plane waves in homogeneous, elastic anisotropic media (equation 2.6).

$$c_{ijkl} \mathbf{N}_j \mathbf{N}_k \mathbf{a}_l = \rho \nu^2 \mathbf{a}_i, \quad (2.6)$$

Defining a density normalized stiffness tensor (c_{ijkl}/ρ) as C_{ijkl} , the Christoffel equation (equation 2.7) with the Christoffel matrix Γ_{il} ($\Gamma_{il} = C_{ijkl} \mathbf{N}_j \mathbf{N}_k$) can be obtained from the equation 2.6.

$$\Gamma_{il} \mathbf{a}_l = \nu^2 \mathbf{a}_i \quad (2.7)$$

Here, ν^2 and $\mathbf{a}_i$ are the eigenvalues and the eigenvectors of Γ_{il} . From this equation we can obtain three eigenvalues and three mutually perpendicular eigenvectors that correspond to the three different wave types in a homogeneous anisotropic medium.

2.2.3 Plane waves in hexagonal anisotropy systems

Transversely isotropic materials that are characterized by a single plane of isotropy and one axis of rotational symmetry exhibit *hexagonal symmetry*. Thus, such systems are also known as hexagonal anisotropy systems. The 21 elastic coefficients discussed above in a general anisotropic medium are further reduced depending on

the symmetry of the system. In an isotropic case, only two elastic coefficients (also known as *Lamé elastic moduli*) are enough to fully describe the elastic properties of the medium. Hexagonal anisotropy symmetry is the most common system that is used to understand seismic wave propagation in different Earth media. Therefore, we will consider the plane wave propagation in a hexagonal anisotropy system in detail. Horizontal lithologic layering with a vertical symmetry axis (Figure 2.1) and vertically aligned cracks with a horizontal symmetry axis (Figure 2.2a) are two natural examples of hexagonal symmetry. To characterize the elastic properties of such a system, one must have five elastic coefficients (A , F , C , L , and N) [Love, 1944]. The stiffness tensor of the hexagonal anisotropy system can be denoted using the above coefficients as given below:

$$c_{mn} = \begin{pmatrix} A & A - 2N & F & & & \\ A - 2N & A & F & & & \\ F & F & C & & & \\ & & & L & & \\ & & & & L & \\ & & & & & N \end{pmatrix}$$

where, A , F , C , L , N and $A - 2N$ are also known as Love's coefficients after Love [1944].

Let us characterize the properties of plane waves that propagate in two main directions: **(a)** Plane wave propagating in the direction of a plane perpendicular to the symmetry axis (blue arrow in figure 2.1) **(b)** Wave propagating in the direction parallel to the symmetry axis (red arrow in figure 2.1) using the above c_{mn} matrix of the hexagonal system.

a. Wave propagating in the plane perpendicular to the symmetry axis:

For this type of wave propagation system, the Christoffel matrix (Γ_{il}) yields:

$$\Gamma_{il} = \frac{1}{\rho} \begin{pmatrix} A & 0 & 0 \\ 0 & N & 0 \\ 0 & 0 & L \end{pmatrix}$$

Figure 2.1 Layered anisotropic system with hexagonal symmetry (also equivalent to transverse isotropy with vertical axis of symmetry). x_3 axis denotes the symmetry axis. Blue and red arrows indicate the S-wave propagation parallel to the plane perpendicular to the symmetry axis and parallel to the symmetry axis, respectively. The presence of quasi S phases upon propagating through the media is denoted by S_f (fast) and S_s (slow).

This matrix has three different eigenvalues λ_1^2 , λ_2^2 and λ_3^2 that correspond to the three eigenvectors a_1 , a_2 and a_3 , respectively. These eigenvectors indicate the polarization directions of the three different wave types with varying velocities along x_1 direction:

$$(a_1) = \begin{vmatrix} 1 \\ 0 \\ 0 \end{vmatrix} \quad (a_2) = \begin{vmatrix} 0 \\ 1 \\ 0 \end{vmatrix} \quad (a_3) = \begin{vmatrix} 0 \\ 0 \\ 1 \end{vmatrix}$$

The phase velocities of the waves are given by:

$$\lambda_1 = \sqrt{\frac{A}{\rho}}$$

$$\lambda_2 = \sqrt{\frac{N}{\rho}}$$

$$\lambda_3 = \sqrt{\frac{L}{\rho}}$$

Here, λ_1 is the velocity of the fastest wave, which is polarized parallel to the propagation direction. Thus, this wave denotes the P wave. The other two mutually perpendicular phases are polarized perpendicular to the wave propagation direction (S waves) and travel with different velocities (λ_2 and λ_3). In the absence of anisotropy (or isotropic case), $\lambda_2 = \lambda_3$ and there will only be a single phase that is polarized in the direction perpendicular to the direction of wave propagation. This phase represents the unsplit shear (S-wave) wave.

Thus, if a wave propagates perpendicular to the symmetry axis there will be two mutually perpendicular quasi S phases (S_f and S_s in figure 2.1) that are polarized perpendicular to the wave propagation direction. This phenomenon is known as shear-wave splitting.

b. Wave propagation in the plane parallel to the symmetry axis:

The Christoffel matrix (Γ_{il}) for the wave propagated in x_3 direction is given by:

$$\Gamma_{il} = \frac{1}{\rho} \begin{pmatrix} L & 0 & 0 \\ 0 & L & 0 \\ 0 & 0 & C \end{pmatrix}$$

$$(a_1) = \begin{vmatrix} 1 \\ 0 \\ 0 \end{vmatrix}$$

$$(a_2) = \begin{vmatrix} 0 \\ 1 \\ 0 \end{vmatrix}$$

$$(a_3) = \begin{vmatrix} 0 \\ 0 \\ 1 \end{vmatrix}$$

The phase velocities of the waves in each direction are given by:

$$\lambda_1 = \sqrt{\frac{L}{\rho}} \qquad \lambda_2 = \sqrt{\frac{L}{\rho}} \qquad \lambda_3 = \sqrt{\frac{C}{\rho}}$$

There are two degenerate eigenvalues (λ_1^2 and λ_2^2), which represent the velocity of the unsplit S-wave that is polarized perpendicular to the symmetry axis (S in figure 2.1). The other eigenvalue (λ_3) is the velocity of the P-wave that is polarized parallel to the propagation direction or the symmetry axis (x_3). Except when a wave's propagation direction is along the symmetry axis, the other propagation directions in the anisotropic medium will display S-waves splitting, but the propagation velocities of split S-waves may vary depending on the angle between symmetry axis and the ray propagation direction (Fig. 2.2 and Fig. 1.6).

Figure 2.2 S-wave splitting in different anisotropic media when propagation in the plane perpendicular to the symmetry axis (x_3). Splitting of S-wave is marked in blue. δt indicates the time lag between fast (S_f) and slow (S_s) split S-waves, and ϕ denotes the polarization direction of the fast split wave. (a) a system with vertically aligned cracks. SH_{max} denotes the maximum compressive direction. (b) mineral alignment due to flow or shear.

2.3 Shear-wave splitting and splitting parameters

The presence of quasi shear-waves is a distinct property of elastic waves traversing an anisotropic medium, satisfying the relationship between the direction of propagation and symmetry of the medium. Most importantly, this property can be used to

distinguish anisotropy from isotropy and provide detailed information about the anisotropic medium. Because polarization directions of the split waves are controlled in part by the alignment of the anisotropic medium, this is also known as polarization anisotropy.

As highlighted in section 2.2.3, fast (S_f) and slow (S_s) split S-waves are polarized in mutually perpendicular directions depending on the alignment of the anisotropic medium and the wave propagation direction (Figure 2.2). Usually, the polarization direction of the S_f wave (ϕ) is related to the alignment of the anisotropic medium, and the delay time between S_f and S_s depends on the extent and/or degree of anisotropic medium and path length of the ray within the medium, but it also depends on the propagation direction [Savage, 1999]. When these two split waves propagate with two different velocities, β_f (velocity of S_f wave) and β_s (velocity of S_s wave), they will accumulate delay time (δt) along the travel path. If the length of the travel path is l and the isotropic velocity of the shear-wave is β_0 , for small anisotropy the relationship between δt and l is given by the following linear equation [Silver and Chan, 1991]:

$$\delta t = l \delta \hat{\beta} / \beta_0 \quad (2.8)$$

Here, $\delta \hat{\beta}$ is a fraction of velocity anisotropy calculated from the S-wave splitting (i.e. $\delta \hat{\beta} = (\delta \beta_f - \delta \beta_s) / \beta_0$; $\beta_f = \beta_0 + \delta \beta_f$ and $\beta_s = \beta_0 - \delta \beta_s$). Accordingly, $\delta t^\dagger$ is proportional to l . Thus, δt and $\phi^\dagger$ are considered as the main splitting parameters that could be determined from shear-wave splitting to investigate the subsurface anisotropic structure in homogeneous media.

[†]hereafter, δt denotes delay time/s and ϕ denotes fast direction/s.

2.4 Origin of anisotropy

Anisotropy occurs at various depths at different length scales depending on the elastic properties of the materials and the processes associated with the formation (e.g. geodynamic and deformational process). Anisotropy has two main forms: 1) extrinsic and 2) intrinsic anisotropy. Extrinsic anisotropy is related to the structural arrangement of the Earth's material, therefore, it is also known as shape preferred orientation (SPO). Intrinsic anisotropy is predominantly caused by the lattice preferred orientation (LPO) of intrinsically anisotropic minerals [*Kendall, 2000, Montagner and Guillot, 2002*].

SPO fabrics are mainly observed at crustal depths and possess either stress or structural origin. Stress-induced anisotropy is observable under low confining pressure (<200 Mpa) [*Kern and Wenk, 1990*] in the uppermost crust as a result of the preferential orientation of fluid-filled (usually water or melt) micro-cracks (Figure 2.2a). In addition to stress controlled sources, structural anisotropy is also present at various scales at crustal depths. Macro-scale structures such as faults, bedding planes and foliations are some of the examples for anisotropy caused by structure. Most of these structurally controlled fabrics are also associated with LPO fabrics at mid-lower crustal depths. LPO anisotropy is the dominant source of anisotropy in the upper mantle; it also has a significant effect on the anisotropy in the lower crust [*Mainprice and Nicolas, 1989*].

2.4.1 Extrinsic anisotropy

a) Stress-induced anisotropy

Stress-induced anisotropy is the most abundant source of anisotropy in the upper crust where there is a low confining pressure or high fluid pressure. Earlier studies on elastic properties of rocks under hydrostatic stress suggest that the elastic properties

of rocks under low effective stress ($\lesssim 1$ kb) are primarily controlled by the properties of the micro-cracks. The effect of non-hydrostatic stress (or differential stress) on velocity anisotropy was first studied by *Nur* [1971]. This study showed that when rocks are subjected to non-hydrostatic stress, cracks can be closed preferentially, leaving some cracks that remain open in another direction. This mechanism may change the rock from its isotropic state to an anisotropic state as result of the change in elastic properties within a given direction of the medium. In such a crack-induced anisotropic medium shear-wave splitting is possible and depends on the crack alignment and the stress direction (Figure 2.2a). The polarization direction of the fast shear-wave (ϕ) in such an anisotropic medium is usually parallel to the alignment (or the strike) of the cracks remaining open and hence parallel or sub-parallel to the horizontal maximum compressive direction (SH_{max}) [*Crampin*, 1978, *Crampin and Lovell*, 1991]. *Hudson* [1980] provides the effective elastic constants of the cracked medium in terms of crack density, suggesting that the shear-wave splitting from cracked media does not depend on the crack dimensions, as the wavelength of the shear-waves are much higher than the crack dimensions. However, the crack geometry, crack density and degree of crack saturation (or pore fluid pressure) have significant effects on the shear-wave splitting and the differential shear-wave anisotropy (SWA) [*Crampin*, 1978, 1994].

Crack-induced anisotropy can have different forms such as fluid-filled cracks, micro-cracks and aligned pore spaces. In shallow sedimentary rocks, fluid-filled cracks are the main source of anisotropy [*Mueller*, 1991] and, in igneous and metamorphic rocks anisotropy is attributed to aligned dry cracks and micro-cracks [*Crampin*, 1994] (see table 2.1). It is usually assumed that the split shear-waves from steep arrivals (incidence angle $\lesssim 45^\circ$) are mainly due to vertically aligned cracks at shallow crustal depths [*Leary et al.*, 1990]. This system (i.e. vertical crack planes) acts as a medium with transverse isotropy with horizontal symmetry axis (Figure 2.2a) [*Crampin and Lovell*, 1991, *Crampin*, 1994]. Because the alignment of vertical cracks is primarily

Table 2.1 An average differential SWA measured at different geological terranes (adapted from *Crampin* [1994]).

Rock type	Depth (km)	Percent SWA (%)
Volcanic rocks	10	4.0-5
Volcanic caldera	10	~ 6.5
Granite (geothermal area)	2	8
Metamorphic rocks	5	4
Metamorphic rocks (after shocks area)	20	7
Granulite facies	1	10
Shale and lime stone	1.1-2	2
Fault zone (sedimentary)	0.3-1.4	7-8
Sand stone and clay	0.04 (very shallow)	12-14

controlled by the local stress field, shear-wave splitting in cracked media can be used as a tool to investigate the changes in the local stress field over time. This concept has been successfully tested as a stress monitoring tool at volcanoes [*Miller and Savage*, 2001, *Gerst and Savage*, 2004] and in various other tectonic domains [*Crampin et al.*, 2004, *Peacock et al.*, 1988].

b) Structurally-induced anisotropy

Structural anisotropy is mainly caused by the alignment of foliation planes, bedding planes or macro-scale faults (e.g. San-Andreas fault, Alpine fault). This has been frequently observed at crustal depths. *Barruol et al.* [1997] suggested that foliations control most of the splitting in the mid-lower crust. They explained this in terms of vertically aligned foliations that are developed in the middle to lower crust where there is a high enough temperature to develop foliations from vertically aligned phyllosilicates (e.g. Mica) in felsic rocks. An important observation of this study is that higher δt are recorded from the S-waves that traverse parallel to the foliation plane, whereas no or little splitting occurs if the propagation direction is normal to the foliation planes. Vertical ray paths that travel in regions with steeply dipping foliations (e.g. strike-slip and convergent plate-boundaries) would produce δt within the range of 0.3-0.5 s [*Barruol et al.*, 1997]. This suggests that not only the mineral

composition but also the orientation of foliation plane with respect to ray path of the S-wave, is crucial in producing splitting. Studies also showed that rocks with a single set of foliations produce more pronounced splitting compared to rocks with multiple sets of foliations [Lloyd *et al.*, 2009]. Other than foliated structures, aligned melt-inclusions or melt-bands at the lower-crustal or the upper-mantle depths in the tectonic settings, such as mantle wedge and volcanic terranes, also play a dominant role in producing a structurally-induced anisotropy [Holtzman *et al.*, 2003].

2.4.2 Intrinsic anisotropy

Unlike extrinsic anisotropy, intrinsic anisotropy is independent of the wavelength of the S-wave and results either from strain-induced mineral alignment (also known as lattice preferred orientation) or the preferred orientation of individual mineral grains during the rock's genesis [Kendall, 2000].

Lattice preferred orientation (LPO)

LPO is caused by the preferred orientation of intrinsically anisotropic crystals of rock-forming minerals (Figure 2.2b). Changes in the strain field due to plastic deformation mainly control the preferred orientation of the minerals and the strength of LPO depends on the finite strain. Dislocation creep and diffusion creep mechanisms govern the strain-induced deformation at high temperatures ($\sim 0.5T_m$, where T_m is melting temperature) in rocks (section 2.5; Kameyama *et al.* [1999]). Diffusion creep activates under low stress and/or in material with small grain size. This may lead to the development of randomly oriented fabric with no anisotropy or straining in a preferred direction. On the other hand, dislocation creep plays a dominant role in producing LPO fabrics in the upper-mantle and the lower-crust via strain-induced alignment. This mechanism operates under high stress conditions and/or in media with large grain sizes. However, the resultant anisotropic properties depend on the geometry of deformation and physical/chemical conditions during the

deformation [Karato *et al.*, 2008].

The majority of LPO anisotropy occurs in the upper-mantle and the lower-crust, where the rocks deform in a ductile manner and have anisotropic mineral aggregates. In addition to strain, Temperature (T), pressure (P) and mineralogy are also important factors that control the strength of the LPO fabric. In general, the strength of the LPO increases with increasing T and decreases with increasing P. One of the most important factors affecting the LPO strength is the mineralogy of rocks. Advances in mineral physics and petrographic analysis provide detailed information about the anisotropic properties of minerals (single crystals and polycrystals) and the overall anisotropy of rocks. The findings of these studies are also useful in interpreting an in-situ shear-wave splitting anisotropy to understand the LPO contribution to total splitting. Anisotropy of a single crystal is expressed by percentage anisotropy (A) as given below:

$$A = 200 \left(\frac{V_{max} - V_{min}}{V_{max} + V_{min}} \right) \quad (2.9)$$

where, V_{max} and V_{min} are the maximum and minimum velocities of either the P (V_p) or S (V_s) phases. In the upper-mantle, olivine, enstatite (orthopyroxene) and diopside (clinopyroxenes) are considered to be the main intrinsically anisotropic minerals [Mainprice *et al.*, 2000]. Olivine is the volumetrically most important mineral in the upper mantle rocks that could generate strong LPO fabrics. However, when olivine combines with orthopyroxene and clinopyroxene (e.g. in the upper-mantle rocks), the resultant velocity anisotropy can vary depending on the mineral composition. Polycrystalline aggregates of olivine in the upper-mantle show the highest V_p anisotropy although the single crystal anisotropy of lower-crustal minerals is higher than that of upper-mantle anisotropy [Mainprice *et al.*, 2000]. In the lower-crust, hornblende, plagioclase, and diopside play a dominant role in the development of LPO fabrics (Table 2.2). As suggested by Mainprice and Nicolas

Table 2.2 Single crystal and polycrystal velocity anisotropy (V_p) of intrinsically anisotropic minerals of the lower-crust and the upper-mantle [*Mainprice and Nicolas, 1989, Mainprice et al., 2000*].

Lower crust		
Mineral	Single crystal V_p (%)	Polycrystal V_p (%)
Quartz	24.4	6
Plagioclase	27.4	13
Diopside	29	5
Hornblend	23.9	12
Upper mantle		
Olivine	25	~ 15
Enstatite	15 - 12	~ 5
Diopside	29	5

[1989], monomineralic plagioclase that deforms plastically in the lower crust can also act as a strong source of anisotropy.

2.5 Deformation kinematics and anisotropy

The elastic anisotropy measured from shear-wave splitting is indirect evidence for deformation of the Earth's interior. As explained in sections 2.4.1 and 2.4.2, rocks acquire anisotropic properties due to the preferential orientation of minerals at deep crustal or mantle depths and micro-cracks at shallow depths. Viscoelastoplastic rheology can be used to fully characterize the lithospheric deformation, assuming that the shallow brittle zone and mid-transition zone has elasto-plastic behavior and the mantle rocks follow the viscous behavior. The resultant total strain rate is expressed as the sum of elastic $\dot{\epsilon}^{el}$, plastic $\dot{\epsilon}^{pl}$ and viscous $\dot{\epsilon}^{vis}$ strain rates [*Kameyama et al., 1999*].

$$\dot{\epsilon} = \dot{\epsilon}^{el} + \dot{\epsilon}^{pl} + \dot{\epsilon}^{vis} \quad (2.10)$$

Here, $\dot{\epsilon}^{el} = \frac{1}{\mu} \frac{\partial \sigma}{\partial t}$; where μ is shear modulus that represents the linear rheology (elastic deformation) in which the rate of deformation (or strain rate) linearly varies with

the dynamic stress (σ). In the elastic (brittle) domain, intercrystalline deformation kinematics (e.g. grain boundary sliding and fracturing) govern strain accumulation. $\dot{\epsilon}^{pl}$ and $\dot{\epsilon}^{vis}$ terms denote the visco-plastic deformation in the transition zone (mid-lower crust) and the upper-mantle respectively. $\dot{\epsilon}^{pl}$ indicates a low-temperature stress-controlled plastic deformation process that occurs via the Peierls mechanism as given in the equation 2.11 [Kameyama *et al.*, 1999]. Peierls mechanism is attributed to a low-temperature ($\leq 0.3T_m$) dislocation glide (section 2.5.1) within the crystal lattice, which causes permanent strain in the lattice. Compared to power-law creep (below), Peierls mechanism is less dependent on the temperature, but greatly controlled by the differential stress (~ 500 MPa).

$$\dot{\epsilon}^{pl} = A \exp[-H/RT(1 - \frac{\sigma}{\sigma_p})^q] \quad (2.11)$$

where H is activation enthalpy of the glide motion, $\mathbf{A}$ and $\mathbf{q}$ are material constants, R denotes the gas constant, and T is temperature. σ_p represents the Peierls stress (i.e. the minimum differential stress that is needed to activate dislocation glide at very low temperatures). This low-temperature plastic deformation activates at large differential stresses once the stress reaches the yield stress level.

$\dot{\epsilon}^{vis}$ is a product of high-temperature plastic deformation (viscous creep) that is usually produced via two creep mechanisms: diffusion creep and dislocation/power-law creep (equation 2.12), under upper-mantle conditions. Thus, $\dot{\epsilon}^{vis} = \dot{\epsilon}^{dis} + \dot{\epsilon}^{dif}$, where $\dot{\epsilon}^{dis}$ and $\dot{\epsilon}^{dif}$ are strain rates due to dislocation and diffusion creep respectively. The activation of different types of deformation kinematics primarily depends on the mineralogy and grain size of the rock fabric, the geothermal gradient, state of stress, and finite strain [Kirby and Kronenberg, 1987]. Figure 2.3 [Kameyama *et al.*, 1999] explains which creep mechanism becomes more favorable in olivine grains (constant grain size of 0.1 mm) under different temperature, stress, and strain conditions.

Figure 2.3 Deformation kinematic diagram as function of stress, temperature, and strain (after *Kameyama et al. [1999]*) when olivine grain size is equal to 0.1 mm. Bold lines indicate constant strain rates (s^{-1}) in log scale.

2.5.1 Dislocation creep and diffusion creep

The two forms of viscous deformation kinematics significantly alter the mantle fabrics, reflecting the most dominant kinematic under existing physical conditions. LPO anisotropy is such a deformed fabric that results from the dislocation creep of intrinsically anisotropic minerals such as olivine (section 2.4.2) in the upper-mantle [*Karato and Wu, 1993*]. When minerals in rocks are subjected to dislocation creep, they will produce a preferred orientation through the mechanisms of 1) intracrystalline slip, 2) recrystallization, and 3) grain boundary sliding [*Kirby and Kronenberg, 1987*]. Dislocation mechanisms activate due to the stress fields around the crystal dislocations. Dislocations can move through the crystal lattice along the favorable slip plane towards the crystal boundaries, resulting in a strain localization along that slip plane. Because of the intrinsic anisotropy of some minerals (e.g. olivine, orthopyroxene), crystals may have a favorable and active slip plane along where dislocations can migrate easily. This process is known as dislocation glide (edge dislocation). For example, the $[100](010)$ slip system in the orthorhombic

Figure 2.4 Velocity anisotropy of olivine crystal. Velocities of S and P waves along different propagation directions within the crystal are marked by double headed arrows and dotted lines, respectively. Note that the highest P-wave velocity is along the direction of the a-axis [100]. Typical active slip system is $[100](010)$ and this system aligns with the flow direction, shearing (or maximum extension). In this example, flow direction could be in either direction as indicated by two arrow heads parallel to [100] axis. Note b-axis and c-axis are marked by [010] and [001] respectively.

olivine crystal (Figure 2.4) is the most active slip system at high-temperature and at low differential stresses. Impurities and stress around the other dislocations in the lattice sometimes hinder the progression of dislocation glide towards the boundary. Consequently, the dislocation climb mechanism (screw dislocation) is activated by crossing the slip planes. Dislocation creep is defined as a combination of both dislocation glide and climb mechanisms. For naturally deformed olivine crystals, the viscous strain rate ($\dot{\epsilon}_s$) resulting from either dislocation creep or diffusion creep (grain-size dependent creep) can be expressed in the form of power law rheology that explains the non-linear viscous deformation (equation 2.12) [Karato and Wu,

1993]:

$$\dot{\epsilon}_{vis} = A(\sigma/\mu)^n(d/b)^m \exp[-(E^* + PV^*)/RT] \quad \text{Karato and Wu [1993]} \quad (2.12)$$

where A is the material constant, σ is the differential stress, μ is the shear modulus, E^* and V^* are activation energy and activation volume respectively, P is the hydrostatic pressure, R is the gas constant, and T is the absolute temperature. d , b , m , and n are the extra parameters that denote the grain size, the length of the Burgers vector, the grain-size exponent and the stress exponent respectively. $m = 0$ (grain size independent) for the dislocation creep mechanism, which is promoted at high-stress levels ($\gtrsim 10$ MPa) and at high temperatures (Fig. 2.3). However, dislocation creep deformation is likely influenced by the chemical environment (water content and mineral chemistry), hydrostatic pressure and dynamic recrystallization processes [Karato and Wu, 1993]. Dynamic recrystallization is a temperature dependent mechanism that is favorable under high-strain and high-temperature conditions [Bystricky et al., 2000]. It can produce more pronounced anisotropic fabrics (~ 20 % P-wave anisotropy) than dislocation creep, rotating the fast direction into the plane of shear or the flow direction [Ismail and Mainprice, 1998, Zhang and Karato, 1995]. This controls the grain size in the dislocation creep regime due to formation of new grain boundaries and/or change in existing grain-boundaries. Grain sized under dislocation creep conditions is given by:

$$d = Kb(\sigma/\mu)^{-q} \quad \text{Karato and Wu [1993]} \quad (2.13)$$

where K and q are temperature insensitive constants. According to equation 2.13, grain size reduces as the stress level increases. If significant grain size reduction occurs during the dynamic recrystallization, especially at low temperatures, then the grain-size sensitive creep (or diffusion) [Poirier, 1990] is enhanced and the transition from dislocation creep to diffusion creep takes place (Fig. 2.3). Most

Table 2.3 Flow law parameters in equation 2.12 for olivine deformation in dislocation and diffusion creep regime under dry conditions [*Karato and Wu, 1993*]

<i>Parameters</i> [†]	Dislocation creep	Diffusion creep
A(s ⁻¹)	3.5×10 ²²	8.7×10 ¹⁵
n	3.5	1.0
m	0	~2.5
E (kJ mol ⁻¹)	540	300
V(cm ³ mol ⁻¹)	~15 -25	6

[†]Parameters A, n, m, E and V are explained in section 2.5.1

importantly, effective viscosity in the dislocation regime is a function of differential stress, whereas, in the diffusion creep regime, effective viscosity is sensitive to the grain size [*Kirby and Kronenberg, 1987*].

Materials with small grain sizes undergo diffusion creep at low stress levels and low temperatures. This deformation is caused by the diffusion of vacancies within the crystal lattice. An individual crystal strain is caused by the directional diffusion of vacancies and atoms. Diffusion rate is proportional to the applied differential stress. The vacancies usually diffuse from low to high pressure (stress) regions within the lattice, and in the opposite direction to the atomic flux. Unlike in dislocation creep, grain sizes in diffusion creep are controlled by both dynamic recrystallization and grain growth.

2.5.2 Factors controlling deformation and implications for mantle dynamics

As discussed in section 2.5.1, olivine can be subjected to either dislocation or diffusion creep, depending on the flow parameters in equation 2.12 (see table 2.3). Olivine deformation has the following characteristics [*Karato, 1998, Karato et al., 2008*]:

1. Both dislocation and diffusion creep mechanisms can be enhanced and/or altered in the presence of water. Consequently, the activation of different slip systems (e.g. changing from the typical $[100](010)$ system to $[001](010)$) and different diffusion mechanisms (e.g. wet diffusion) are possible [Karato *et al.*, 2008]. However, dislocation to diffusion transition depths are independent of the water content.
2. Both creep mechanisms are sensitive to geothermal gradients, which determines the transition between the two creep mechanisms. For example, the upper-mantle beneath mid-oceanic ridges has a high geothermal gradient and, therefore, dislocation creep dominates. But the geothermal gradient in the cold (or old) upper-mantle is lower than that in the mid-oceanic ridges. Thus, diffusion creep dominates in such a domain.

Implication for mantle dynamics

The above controlling factors will affect the creep mechanisms, which in turn control mineral orientation and hence anisotropy. Thus, implications for mantle dynamics and factors controlling mantle deformation can be gained from the study of in-situ seismic anisotropy and laboratory studies of mantle-derived rocks (e.g. Xenoliths). Typical differential stress levels in the upper-mantle ($\lesssim 10$ MPa) of tectonic domains away from the mid-oceanic ridges are not enough to activate the dislocation creep mechanisms, unless there are high-temperature conditions. Thus, in such systems, diffusion creep dominates in the shallow (~ 100 km) upper-mantle (or in the deep lithosphere) and results in a rheologically weakened upper-mantle with little anisotropy [Karato and Wu, 1993]. The higher seismic anisotropy observed in such a setting is attributed to frozen LPO anisotropy that formed when this region was relatively hot and/or experiencing very high tectonic stresses [Silver, 1996]. Due to high temperatures at the intermediate depths (~ 100 -150 km) of the upper-

mantle (i.e. shallow asthenosphere) compared to the shallow upper mantle (~ 40 -100 km), dislocation creep becomes active and reflects the present-day asthenospheric flow or shear [*Karato and Wu*, 1993]. In the deep upper-mantle below the shallow asthenosphere, again diffusion mechanisms are favorable and more rapid as a result of grain boundary sliding mechanisms that are activated at relatively high-strain conditions compared to that in the usual diffusion creep conditions. This process may result in super-plasticity with little or no anisotropy in the deep mantle [*Poirier*, 1985].

Continental environments, such as collisional zones in plate-boundaries, are more likely to be characterized by large differential tectonic stresses (≥ 100 MPa; *Karato and Wu* [1993] and Fig. 2.3). As a result, dislocation creep dominates in the shallow-mantle. Increase in stress results in instantaneous increase in strain at a given grain size, but increase in stress will also reduce the grain size. *Karato and Wu* [1993, Fig. 4] demonstrated that this grain size reduction will promote diffusion creep if this process is close to the transition boundary between dislocation and diffusion creep. In summary, whether dislocation or diffusion creep is favorable depends on the tectonic setting and the physical/chemical conditions of the setting. As concluded by many studies, activation of these mechanisms at various depths may result in mechanical decoupling between Earth layers and deformation localization [*Silver*, 2002, *Vaucher and Nicolas*, 1991, *Avouac and Tapponnier*, 1993].

2.6 Sampling anisotropy using different S-phases

Anisotropy at different depths can be investigated through the shear-wave splitting of different S phases (e.g. SKS, teleseismic S, ScS and local/regional S) (Figure 2.5). SKS splitting gives constraints on the receiver side mantle anisotropy because of its phase conversions (S-P(K)-S) at the core mantle boundary. This is the most commonly used and earliest-established method to characterize the receiver-side mantle dynamics and deformation [Silver, 1996]. Although its vertical resolution is poor, SKS splitting provides a good lateral resolution to constrain the dynamics and deformational patterns in the upper-mantle and asthenosphere. The study of local/regional S- wave splitting can resolve the vertical anisotropic structure beneath the station. Most of the SKS splitting studies assume that crustal and/or subcrustal anisotropy is negligible compared to mantle anisotropy. Several crustal anisotropic studies have noted that such an assumption is not valid in most of the active tectonic domains [Godfrey *et al.*, 2000, Barruol, 1993]. Furthermore, to fully characterize the plate-boundary processes or deformation around large scale faults, both mantle and crustal deformation must be taken into account.

In tectonic domains (e.g. subduction zones, collisional and strike-slip plate boundaries etc.), it is also important to understand how the lithosphere responds to active tectonics and geodynamic processes. These domains are likely to experience significant deformation and, hence, be characterized by anisotropic structure. For example, investigation of seismic anisotropy via splitting of local and regional events recorded at varying depths in subduction zones (subduction zone in figure 2.5) allows the isolation of anisotropy in the overriding plate, mantle wedge, subducting-slab, and sub-slab. Such an investigation with good depth resolution enables capturing the subduction zone complexities [Long and Silver, 2009, Xie, 1992, Huang *et al.*, 2011b]. Therefore, the combination of the splitting from different S-phases can be used to depict the 3-dimensional anisotropic structure of tectonic domains. However,

Figure 2.5 Different S-phases utilized in S-wave splitting to investigate the anisotropic structure at various depths. Ray paths of SKS, teleseismic S-phases and regional S-phases are marked in the black color curved paths in right-side figure. The left-side figure illustrates the possible ray paths of local S-waves in a subduction setting.

frequency dependent effects (section 3.5.2) and ray-path ambiguity can make such combinations difficult to interpret.

2.7 Characterizing tectonic regimes using S-wave splitting

Patterns of S-wave splitting directions and the strength of anisotropy are characteristics of certain tectonic domains and their prevailing physical and chemical conditions. Our study area consists of both subduction and transpressional tectonic domains, where anisotropic properties may vary distinctly as predicted by shear-wave splitting studies in similar global-tectonic settings. Let us consider how S-wave splitting can be used to determine deformation and dynamics in these systems.

2.7.1 Transpressional zones: S-wave splitting

The majority of the plate motion in the transpressional domain in New Zealand is predicted to occur parallel to the plate-boundary. As a result, transcurrent flow dominates around the Alpine fault, Marlborough fault zone, and North Island dextral fault-belt. Global studies have suggested that shear-wave splitting in such domains is governed by mantle-flow derived anisotropy at upper-mantle depths [Silver, 2002, Vauchez and Nicolas, 1991], structural anisotropy due to vertical foliations [Barruol, 1993] in the mid-lower crust, and stress-aligned cracks [Peacock *et al.*, 1988] or sheared-cracks in the shallow crust. Lithospheric shear-zones associated with the large strike-slip faults (e.g. San-Andreas fault, Alpine fault), have the sense of motion parallel to the fault. In that case, intrinsically anisotropic minerals at mid-lower crustal depths may align parallel to the shear, forming vertical or horizontal foliations [Mehl *et al.*, 2003]. Depending on the rock fabric and metamorphic conditions (e.g. high-grade metamorphism), such foliated fabrics could also be present in the upper crust (e.g. Schists). These crustal shear zones can extend into the upper-mantle and possibly down to the asthenosphere as mantle shear-zones. In the lower lithosphere, such shear zones would generate anisotropic mineral alignment that is parallel to the plate-boundary transcurrent movement. In both cases, vertically propagating S-waves can split by the fabric alignment and the faster S-wave will be polarized in the direction parallel to the transcurrent shear and the alignment of vertical foliations with high strength of anisotropy [Vauchez and Nicolas, 1991]. The strength and fast polarization direction may vary depending on the crust and mantle contributions [Barruol *et al.*, 1997]. For instance, if the crustal foliations are not well developed or anisotropy is altered by other mechanisms to produce different geometry or different types of anisotropy, then crust-mantle decoupling that may result in complex shear-wave splitting is also possible.

Collision and orogenesis in South Island, New Zealand are more recent tectonic events

that initiated with the migration of the Australian passive-margin into transcurrent plate-boundary zone (section 1.3.1). According to [Nicolas, 1993], when passive margins are subjected to transcurrent motion as a result of plate collision, two types of anisotropic fabrics may occur: 1) frozen mineral alignment from asthenospheric flow parallel to rift axis and 2) mineral alignment parallel to transcurrent motion. Although these two anisotropies are expected to have similar orientations, changes in plate-boundary orientation during the two tectonic events may alter the general trend of fabric orientations, resulting in spatially or depth varying anisotropy. Thus, deflection of plate-boundary parallel fast azimuth of shear-wave splitting can be attributed to changes in tectonic environment over the geologic timescale [Nicolas, 1993]. Such variations have also been attributed to present-day stress governed anisotropy in the shallow crust by some local shear-wave splitting studies [Peacock *et al.*, 1988]. In that case, stress-induced alignment of the cracks at shallow depths would produce vertical crack planes that are parallel to the maximum horizontal compressives direction (SH_{max}), so that the fast split S-wave from a vertical ray path will be polarized parallel to the SH_{max} .

2.7.2 Subduction zones: S-wave splitting and dynamics

Subduction zones, where the descending slab lithosphere interacts with the over-riding lithosphere, are expected to be characterized by complex mantle dynamics and lithospheric deformation. The complexities can vary from one subduction zone to another depending on the deformation conditions (e.g. pressure, temperature, strain rate and/or stress), subduction zone parameters such as convergence velocity, slab dip, trench migration velocity [Long and Silver, 2009] and the properties of subduction zone material. The study of shear-wave splitting from core phases (e.g. SK(K)S, ScS) and teleseismic S-phases and geodynamic modeling allows us to constrain the mantle flow in the subduction systems. Mantle flow governs the

development of LPO by aligning the crystallographic axes of minerals in a preferred direction, most commonly parallel to the mantle flow direction [*Savage, 1999*].

Shear-wave splitting anisotropy in subduction systems

Vertically traveling S-waves in a subduction zone can be split by anisotropy in the sub-slab region, subducting slab, mantle wedge and the overriding plate [*Long and Silver, 2009*]. These zones can be characterized by large-scale intrinsic anisotropy that is present at mantle depths and small-scale crustal anisotropy due to cracks/microcracks and melt/fluid-filled microcracks or pockets, faulting structures in the upper-plate and bending related cracks in the slab [*Morley et al., 2006, Gledhill and Stuart, 1996*]. Despite the complex nature of the local/regional S-wave splitting, the good depth resolution of those phases enables us to integrate them with the SKS splitting to isolate the anisotropic signatures in the sub-slab region, mantle wedge, slab and overlying plate [*Savage, 1999*]. If the length-scale of anisotropy and/or orientation of the alignments in those zones are significantly variable, frequency dependent splitting and/or multilayer/two-layer modeling approaches (sections 3.5.1 and 3.5.2) incorporated with splitting analysis can be used to resolve the complexities in subduction zones.

Origin and factors controlling the slab, mantle wedge and sub-slab anisotropy

In subduction zones, the mantle wedge and sub-slab regions are usually considered to be the dominant anisotropic zones. Nevertheless, recent studies have suggested that the slab itself also contributes a considerable amount to the total anisotropy in some systems [*Faccenda et al., 2008, Morley et al., 2006*]. Slab anisotropy can have a different origin of anisotropy and SKS/Tel.S wave splitting may not be capable of detecting such relatively small-scale localized anisotropic structures in the slab.

As inferred by an integrated study of local S to P conversions, teleseismic receiver function studies and Pn azimuthal anisotropy studies [Song and Kawakatsu, 2012], the slab itself may consist of interior layering and the uppermost mantle region of the slab may be characterized by a localized anisotropic zone. Song and Kawakatsu [2012] attributed such a localization of anisotropy to the anisotropy preserved in the subducting ocean lithosphere from paleo-spreading. However, Long and Silver [2008] found no correlation between such fossil spreading directions and ϕ (polarization direction of the fast S-wave) from splitting anisotropy in the global subduction settings, and they concluded that there is no clear relationship between the age of the subducting slab (or thickness of the slab) and the amount of observed δt .

In addition to anisotropy derived from paleo-spreading, anisotropy in the slab or mantle wedge can be modified if the system is hydrated. As explained in Long [2013], the shallow slab is most likely to retain water in the form of hydrous minerals (e.g. Serpentine minerals like Antigorite and Lizardite, and clay minerals like talc and chlorite). Such a mineralization is possible during the formation of oceanic crust in the oceanic environments (spreading centers) and/or when the topmost layer of newly created crust is subjected to hydrothermal alterations [Boudier *et al.*, 2010, Long, 2013] and/or hydration/serpentinization of the bending-induced faulting in the shallow layer of the subducting slab [Faccenda *et al.*, 2008, Ji *et al.*, 2013]. Compared to Antigorite, Talc is more stable under high temperature conditions and can be present from the sea-floor to 150 km in the subduction zone. Talc can be derived from the unstable Antigorite under high temperature/pressure conditions especially in the mantle wedge [Mainprice and Ildefonse, 2009]. These hydrous minerals are assumed to produce strongly anisotropic media because of their relatively high single crystal anisotropy (e.g. SWA of antigorite and talc $\simeq 68\%$ and Lizardite $\sim 81\%$) [Long, 2013].

The majority of studies agree with trench parallel flow below and above the slab (close to trench) as predicted by the slab-entrained flow (section 2.7.2) . However, a

few other studies have noticed trench perpendicular/oblique ϕ in the back-arc region (mantle wedge) [Morley *et al.*, 2006]. Fischer *et al.* [2000] proposed the 2D-corner flow model to explain the trench perpendicular splitting observed in the back-arc regions (section 2.7.2). Splitting estimations from the local shear-wave splitting and some SKS-splitting studies have determined that anisotropy in the wedge is more complex than anywhere else in the subduction zone [Wirth and Long, 2010, Abt and Fischer, 2008]. Several studies argue that the complex anisotropy in the mantle-wedge above the slab can also be due to B-type olivine (010)[001] as result of the high H_2O content under low temperature and high stress conditions [Karato *et al.*, 2008], or high pressure conditions (deeper than ~ 90 km, or pressure around 3 GPa) [Jung *et al.*, 2009, Long, 2013]. Serpentinization or mineralization of clay minerals can occur in the mantle wedge as a result of the water-rich sediments or dehydration of the slab-mineral. In that case, the wedge may be characterized by highly anisotropic minerals other than olivine (e.g. antigorite, talc) with variable anisotropic symmetry properties. Mehl *et al.* [2003] demonstrated that presence of high H_2O content could result in the activation of (001)[100] slip in C-type olivine, producing horizontal foliations rather than typical vertical foliations that are caused by the (010)[100] slip. In both cases, the S-wave splitting will yield trench parallel ϕ , but δt resulting from the horizontal foliations can be higher than δt resulting from the vertical foliations. Holtzman *et al.* [2003] ascribed a complex splitting in the wedge to the effect of melt on LPO of olivine. Deviatoric stress can create an anisotropic alignment of the melt pockets and this alignment may cause girdles in a- and c-axes of olivine in the absence of melt segregation. However, if the melt is segregated into melt rich shear zones, then the olivine a-axis can be rotated 90° from the shear plane, resulting in trench perpendicular or oblique ϕ [Holtzman and Kendall, 2010].

Sub-slab region anisotropy can be approximately estimated by either a combined study of the splitting of SKS and local S-phases that originate in the slab, isolating

the splitting in the mantle below the slab [Long and Silver, 2009] from the mantle-wedge/slab anisotropy or by source-side splitting of down-going teleseismic S and correction for receiver-side splitting [Russo, 2012, Lynner and Long, 2013]. Long and Silver [2009] showed that most of the subduction systems are characterized by trench-parallel flow beneath the slab, and that the observed splitting delays agree with the anisotropy due to flow in the upper-mantle and/or in asthenosphere. Song and Kawakatsu [2012] proposed a different hypothesis to characterize the observed splitting in sub-slab region. The proposed idea is that the sub-slab oceanic asthenosphere is coupled with the subducting slab. As a result of this slab-coupled subducting asthenosphere, the sub-slab region can be characterized by orthorhombic symmetry as predicted in the oceanic asthenosphere, but the symmetry axis is tilted because of the subduction. Inferred variations of splitting parameters from this method tend to agree well with splitting observations in shallow (dip $<15^\circ$) subduction zones [see Song and Kawakatsu, 2012, Fig. 2]. However, in steep subduction settings the predicted ϕ with this model are parallel or subparallel to the trench as predicted by the trench parallel flow model (next section and Fig. 2.6) and, therefore, it is hard to discriminate between the two models.

Subduction zone flow models

If the anisotropy in subduction settings is mainly governed by mantle flow, the ϕ derived from the S-wave splitting must be either parallel to the relative plate motion direction (RPM) and/or to the absolute plate motion direction (APM). Deviation from these simple trends implies that there are other mechanisms that could be affecting the anisotropy or flow pattern in subduction settings. Based on the splitting parameters from various subduction settings, several flow models have been proposed to describe the flow geometries in those settings. Some models are focused on the flow patterns in certain regions (e.g. subslab, mantle-wedge, back-arc). Russo and Silver [1994] proposed the first subduction zone flow model to

explain the trench parallel splitting from the SKS phases. This model consists of trench-parallel flow just beneath the slab that results from the retrograde motion of the slab relative to mantle flow. This model is comparable to entrained flow around the slab that is proposed by other recent studies (Fig. 2.6). The trench parallel flow is, however, assumed to decouple from the upper-mantle asthenospheric flow, which represents the APM direction at depths or away from the subduction setting.

Trench parallel ϕ above the slab, which is determined from large δt ($\gtrsim 1$ s) of local S-wave splitting [Smith *et al.*, 2001], is assumed to result from the along-arc mantle flow above the slab. Such a flow pattern can be generated from the slab rollback effect and/or a trench parallel component of the APM/RPM direction in oblique subduction domains. In addition to that, if the overriding plate consists of large scale strike-slip faulting parallel to the trench due to RPM in the forearc region (e.g. southern Hikurangi subduction zone), then mantle shear zones beneath those faults could also produce splitting that is parallel to the faulting and the trench. The two-dimensional back-arc flow (also known as 2-D corner flow) model explains trench perpendicular ϕ with strong splitting in the back-arc regions [Fischer *et al.*, 2000]. Furthermore, models with olivine transitioning from A- to B-type in the mantle wedge, models with horizontal and vertical foliations that result from trench parallel flow in the transpressional subduction settings [Mehl *et al.*, 2003] and the recent model of slab-coupled asthenospheric flow by Song and Kawakatsu [2012] have been proposed to explain the variable splitting parameters in subduction zones by recent studies (Long [2013]; see Fig. 2.6).

Because of the complexities in the different subduction settings it is impossible to provide a distinct flow model that generalizes the flow patterns in various subduction regimes. Long and Silver [2008] utilized a combined flow model (Fig. 2.6-C) that is characterized by 2D corner flow in the back-arc region and 3D flow around the slab (similar to trench parallel slab-entrained flow) to test whether there is a relationship between splitting parameters from various subduction zones and subduction zone

Figure 2.6 Schematics display the main flow models proposed for subduction zones. (a,b) Trench parallel flow in the sub-slab and above the slab [Russo and Silver, 1994, Smith *et al.*, 2001]. (c) Combined flow model Long and Silver [2008] (2-D corner flow model [Fischer *et al.*, 2000] and 3-D slab entrained flow). (d) Slab-coupled asthenospheric flow [Song and Kawakatsu, 2012].

parameters (see 2.7.2). 2D corner flow model is assumed to derive from the down dip motion of the slab, which possibly controls the trench perpendicular ϕ in the back-arc region as noticed by many studies. Conversely, 3D flow or slab entrained flow can be approximated by trench parallel flow below and above the slab and is mainly governed by the trench migration velocity relative to the hot spot reference frame. The trench parallel flow in the wedge, however, can be altered by the trench perpendicular or oblique ϕ in the presence of B-type olivine fabrics. Long and Silver [2009] concluded that trench parallel flow in the sub-slab is mainly governed by the trench migration velocity, and trench parallel flow (relative to surrounding mantle) of the fast-moving trenches is more likely to produce stronger anisotropy than that of the slow or nearly stationary trenches. As estimated by Gledhill and Gubbins [1996], sub-slab splitting anisotropy of the subduction zone in New Zealand (~ 1.5 s) is also higher than most of the global subduction zones and, therefore, is expected to be characterized by a significant trench parallel flow.

Chapter 3

Data and methodology

This chapter discusses the instruments, quality of the S-phases and the main methods that have been utilized in this study. In some sections theories related to methodology are discussed in detail. Hereafter South Island and North Island are denoted by the acronyms of SI and NI, respectively.

3.1 Experiment and Data

The thesis consists of three main research projects (1) on-land SI, 2) offshore SI, and 3) southern NI). All three projects were focused on investigating S-wave splitting anisotropy. The on-land and offshore SI experiments utilized local and regional high frequency S phases to explore S-wave splitting parameters in the plate-boundary zone in SI, New Zealand, whereas the southern NI experiment utilized low frequency SKS, ScS and teleseismic S-phases to examine the subduction zone anisotropy. In all three studies, seismic records from broad-band seismographs are used to determine S-wave splitting. Local/regional S arrivals recorded on Ocean Bottom Seismographs (OBS) are used to investigate S-wave splitting anisotropy in the offshore, SI.

S-wave signals and quality of the seismograms

Shear-wave splitting analysis window must be free from noise and a clear S wave onset is preferred for reliable splitting measurements. To minimize the distortion of S-waves at layer boundaries and/or at the free surface, S-wave splitting analysis is usually carried out using the S-waves that are within the incidence angle (i_c) of 35° degrees (See section 3.3) [Nuttli, 1961]. S-waves are also filtered to enhance the S-arrival and to increase the Signal-to-Noise ratio (SNR) by removing unnecessary

noise. Thus, all three studies used suitable band pass filters, which were two-way 4-pole Butterworth filters. The applied filter frequency bands were dependent on the different types S-phases that have been used to examine splitting (see sections 3.1.1, 3.1.2, and 3.1.3). Figure 3.1 shows the comparison between unfiltered (top two) and filtered (bottom two) S phases (local S and SKS). The filter bandwidth was selected so that the filtering only enhanced the S-wave onset without distorting the S-phase.

Figure 3.1 The comparison between filtered and unfiltered S phases. A1 and A2 are the unfiltered and filtered (band pass between 0.2 and 3 Hz) high frequency local S phases. B1 and B2 shows the SKS phases unfiltered and band pass filtered between 0.04 and 0.1 Hz. Note the enhanced S-wave onset in the filtered records.

3.1.1 On-land SI: anisotropy of the on-land plate-boundary zone

Focus: We Investigate shear-wave splitting in SI New Zealand to understand the spatial and depth distribution of S-wave splitting anisotropy and to consider implications for the distribution of deformation in the plate-boundary zone.

Experiment

Local S-phases that were recorded on both permanent and temporary broadband stations in SI are used to investigate seismic anisotropy in the lithosphere using the shear-wave splitting technique. We used thirty 3-component broad band stations. Twenty six of them were GeoNet permanent stations and the other four stations were temporary PASSCAL seismographs that were deployed by Victoria University of Wellington in collaboration with the University of Colorado to enhance the seismic coverage. Both sets of instruments consisted of Guralp-3ESP and Kinematics FBA-ES broad-band seismographs (Fig. 3.2). The PASSCAL utility package and Antelope software are used to convert RT130 data into SAC format (appendix D).

Data selection

We used local and regional S-waves recorded for the period of 2004-2010 on GeoNet network stations and events recorded for an eleven month period (2009 February - 2010 January) on the temporary stations in the northern SI (Fig. 3.2 and squares in Fig. 3.3).

Figure 3.2 Temporary deployments (2009-10) and MONA OBS recovery in 2010.

Figure 3.3 (a) and (b) maps showing the distribution of earthquakes that yield good quality splitting measurements in SI. The earthquakes are color coded according to earthquake depth. (a) displays events recorded from all possible incidence angles and (b) shows the events that are within the i_c of 35° (see table 3.1). Earthquakes within three blocks (northern SI, central SI and southern SI) are analyzed only using the station within that block, but the stations at the block boundaries (MSZ, WKZ, ODZ, CASS and LTZ) used the events recorded on both sides. Squares and triangles display the temporary stations and permanent stations (GeoNet network) respectively.

Table 3.1 Event selection criteria and event/measurement information.

Block	min. magnitude	tot.events	max.event depth (km)	no. meas.(AB*)	AB*/ $i_c \leq 35^0$
ALL	-	2851	244.00	8336	3183
Northern SI	3.5	1070	243.85	1925	1167
Central SI	3.0	470	55.5**	1898	189
Southern SI	3.5	769	177.42	2910	1213

*AB grade measurements (section 3.2.3)

**depth indicates event depths of the majority of events in the central SI, but a few subduction zone events that were recorded on the central SI stations had the maximum depth of ~ 90 km.

The complete region was divided into three geographic units: northern, central and southern SI and events in each block were collected from the stations in that particular block during the initial retrieval of data (Fig. 3.3b). Stations that are close to the borders (WKZ, MSZ, ODZ, CASS and LTZ) of a block used the events recorded from either side. This criterion is used to control the epicentral distances and to avoid analyzing unnecessary waveforms at each station.

The number of events that give acceptable quality splitting measurements was further reduced by limiting events that have incidence angles less than 35^0 (Fig. 3.3-b, section 3.3). To obtain good signal-to-noise ratio, we used a threshold magnitude of 3.5 in northern and southern SI. Due to the lack of seismicity and deep events in central SI, the threshold magnitude of 3.0 was chosen to obtain more events (table 3.1).

In addition to the above analysis, splitting parameters from smaller earthquakes ($2.0 \leq \text{magnitude} < 3.0$) are combined with the above measurements to investigate the in-situ seismic anisotropy of the Haast schist belt in central-southern SI. For this analysis, earthquakes that are within 60 km epicentral distances and depths down to 40 km are extracted from 10 stations (WKZ, EAZ, TUZ, JCZ, FOZ, WVZ, LBZ, ODZ, MLZ, and OPZ) in or near by regions of the schist (Fig. 3.4).

Figure 3.4 Events (mag. ≥ 2.0) used to estimate splitting anisotropy of the schist rock in Central SI. Pink to purple colors indicate the schist rock of low to high grade metamorphism.

3.1.2 Offshore SI: Local and regional shear-wave splitting in the offshore SI

Focus: We examine the anisotropic structure in the offshore SI to study deformation distribution around the plate-boundary zone. This is an extension of the on-land study.

Experiment

Local and regional S phases recorded on the MOANA (Marine Observation of Anisotropy Near Aotearoa) Ocean Bottom Seismographic (OBS) network (Fig 3.5) are used to determine splitting parameters in the eastern and western offshore regions

of the SI [Collins *et al.*, 2010]. The initial deployment consisted of 30 OBS and data were collected from January 2009 to February 2010. OBS instruments used for the experiment belong to U.S. Ocean Bottom Seismic Instrument Pool. Twenty-nine OBS seismographs were Nanometric Trillium 240 and one was a Nanometric Trillium 40 seismograph. They were set up to record continuous wave-form data with a 50 sps sampling rate [Collins *et al.*, 2010, Yang *et al.*, 2012]. The seismometer orientations of the OBS on the seafloor were estimated from the Rayleigh wave arrival azimuths that are determined from the polarization analysis of the Rayleigh waves [Stachnik *et al.*, 2012]. All the OBS seismograms have been corrected for sensor orientations that were determined by Stachnik *et al.* [2012] using SAC software prior to processing data for splitting. Except for NZ01 and NZ17, all the other stations recorded good quality waveforms that can be used for observing splitting.

Figure 3.5 MOANA OBS network (white triangles). Circles indicate the events with AB quality splitting measurements and they are colored according to event depth.

Data selection

Seismicity in the area is mostly distributed along the plate-boundary outside of the OBS network, roughly centering between the western and the eastern OBS deployments (Fig. 3.5). We used ~ 2000 S-phases from events with magnitude (M) ≥ 3 in western SI and $M \geq 3.5$ in the eastern South Island. The event information in the GeoNet catalog from February 2009 - January 2010 was used to identify the events in OBS records. We have eliminated poor quality events and/or measurements when picking the S-phases. Because of the noisy nature of the OBS seismograms, S-waves were subjected to filtering using a range of two-pole Butterworth bandpass filters. In general, the majority of wave-forms consisted of high frequency (0.5 - 10 Hz) S-phases.

3.1.3 Southern NI: Deep anisotropic structure beneath the Hikurangi subduction zone

Focus: This study aims to expand the understanding of the deep anisotropic structure beneath the southern Hikurangi subduction zone using teleseismic S-phases recorded on the SAHKE broadband-transect across the southeastern NI.

Experiment

The main aim of the Seismic Array Hikurangi Experiment (**SAHKE**) is to investigate the tectonics and subduction interface processes at the southernmost part of the Hikurangi subduction margin. The project had both onshore (ARRAY, TRANSECT, WANGANUI, and NORTHERN lines) and offshore (OBS-TRANSECT) deployments and aimed to record local, regional and global passive seismicity and regional on-land and offshore active source signals [*Seward et al.*, 2010,

Henry et al., 2013]. For our study, we only use the teleseismic (Tel.) and regional passive seismic data recorded on the broadband transect (SAHKE transect) that extends from the center to the east-coast of the southern NI (Fig. 3.6). Shear-wave splitting is used to determine the anisotropy in this region. Tel. S, direct S, SKS, and ScS splitting was not well explored in the southeast of North Island by previous studies. Therefore, this study enables us to obtain a complementary set of splitting data to expand our knowledge on the anisotropic structure in Hikurangi subduction zone. Combining our measurements with the existing splitting measurements for the western region of the lower NI, we examine splitting anisotropy variations across the Hikurangi subduction zone.

The northwest-southeast oriented broadband transect consisted of 10 3-component seismographs. Most of these sites had CMG40T sensors, but two sites had CMG3ESP sensors (Fig. 3.6 and appendix C.1). Either Reftek-130s or Earth Data PR6-24 data loggers were used at each site, therefore both the PASSCAL utility package and integrated methods were used to convert RT130/mseed data to SAC format (appendix D). The broadband stations were installed between November 2009 and December 2010. Stations had variable and non-continuous recording periods ($\sim$ 6 -18 months). As a reference set of data, we have also investigated splitting from the teleseismic S-waves recorded between 2009 November and 2011 November at the permanent GeoNet broadband station (WEL), which is located in the Wellington region.

Data selection

Global events with magnitudes ≥ 6 that were obtained from the IRIS earthquake catalog have been used to extract SK(K)S, ScS and Tel. direct S arrivals from the seismograms. We used previously established event selection and filtering criteria from *Marson-Pidgeon and Savage* [1997] (table 3.2). These criteria have been

Figure 3.6 Locations of the SAHKE transect and WEL-GeoNET permanent station (white triangles). White squares display the locations (Taratua array and SNZO permanent station) of previous teleseismic-splitting studies in the southern NI. Black lines and dotted lines display the active faults and bathymetry respectively.

applied to examine the S-wave splitting anisotropy in Wellington region before and such selection criteria allow us to extract unambiguous phases and nearly vertical incident waves with less noise contamination than the raw records. The event selection, filtering criteria, and measurement information are given in the table 3.2.

The expected phase arrival times are identified using the Taup toolkit [Crotwell *et al.*, 1999] and the actual arrivals are picked manually. Figure 3.7 displays ray paths from teleseismic events that produce good quality splitting parameters in the

Table 3.2 Event selection criteria for SKS phases and measurement information.

phase	searched depths* (km)	distance(°)	filter (Hz)	no of measurements	min.depth (km)
SKS	all depths	85 - 125	0.04 - 0.1	48	-
ScS	>390	≤ 35	0.1 - 0.3	21	>400 km
Tel. S	>200	60 - 80	0.04 - 0.1	33	>450 km

*Search depth range for source side

analysis. We used the transverse component minimization technique [Silver and Chan, 1991] to determine splitting parameters from SKS phases (see section 3.2.2). Splitting parameters of ScS and Tel. S phases are determined from the eigenvalue minimization technique (section 3.2.3).

Figure 3.7 Ray paths distribution of SKS (red), ScS (blue), teletismic S (yellow) around the SAHKE transect in Fig. 3.6

3.2 Methodology

Shear-wave splitting is used to investigate seismic anisotropy in all three case studies. Different types of S-phases (local/regional S, teleseismic S, ScS and SKS) have been utilized to investigate anisotropy and, therefore, different S-wave splitting techniques (e.g. Transverse and eigenvalue minimization) are implemented to measure splitting parameters.

3.2.1 Summary of different S-wave splitting analysis methods and their applications

Seismic shear-wave measurements contain a range of frequencies and therefore S-wave splitting can be sensitive to the structures of different scales-lengths [Wu, 1982, Long and Silver, 2009]. Both high-frequency and broad-band seismograms (Fig. 3.1) can be used to estimate shear-wave splitting parameters, but the splitting of high-frequency phases are more sensitive to the upper-most crustal anisotropy [Rumpker and Silver, 1998]. If the split S-phases exhibit a clear separation in the seismogram, splitting parameters can be easily estimated using a simple particle motion analysis [Vidale, 1986]. This is possible with most of the high-frequency S-phases, however such a simple approach is not suitable to determine splitting parameters from noisy data and/or long-period (period of S-phase $\gg \delta t$) S-phases such as SKS (period > 10 s) [Vecsey et al., 2008, Long and Silver, 2009]. The various techniques of S-wave splitting are available to study splitting parameters depending on the different types of S-phases used in the experiment (e.g. SKS, local/regional, teleseismic S and ScS etc.). Here we discuss some of the commonly used methods. Eigenvalue minimization and transverse component minimization techniques are discussed in detail in section 3.2.2 as the work presented here is primarily based on those two techniques.

1. Transverse component and eigenvalue minimization methods

The transverse component and eigenvalue minimization methods are two of the commonly used techniques to determine shear-wave splitting parameters. The transverse component minimization technique is commonly used with SKS splitting analysis, whereas eigenvalue minimization is most often utilized in local and regional splitting analysis and is also applicable to teleseismic splitting. Commonly used splitting techniques, such as SplitLab and Mfast, are based on the transverse component and eigenvalue minimization methods. Although these two methods are first introduced in some early literature (e.g., *Silver and Chan* [1988]), detailed information on the methodology is given in *Silver and Chan* [1991]. Both methods use grid searches over splitting parameters (δt and ϕ) to find the pair that best minimize the energy on the transverse component or the pair that best linearize the corrected particle motion [*Long and Silver*, 2009]. The theoretical explanation of these two methods is given in section 3.2.2.

2. Cross-correlation method

This method assumes that the two split S-phases in anisotropic media exhibit similar shape waveforms. As explained by *Bowman and Ando* [1987], the method involves usage of particle motion plots of horizontal components, the rotation of horizontal components and calculation of the cross correlation coefficient between those two components for time lags with constant intervals (usually ranging from +1 to -1 s). The angle and time lag that gives the maximum correlation coefficients determine the polarization direction of the fast S-wave and delay time between two split phases [*Fukao*, 1984]. Similar to the eigenvalue minimization method, this method utilizes horizontal particle motion plots, which are created in the geographic coordinate system to check the linearity of particle motion when the correlation coefficient is maximized.

3. Splitting intensity method

Chevrot [2000] developed this technique to analyze splitting parameters from multi-events and this method is complementary to the transverse minimization technique [*Silver and Chan*, 1988, 1991], which applies to single records. The method defines a splitting intensity ($\mathbf{I}$), which is a measure of amplitude of the transverse component relative to the derivative of the radial component. $\mathbf{I}$ can be derived from the relationship given in equation 3.10 [*Vinnik et al.*, 1989b] for an individual record, defining splitting intensity as: $\mathbf{I} = \delta t \sin 2\beta$, where β is the angle between ϕ and initial polarization direction (ϕ_p). $\mathbf{I}$ is then determined from the singular value decomposition (SVD) procedure discussed in *Chevrot* [2000]. It requires a wide range of ϕ_p (BAZ in the case of SKS splitting investigations) to determine splitting parameters accurately. $\mathbf{I}$ is plotted with respect to ϕ_p (also known as the splitting vector; *Long and Silver* [2009]) to find the best fitting $\sin 2\beta$ curve and to determine δt and ϕ . The amplitude of this sinusoidal curve is equivalent to δt and ϕ can be determined from the phase at the origin. The main advantage of the splitting intensity method is that unlike the splitting operator [*Silver and Chan*, 1991], it is commutative and thus it follows the main assumption of single homogeneous anisotropy approximation. As shown by *Silver and Long* [2011], splitting intensity, therefore, will not provide information on the apparent splitting parameters (or the complex anisotropy) as a splitting operator does (section 3.2.2). This method is ideal for tomography inversions as it assumes the homogeneous single layer anisotropy [*Silver and Long*, 2011], but results could be ambiguous because the splitting intensity method cannot distinguish in which order layers appear.

3.2.2 Transverse component and eigenvalue minimization: theoretical overview

Eigenvalue minimization and transverse component minimization methods by [*Silver and Chan*, 1991] (here after SC91) assume that the shear-wave splits into two orthogonally polarized (radial and transverse or fast and slow) quasi S-phases as it traverses through an anisotropic medium and therefore, exhibits elliptical particle motion, indicating energy on both horizontal components.

Eigenvalue minimization method:

This method uses the same approach implemented in *Vidale* [1986], which analyzes the polarization directions of three component seismographs using a covariance matrix of the analytic signals and calculates eigenvectors associated with three eigenvalues that corresponds to the particle motion directions of each component. But in the splitting analysis, only two orthogonally polarized horizontal-components are considered.

If a shear wave ($\tilde{u}(\omega)$) propagated through length L in an isotropic, homogeneous medium and reaches the surface at T_0 time, its propagation can be represented mathematically as (SC91):

$$\tilde{u}(\omega) = \mathbf{A}\Psi(\omega)\exp[-i\omega T_0] \quad (3.1)$$

where, $\mathbf{A}$ and $\Psi(\omega)$ are the complex vector amplitude and the wavelet function, which is a function of frequency(ω), respectively. $\mathbf{A}$ can be approximated by complex scalar (A) and real unit vector ($\hat{p}$) that indicates the polarization direction of the S-wave (i.e. $\hat{p}$ is perpendicular to the wave propagation direction). By replacing $\mathbf{A}$ with $A\hat{p}$, equation 3.1 can be rearranged as:

$$\tilde{u}(\omega) = \Psi(\omega) \exp[-i\omega T_0] \hat{\mathbf{p}} \quad (3.2)$$

Then, the covariance matrix ($c_{ij}(\phi, \delta t)$) between the two split S-waves ($u_i(t)$ and $u_j(t - \delta t)$) in an anisotropic medium, arriving at the surface with the time lag (δt) is given by:

$$c_{ij}(\phi, \delta t) = \int_{-\infty}^{\infty} u_i(t) u_j(t - \delta t) dt \quad (3.3)$$

Here, ϕ defines the fast polarization direction with respect to $\hat{\mathbf{p}}$. SC91 method defines an inverse splitting operator (Γ^{-1}) which reverses the split phases into unsplit stage. Splitting operator Γ denotes the splitting of S-wave into fast and slow waves with time lag of δt and mutually perpendicular polarization directions of $\hat{\mathbf{f}}$ and $\hat{\mathbf{s}}$. The mathematical explanation of the Γ as given in SC91:

$$\Gamma_{(\phi, \delta t)} \equiv \exp[i\omega \delta t / 2] \hat{\mathbf{f}} \hat{\mathbf{f}} + \exp[-i\omega \delta t / 2] \hat{\mathbf{s}} \hat{\mathbf{s}} \quad (3.4)$$

Equation 3.2 can, therefore, be redefined incorporating Γ to denote a split shear wave ($u_s(\omega)$) as:

$$u_s(\omega) = \Psi(\omega) \exp[-i\omega T_0] \Gamma_{(\phi, \delta t)} \hat{\mathbf{p}} \quad (3.5)$$

To compact the splitting operator, a time difference tensor is defined as [*Silver and Chan, 1991, Walsh, 2012*]: $\delta T = \delta t / 2 (\hat{\mathbf{f}} \hat{\mathbf{f}} - \hat{\mathbf{s}} \hat{\mathbf{s}})$ and, thus, the equation 3.4 can be re-written as:

$$\Gamma_{(\phi, \delta t)} \equiv \exp[i\omega \delta T_{(\phi, \delta t)}] \quad (3.6)$$

Correction for splitting is done by searching for the inverse splitting operator that successfully reverses splitting to its unsplit stage ($\tilde{u}(\omega)$):

$$\tilde{u}(\omega) = \Gamma^{-1} * u_s(\omega) \quad (3.7)$$

In the absence of anisotropy or if the the seismogram is fully corrected for splitting, $c_{ij}(\phi, \delta t)$ (equation 3.3) will only have one eigenvalue and its eigenvector is equivalent to the initial polarization direction ($\hat{p} = \phi_p$). $c_{ij}(\phi, \delta t)$ will have two eigenvalues, if there are two split shear-waves and the eigenvector associated with the smaller eigenvalue (λ_2) is equal to the ϕ . The best Γ^{-1} is determined from the eigenvalues of the $c_{ij}(\phi, \delta t)$ that are calculated from the rotated and shifted seismograms over every possible pair of ϕ and δt using the grid search approach (Fig. 3.8-3).

Because of the noise associated with seismic records, this method looks for the pair (λ_1 and λ_2) with the smallest $\lambda_{2(min)}$ (or the most nearly singular matrix) to define the best splitting parameters ($\phi, \delta t$) and the inverse splitting operator that best resolves for splitting (Fig. 3.8). Unlike other methods, this method can be used to estimate splitting parameters even if the initial polarization direction (ϕ_p) is not known and it is commonly used to examine splitting parameters from local and regional shear-waves. The minimization of λ_2 is equivalent to maximizing the λ_1/λ_2 as in the aspect ratio method [Shih *et al.*, 1989]. Maximizing correlation coefficients of the cross-correlated S-waves as in Bowman and Ando [1987] is also equivalent to maximizing the determinant of the $c_{ij}(\phi, \delta t)$ [Silver and Chan, 1991]. The corrected S-wave is rotated into its initial polarization direction to achieve the original state of the S-wave. Unlike in SKS phases, ϕ_p direction of the local and regional S-waves is not known. The eigenvector of the larger eigenvalue, when the other one is minimized, gives the ϕ_p .

Figure 3.8 The main steps of measuring S-wave splitting parameters using Mfast [Savage et al., 2010]: **1)** Picking the S-arrival and filtering the three-component seismogram (e.n, and z) **2)** Measuring splitting parameters by rotating (p and $p_{\perp}$), shifting (corrected p and $p_{\perp}$) seismograms and calculating eigenvalues of the covariance matrix, which is calculated from the corrected components. Here, dashed lines indicate the minimum starting time and maximum end time of the different S-wave analysis windows **3)** Repeat step 2 applying a grid search over ϕ and δt to find the smallest second eigenvalue ($\lambda_{2(min)}$). $\lambda_{2(min)}$ is denoted by the blue cross, which is placed within the 95% confidence limit (dark black line) in the contour plots (diagnostic plot iii) of second eigenvalues (or minimum energy). The diagnostic plots of wave form and particle motion plots show the correction for splitting using the best splitting parameters at $\lambda_{2(min)}$. **4)** Repeat step 2 and step 3 for different shear-wave analysis windows that are defined between minimum (W_{st})/maximum (W_{s2}) starting time and minimum (W_{e1})/maximum (W_{e1}) end time (step 2) and find the most stable splitting parameters from the cluster analysis [Teanby et al., 2004b].

Transverse component minimization method:

Transverse component minimization is a frequently used approach to determine splitting parameters from core phases, such as SKS and SKKS. As a result of the P-to-S conversion at the core-mantle boundary (CMB), when SK(K)S phases propagate through the receiver side CMB, the phases exhibit no splitting or energy on the transverse component that remains from the source side anisotropy. Thus, in the case of SK(K)S splitting ϕ_p is already known and it is equivalent to the direction of radial polarization.

If a radially polarized S-wave traverses through an anisotropic medium, S-wave splitting results in energy on the both radial (u_r) and transverse (u_t) components. As describe in [Silver and Chan, 1988], the horizontal projection of these two components on the RTZ coordinate system can be expressed as a functions of the radial wave in the unsplit stage ($\tilde{u}_r$), δt and ϕ as:

$$u_r(t) = \tilde{u}_r(t) \cos^2 \phi + \tilde{u}_r(t - \delta t) \cos^2 \phi \quad (3.8)$$

$$u_t(t) = -\frac{1}{2} \sin 2\phi [\tilde{u}_r(t) - \tilde{u}_r(t - \delta t)] \quad (3.9)$$

As derived by Vinnik *et al.* [1989a], time derivative of $u_r(t)$ ($=\dot{u}_r(t)$) is a function of $u_t(t)$.

$$u_t(t) \simeq -\frac{1}{2} \delta t \sin(2\phi) \dot{u}_r(t) \quad (3.10)$$

δt will be zero in the absence of anisotropy and thus component $u_t(t)$ does not exist and the energy on the $u_t(t) \simeq 0$. Before splitting, the initial polarization direction (ϕ_p) of SKS phase can be approximated as the radial polarization direction and it is equal to BAZ direction. Therefore, the seismogram is first rotated to radial (R) and transverse (T) directions and a grid search is performed to minimize the energy on the corrected $u_t(t)$. Since SKS splitting usually produces large δt compared to local and regional splitting, the grid search of ϕ and δt is done with 1° and 0.05 s

increments respectively. Instead of minimizing λ_2 as in the eigenvalue minimization, one can perform a grid search to find the minimum energy on the corrected transverse component ($\tilde{u}_t(t)$), calculating the energy $E_t(\phi, \delta t)$ on th ($\tilde{u}_t(t)$) for each $\phi, \delta t$ -pair using the following estimation:

$$E_t(\phi, \delta t) = \frac{1}{T} \int_0^T \tilde{u}_t^2(t) dt \quad (3.11)$$

Error estimation

Errors of the best splitting parameters ($d\phi$ and $d\delta t$) are determined from the 95 % confidence region in the 2-D contour plots (ϕ and δt) of $\lambda_{2(min)}$ or minimum energy ($E_t^{min}(\phi, \delta)$). *Silver and Chan* [1991] demonstrated that λ_2 is a measure of variance of the noise. Thus, the confidence region of the $\lambda_{2(min)}$ or minimum energy can be used to constrain the errors of the splitting parameters. $\lambda_{2(min)}$ values within the 95% confidence region satisfy the following relationship, determined from the F-distribution [*Silver and Chan*, 1991].

$$\frac{\lambda_2}{\lambda_{2(min)}} \leq 1 + \frac{k}{\nu - k} F_{k, \nu - k}(1 - \alpha) \quad (3.12)$$

Where, ν is the degrees of freedom, α is the 95% confidence interval ($= 0.05$), and k is the number of free parameters ($k= 2$: ϕ and δt). $F_{k, \nu - k}(1 - \alpha)$ is the inverse of the F-distribution. ν depends on the instrument response and noise spectrum and it can be determined from the data as it is a measure of number of uncorrelated time points in the digital-time series [*Silver and Chan*, 1988]. Errors of the best ϕ and δt , are the maximum lengths of the 95% confidence interval along the ϕ and δt axes respectively (contour plot in Fig. 3.8). A detailed explanation on this error estimation is given in *Walsh* [2012].

In the case of transverse energy minimization, 95% confidence region of the ϕ and δt values agrees with the relationship: $E_t(\phi, \delta t)/E_t^{min}(\phi, \delta t) < F_\alpha(\nu, \nu)$, where, $E_t^{min}(\phi, \delta t)$ is the threshold limit of the 95% confidence ($\alpha = 0.05$) and ν is the degrees of freedom [Silver and Chan, 1988]. This error calculation of SC91 is being re-evaluated by [Walsh et al.], but this thesis uses the original error estimation procedure.

3.2.3 Mfast

The Mfast technique [Wessel, 2010, Savage et al., 2010] is a fully-automated shear-wave splitting software, which was developed on the basis of the eigen-value minimization method [Silver and Chan, 1991] and the cluster analysis method [Teanby et al., 2004b]. The method is suitable for the routine analysis of splitting measurements on a large set of local and regional events.

Before processing data with Mfast, S arrivals are hand picked on the 3-component seismograms with the help of expected arrival times that are determined using the Taup toolkit [Crotwell et al., 1999]. During picking, clearly visible poor quality seismograms can be eliminated. The first step of Mfast involves searching over a given set of 14 band pass filters, which are two-pole Butterworth filters with frequency band-widths within the range of 0.4 Hz to 10 Hz, and calculating the signal-to-noise ratio (SNR) to find filters that give the highest SNR-bandwidth product (≥ 3) for each earthquake-station pair.

- **SNR:** The SNR of the filtered wave is determined from the 3 second window lengths relative to the S pick ($t=0$ s) before (noise) and after (signal). Mfast assumes 0.05 s S arrival errors and, thus, noise window ranges from -3.05 to -0.05 s and signal window ranges from 0.05 to 3.05 s. SNR is calculated taking the minimum amplitude ratios of the Fast Fourier Transform of the noise and

the signal windows.

Then the signals obtained from the selected filters (maximum up to three) are used to estimate the splitting parameters. The dominant period of each S-phase is determined from the frequency that corresponds to the frequency with the highest amplitude of the amplitude spectrum (f_d). The amplitude spectrum is calculated from the three second window directly after the S-pick. This period is then used to determine the minimum and maximum cutoff limits of the shear-wave analysis window. The main steps associated with the Mfast are shown in Fig. 3.9.

Figure 3.9 Processing steps of the Mfast program [after *Savage et al.*, 2010].

Cluster analysis minimizes uncertainties in the splitting parameters that are caused by the selection of S-wave analysis windows [*Teanby et al.*, 2004b]. This method

finds the analysis window that gives the most stable splitting parameters with small errors by performing splitting measurements through a given set of analysis windows that are defined between minimum (W_{s1})/maximum (W_{s2}) starting times and minimum (W_{e1})/maximum(W_{e2}) end times (Fig. 3.10). Then the most stable splitting parameters are chosen by applying the cluster grading criteria.

Mfast uses several grading criteria to extract good quality splitting measurements, to minimize the filtering effects on the splitting measurements, and to minimize the scatter in splitting parameters at a single station. Depending on the purpose of analysis, it is possible to use one or more grading schemes for better evaluation of the splitting measurements. The grading criteria used in this study are discussed below.

1. Grading based on the cluster grading and errors associated with splitting measurements

AB graded measurements: The splitting parameters that are given as A and/or B grade criterion (here after AB grade), which is defined in *Savage et al.* [2010], are considered as good quality splitting measurements. Both cluster parameters [*Teanby et al.*, 2004b] and measurement errors have been used to define these grades. We only used AB grade measurements to investigate the anisotropic structure. AB grading agrees with the following criteria:

- $\text{SNR} > 3$, $\delta t < 0.8 * tlag_{max}$ ($tlag_{max}$ is the maximum δt for grid search), maximum error in $\phi < 25^\circ$, if it is not a null measurement, and if the measurement graded as a A or B cluster in the cluster grading. Note that the A grade criteria (a. $\text{SNR} > 4$, b. $\delta t < 0.8 * tlag_{max}$, c. not null and $d\phi < 10^\circ$) are a subset of AB grade criteria, which use more narrow limits for SNR and $d\phi$ compared to AB grade criteria.

- Null measurements are those for which ϕ_p is within 20° of parallel or perpendicular to ϕ (Fig. 3.13). Such measurements can lead to incorrect δt (section 3.3.1).
- B cluster grade [Savage et al., 2010] has $\text{var}(k) < 5 \cdot \text{var}(k)_{\text{best}}$ and $N_{\text{meas}}(k) > N_{\text{cmin}(5)}$ and also $t_{\text{diff}}(k) > t_{\text{lag}_{\text{max}}}/8$ or $180^\circ/8 < \text{phi}_{\text{diff}}(k) < 7 \cdot 180^\circ/8$ (where k , $\text{var}(k)$, $\text{var}(k)_{\text{best}}$, $N_{\text{meas}}(k)$, $N_{\text{cmin}(5)}$, $t_{\text{diff}}(k)$ and $\text{phi}_{\text{diff}}(k)$ are the cluster number, average variance, variance of the best cluster, no of measurements in the cluster, minimum no of events per cluster ($= 5$), time difference, and angular distance respectively within cluster k). Here, $\text{phi}_{\text{diff}} = |\text{phi}(k) - \text{phi}(k)_{\text{best}}|$ and $t_{\text{diff}} = |t(k) - t(k)_{\text{best}}|$. Note that the A grade cluster is a sub-cluster of the B grade cluster.

2. Stable splitting measurements from different filters

If a measurement agrees with the above quality scheme, one can extract the splitting measurements that give the most stable measurements for different filters. However, such a grading is not recommended for a study examining frequency dependent splitting using the same waveform, but different filters. This criterion extracts the measurements that give a similar answer, i.e. if the time difference $< t_{\text{lag}_{\text{max}}}/8$ or if the angular distance $< \pi/8$, for each filter (refers to ‘best filter measurements’ in later sections).

3. Highest quality splitting parameters

This criterion allows us to extract very well constrained splitting measurements with dense error contours around the solution that satisfies the AB best filter grading criteria (as above). Error contours represent the contours of the minimum eigenvalues or the low energy surfaces. Savage et al. [2010] defines λ_{max} as the value that corresponds to maximum error contour when the contours around the 95% confidence level are re-scaled assuming that 95% confidence contour is 1. The highest quality measurements defined in chapter 4 are those of which have A grade best filter

measurements and $\lambda_{max} > 8$ (the highest quality). In chapter 5, measurements with $\lambda_{max} > 5$ are chosen as high quality measurements.

Teanby et al. [2004b] and Mfast cluster analysis

Teanby et al. [2004b] conducts shear-wave splitting measurements on several analysis windows (N_{total}) that are defined using the set parameters in the configuration file. Mfast modified this method to generate a customized configuration file for each event considering the dominant frequency (section 3.2.3) of the S signal and the S arrival time.

Figure 3.10 Defining a set of S-wave splitting analysis windows relative to S-arrival.

The number of start-time windows (N_{beg}) are defined between W_{s1} and W_{s2} with the steps of ΔT_b and the number of end-time (N_{end}) windows are defined between W_{e1} and W_{e2} with ΔT_e steps relative to the S-pick (Fig. 3.10). The minimum and maximum window lengths are chosen to be 1 s ($1/f_d$) and 2.5 s long respectively for local/regional splitting analysis (Fig. 3.10).

After measuring best splitting parameters on N_{total} number of splitting analysis windows, clusters of similar splitting measurements are identified from the criteria based on the variance of clusters and the individual measurements [*Teanby et al.*, 2004b]. These clusters can be identified on the ϕ versus δt plots (measurement shown in Fig. 3.11-a has 6 clusters). The best cluster is identified as the cluster with

Figure 3.11 (a) Stable cluster regions (6 red triangles) that are determined from the cluster analysis [Teauby *et al.*, 2004b]. The best cluster is marked by the blue cross. (b) Splitting parameters versus window number. The most stable measurement with in the best cluster has the minimum splitting errors compared to other measurements.

the lowest variance and low individual measurement errors. Thus, the best splitting measurements are extracted from the best cluster by finding the measurement with the smallest ϕ and δt errors (Fig. 3.11-b).

Mfast incorporates the cluster analysis parameters of the best cluster and the other clusters (e.g. cluster number, average variance, variance of the best cluster, number of measurements in each cluster, time difference and angular distance between best cluster and the other clusters) to grade the splitting measurements of each event as one of the quality control steps (B cluster grade is explained in the section 3.2.3). This grading compares the properties of the best cluster with the other clusters and removes the measurements with similar cluster properties to the best cluster, but with different splitting measurements [Savage *et al.*, 2010].

3.3 Splitting parameters: limitations, quality and statistics

Seismograms with clear S-wave onset are preferred to obtain good splitting measurements. However, in most of the cases seismograms display complicated waveforms due to complexities in the subsurface structures and/or interaction of the wave forms at the interfaces of the layered media. Sometimes even if the seismogram has a clear S arrival, splitting measurements can be distorted by different other phenomena (section 3.3.1).

The polarization of an S-wave incident at the free surface is governed by many factors such as the angle of incidence, frequency of the incident wave, the source depth, and the layering of the crust [Booth and Crampin, 1985]. To minimize the effect of incidence angle on S-wave polarization, splitting analysis is usually carried out inside the shear-wave window (i.e. a few degrees less than critical incidence angle) [Evans, 1984, Nuttli, 1961]. Wave fronts with incident angles beyond the critical angle ($i_c = \sin^{-1}[V_s/V_p]$, where V_s and V_p are the surface S and P velocities) are likely to display complex particle motion due to phase conversions at the free surface. This will cause particle motion to be non-linear even in the absence of anisotropy [Nuttli, 1961] and make the interpretation of S-wave polarization complicated. Thus, splitting analysis is usually restricted to incident angles less than the i_c . However, the i_c can be greater than 35° due to curvature of the wave front of the local ray paths in the presence of low-velocity surface layer [Booth and Crampin, 1985]. In on-land investigation, we assume a low-velocity surface layer with 2.6 km/s S velocity, by taking into account proposed velocity models for South Island [Eberhart-Phillips and Bannister, 2010, Eberhart-Phillips and Reyners, 2001]. In South Island, most of the seismicity is limited to the plate-boundary zone and this narrow distribution significantly reduces the number of local events that can be used to measure the crustal anisotropy far away from the plate-boundary. Thus, the average velocity structure allows us to obtain splitting parameters from ray paths

with larger epicentral distances. Although these long ray paths of local/regional S-waves reach the surface sub-vertically, they are likely to attain a horizontal/sub-horizontal propagation when considering the full curved ray-paths. These sub-horizontal/horizontal ray segments can be sensitive to horizontal anisotropic layers with vertical symmetry axis [Paulssen, 2004], resulting in an integrated effect of anisotropy with different symmetry configurations along the ray paths.

As discussed in the section 3.1, ambiguous splitting parameters are primarily caused by the noise in the seismogram. The application of a proper filter can minimize the effect due to noise and enhance the SNR prior to splitting. However, there are some other cases when erroneous splitting parameters can occur. Such splitting measurements may lead to wrong interpretations. Thus evaluation of splitting parameters should be done with an extreme care to avoid ambiguous splitting measurements. Such measurements are visually clear in the diagnostic plots and can be minimized by applying automatic and manual grading criteria to splitting measurements or sometimes eliminating signals with poor quality waveforms prior to splitting analysis. The Mfast consists of such elimination criteria that allows us to extract robust splitting measurements. The following section explains how to diagnose good splitting measurements and different mechanisms that can produce erroneous measurements.

3.3.1 Diagnosing robust splitting measurements

The diagnostic plots that are produced during splitting can be used to identify the good quality splitting measurements, especially in the case of manual grading. Those plots also have a mathematical meaning that can be incorporated with the automatic grading. Quality measurements usually exhibit following characteristic properties:

Shear-wave analysis window

The selected shear wave analysis window should have a clear S-wave onset with high SNR ($\text{SNR} \gtrsim 2$). Energy on both horizontal components is a clear indication of the S-wave splitting.

Contour plots

Contour plots are generated by contouring the second eigenvalues (or minimum energy) that are obtained from the covariance matrix between two horizontal components for different pairs of ϕ and δt values in $\phi, \delta t$ -space. The contours in those plots, therefore, indicate equal energy surfaces or eigenvalue surfaces. The best splitting parameters can be defined as the values correspond to the smallest second eigenvalue ($\lambda_{2(\min)}$) and the location of this minimum is surrounded by several contours with clear minimum contour (contour plot in Fig. 3.8-3). Good splitting measurements have well defined ellipsoidal contours and a symmetric small 95 % confidence area that indicates a well constrained splitting measurement. According to SC91, non-ellipsoidal contours indicate no splitting due to: $\delta t = 0$, $\phi_p = \phi$, or $\phi_p = \phi + 90^\circ$.

Wave form plots

The waveform plots of the fast and slow components before and after correction for splitting can be used to examine how well they correlate each other (see Fig. 3.8-2). When fast and slow components of an original seismogram are rotated on to fast and slow directions by ϕ and $\phi + 90$, waveform plots show a clear separation with δt time lag between two components and when are they shifted by δt , the two waveforms must match well each other, indicating the unsplit stage of the S-wave.

Once the waveform is unsplit, energy of the waveform must be visible only on the component parallel to the initial polarization direction (ϕ_p) and energy on the other component must be minimized. To check this, corrected fast and slow components are rotated into ϕ_p and the component perpendicular to initial polarization direction

must only display a very little energy that corresponds to the noise in the seismogram (Fig. 3.8-2).

Particle motion plots

Presence of energy on both fast and slow components denotes the split S-wave. The horizontal particle motions of the fast and slow components, thus display an elliptical particle motion as in the particle motion plot in Fig. 3.8. Once the energy on the component perpendicular to initial polarization direction is minimized or the covariance matrix is nearly singular, the particle motion of the two components is nearly linear. The eigenvector corresponding to the larger eigenvalue of the nearly singular covariance matrix therefore represents the particle motion in the horizontal plane, which is mostly linear (Fig. 3.8-2). The direction of this linear particle motion is equivalent to the ϕ_p direction of the unsplit S-wave.

Cycle skipping

Cycle skipping is caused by the mismatch between fast and slow components. This results in a shift in δt , giving either higher or smaller δt compared to correctly fit wave-forms. Cycle skipping is often caused by the S-waves with narrow frequency ranges and/or filtering of the S-waves using narrow band-pass filters. Such a filtering, or the earth's natural filtering may result in sinusoidal waveforms, which are likely to cycle skip. Erroneous fit can happen either in half or full cycles of the wave period, so that the calculated δt from cycle skipped splitting measurements δt_{cyc} can be mathematically represented as [Gerst, 2003]: $\delta t_{cyc} = \delta t \pm n \cdot \frac{T}{2}$

where T is the period of the wave and n is an integer. As shown by the equation a mismatch may occur in two different ways:

1. Trying to fit the fast component with the wrong slow component that arrives one or several half-cycle periods later. This happens if the first arrival of the

slow component has a weak onset. In this case δt is enlarged by $n \cdot \frac{T}{2}$. Such a fitting will only produce error in δt , but not in ϕ (see Fig. 3.15).

2. Switching between slow and fast components. This is possible if the first arrival (fast wave) is weak and the slow component can be mistaken as the fast component. As a result, resolved initial polarization direction will have 90° error and δt will be reduced by $n \cdot \frac{T}{2}$. In such situation several minima can appear 90° apart each other along the fast direction axis.

Figure 3.12 An example of cycle skipped measurement that explains case one (see text). Right: wave form plots of before and after correction for splitting, indicating a waveform mismatch and an enlarged δt . Left: contour plot of errors surfaces (gray contours) with the best splitting parameters (blue cross) and 95% confidence limit (black contour) of the best splitting parameters.

In the case of addition of full or half cycles of wave-periods, δt will display unusually high splitting delays. Hence, the elimination of splitting measurements with high δt is one of the ways to alleviate ambiguities due to cycle skipping. Cycle skipping is quite obvious in the contour plots as they consist of several minima (or similar shape minimum contouring) that are separated by $n \cdot \frac{T}{2}$ along the δt axis. In such cases,

correct measurements are recognized as the minimum with the densest contours

Null measurements

Identifying null measurements is one of the main quality control steps, as those measurements may lead to an erroneous interpretations. Null measurements do not produce split S-waves. Such measurements are possible, if the S-wave satisfies the following conditions [*Silver and Chan, 1991*]:

1. If the initial polarization direction of the unsplit S-wave (ϕ_p) is aligned with either fast ($\phi_p = \phi \pm 180^\circ$) or slow ($\phi_p = \phi \pm 90^\circ$) polarization direction (see Fig. 13(a)).
2. If the media is isotropic or has a vertical axis of symmetry, vertically-traveling S-waves show no splitting and, thus, δt is zero.

If an S-phase satisfies either of the above conditions, the covariance matrix between two horizontal components will yield the same minimum λ_2 for the grid search pairs of ϕ with $\delta t = 0$ and/or the pairs of different δt with $\phi = \phi_p \pm 90^\circ$. The calculated minimum λ_2 , therefore, only represents the correction of noise in the perpendicular component. Thus, δt can not be determined accurately. Null measurements can be clearly recognized on the contour plots (Fig 3.13) with U shape contours [*Gerst, 2003*] and the wave-form plot with little or no energy on the direction perpendicular to ϕ .

Mfast takes into account the cases where null measurements are likely and removes their effects by eliminating the measurements that agree with the above conditions. As explained in the null criteria, it retains only all the measurements with ϕ that agree with: $20^\circ < (\phi - \phi_p) < 70^\circ$ (Fig. 3.13-b). And also, selecting measurements with high SNR (>3 or 4) ensures that the measurement consists of a clear signal.

(a) The contour plot of the null measurement (blue cross) that agree with the first condition. Note the U shape contours and large δt (p. 101).

(b) Quality control step in Mfast to avoid possible null measurements.

Figure 3.13 Example of a null measurements (top) and application of the null-avoiding criteria to all on-land measurements (bottom).

Statistics of the splitting measurements

The application of grading criteria to individual splitting measurements allows us to extract robust splitting parameters and thus, minimize the scatter in splitting measurements at an individual station. Nevertheless, some statistical parameters and analysis are important to explain the single station measurements. Directional data such as ϕ exhibit 180° ambiguity and sometimes have more than a single mode of prominent azimuth within a single population. Therefore, both normal and directional statistics are important to determine statistical measures of the splitting measurements. Previous studies [e.g. *Gerst and Savage, 2004, Savage et al., 2010, Roman et al., 2011*] have successfully utilized such statistical calculations using the explanations given in *Mardia [1972]*. Mean ϕ , length of the mean resultant vector of ϕ (R), circular standard deviation (CSD) and standard error (SE) are some of the directional statistics that have been used in this study.

Mean ϕ

As ϕ is an angular data set, one cannot simply average ϕ measured at a station as in the normal statistics. To calculate mean angular direction, an individual fast azimuth (ϕ_i) should be first transformed into unit vectors in a rectangular coordinate system, so that they can be represented as a point on the 2-D plane. The unit vector ($\bar{r}_i$) of a single measurement can be defined as: $\bar{r}_i = \{\cos f\phi_i, \sin f\phi_i\}$. The directional ambiguity associated with ϕ_i is removed by doubling the ϕ_i and thus $f = 2$ [*Jammalamadaka and Sengupta, 2001*]. The mean resultant vector $\bar{r}$ is:

$$\bar{r} = \begin{pmatrix} \frac{1}{n} \sum_{i=1}^n \cos(2\phi_i) = \bar{C} \\ \frac{1}{n} \sum_{i=1}^n \sin(2\phi_i) = \bar{S} \end{pmatrix}$$

The mean fast azimuth (ϕ_s) is, therefore equal to:

$$\phi_s = \arctan(\bar{C}/\bar{S}) / 2 \quad (3.13)$$

Mean resultant vector length (R), Standard error (SE), and circular standard deviation (CSD)

R is determined from:

$$R = |\bar{r}| = \sqrt{\bar{C}^2 + \bar{S}^2} \quad (3.14)$$

R is a quantitative estimate of circular spread and it is used to determine both standard deviation of the ϕ_s (SE) and SCD of ϕ . If the angular data set is more concentrated along one direction, R is close to 1.

SE of ϕ_s is calculated using R and the concentration parameter (κ), which is a parameter of a Von Mises distribution ($VM_{(\phi_s, \kappa)}$), and SE indicates the accuracy of the ϕ_s . κ can be estimated from R . $VM_{(\phi_s, \kappa)}$ is a continuous distribution on the range of $0 - 360^\circ$ and it is a probability density function that varies with ϕ_s and κ [Mardia, 1972]. Thus, $VM_{(\phi_s, \kappa)}$ is considered to be a circular equivalent of the normal distribution that measures the circular spread. Its probability density function is given by [Mardia, 1972, Philipp, 2009]:

$$p_{(\phi_i, \phi_s, \kappa)} = \frac{1}{2\pi I_0(\kappa)} \exp[\kappa \cos(\phi_i - \phi_s)],$$

where, I_0 denotes the modified Bessel function of order zero. SE can be mathematically expressed as [Philipp, 2009]:

$$SE = \frac{1}{f\sqrt{nR\kappa}} (f = 2) \quad (3.15)$$

Circular variance ($S = 1 - R$) is also a measure of circular spread. S of concentrated and scattered data will be close to 0 and 1, respectively.

Circular standard deviation of the ϕ_i (SCD ϕ) determines the spread of the

measurements around the ϕ_s . SCD of ϕ is defined incorporating S [Mardia, 1972] as:

$$SCD_\phi = \sqrt{-2\ln R} / f \quad (f = 2) \quad (3.16)$$

It is advised to carry out tests for randomness of the data, before analyzing the above statistical measurements. The most commonly used method is *Rayleigh test*, which is ideal to apply for ϕ with unimodal (only one significant ϕ) distribution. This determines the threshold value for R to deviate from uniformity depending on the number of measurements and the significance level.

We have calculated most of the statistics discussed above from the ϕ of our local/regional shear wave splitting measurements. However, when analyzing splitting parameters spanning a large area and depth range, the above statistical estimations at single stations may not provide accurate constraints on the anisotropy as the scatter in splitting parameters or multi-mode distributions can be associated with the heterogeneous, spatial and depth varying anisotropy. In such cases, the elimination of station averages or measurements on the basis of statistical parameters is not suitable.

3.4 Direct estimations from splitting parameters

Other than the alignment of the anisotropic medium, ϕ and δt values can be used to obtain some quantitative estimations of the anisotropic medium. Some of the examples are given below:

Anisotropic layer thickness:

Assuming that the anisotropy is concentrated in a single homogeneous layer anisotropy, *Vinnik et al.* [1989a] and *Silver and Chan* [1991] (see equation 2.8) provide the following approximations to calculate the anisotropic layer thickness.:

$$\frac{\delta t}{t} \simeq -\frac{\delta\beta}{\beta} = \frac{\delta t \cdot \beta_0}{L} \quad (3.17)$$

Here, t denotes the average propagation time of the S-wave in the anisotropic medium with layer thickness L . $\delta\beta$, β and β_0 represent velocity difference between fast and slow waves, the average and isotropic S-wave velocities respectively. *Vinnik et al.* [1989a] demonstrated that 70% anisotropic upper- mantle rock that contains 30% transversely isotropic olivine ($\delta\beta/\beta \simeq 0.02$) should have ~ 230 km layer anisotropic thickness to produce 1 s δt .

Crack density estimation:

One common assumption is that anisotropy observed in the crust is solely due to crack induced anisotropy. In some tectonic settings such as volcanoes, geothermal areas and large-scale fracture zones, this assumption is useful to calculate crack density. Crack density of a medium can be determined the number of cracks (N) with radius a in a unit Volume(V): $\xi = Na^3/V$. Thus the normalized crack density of a unit cubic volume is equal to a^3 [*Hudson, 1980*]. Crack density is equal to the $SWA/100$ (section 2.4.1) and can also be expressed as : $\xi = \delta t \cdot \beta_0 / L$ [*Peng and Ben-Zion, 2004, Hudson, 1980*].

Figure 3.14 Olivine deformation under pure shear and simple shear conditions. Note that the extension axis (e_3) is always perpendicular to shortening direction (e_1) in the case of pure-shear deformation and it does not rotate with the increase in shear ($\Phi = 0$). But in the case of simple shear, extension axis rotates with respect to the shear plane when changing shear. The strain along the extension direction ($|e_3|$) can be calculated from the equations given in the figure (also see text).

Finite strain and shear-direction

Splitting parameters from the S-phases that sample mantle anisotropy can be used to extract information on the magnitude ($|e|$) and the direction of the strain (e_i). Here, i defines three principal strain directions: shortening (e_1), intermediate (e_2), and extension (e_3). Naturally deformed Olivine [100] is predicted to align with the extension direction [Ribe, 1992]. This phenomenon can be used as a proxy for mantle deformation and flow due to relative motion of the plates (RPM) at plate-boundaries and to determine shear or flow associated with absolute plate motion (i.e. differential motion between the overlying plate and mantle). These two processes are comparable with the progressive simple-shear type deformation.

In the case of pure shear deformation of Olivine, [010] is parallel to shortening axis

and $[100]$ is aligned with the extension direction, so that ϕ is parallel to extension. Extension via simple shear deformation is predicted to change the orientation of the extension axis, rotating towards the shear plane with increasing shear (or finite strain). As a result, the angle between the extension direction and shear plane (Φ ; see equation 1.1) becomes smaller [Ribe, 1992] with increasing finite strain. Finite strain associated with this extension (or progressive shear) at a given Φ can be estimated from the relationship : $|e_3| = \cot(\Phi)$ [Molnar et al., 1999].

3.5 Characterizing splitting parameters

It is important to extract the maximum possible information from the complex splitting data set to fully characterize the anisotropic structure. Complexities in splitting parameters are assumed to be caused by: 1) multiple layers of anisotropy, 2) frequency dependent splitting, 3) dipping anisotropic layers, and 4) presence of intrinsically anisotropic minerals with different crystal symmetry or crustal heterogeneities (mixed anisotropy). In some cases, observed splitting parameters can be modeled using forward modeling or inversion approaches to characterize such complexities.

3.5.1 2-layer and multiple layer modeling

The assumption of a single homogeneous anisotropic layer is sometimes not valid, especially in the cases of evaluating splitting parameters in complex tectonic settings. Such settings are likely characterized by layered anisotropic media. *Silver and Savage* [1994] introduced a forward modeling approach to explain the splitting parameters, where the anisotropic structure beneath the receiver is characterized by two or more layers of anisotropy with varying hexagonal symmetry orientations.

Splitting analysis that assumes a single homogeneous anisotropic layer beneath the station will yield apparent splitting parameters if there is more than a single layer of anisotropy. In the presence of two layers of anisotropy, S-waves can be double-split by two layers [Yardley and Crampin, 1991]. In that case, splitting can be described using two splitting operators Γ_1 and Γ_2 that correspond to splitting at each layer (Fig. 3.15). *Silver and Savage* [1994] showed that the apparent splitting operator Γ_a that is a function of both Γ_1 and Γ_2 can be used to characterize the splitting in each layer. They found that apparent splitting parameters from SKS phases exhibit $\pi/2$ periodic patterns with the initial polarization direction (ϕ_p) (Fig. 3.15-right).

Figure 3.15 Modeled apparent splitting parameters in two-layer anisotropic media [after *Silver and Savage*, 1994]. Left: Double split S-phase from two anisotropic layers (top layer: $\phi_2=140^\circ$, $\delta t_2=1$ s; bottom layer: $\phi_1=90^\circ$, $\delta t_1=1$ s). Right(top/bottom): Variation of apparent splitting parameters with respect to ϕ_p . Curved lines in Figs. **a** and **b** represent the predicted best-fitting splitting parameters from the 2-layer model at different S-wave periods (dotted line - 5 s, solid line - 8 s, and dashed - 20 s).

However, such a method requires apparent splitting measurements from a wide range of backazimuths to constrain the splitting parameters accurately. *Ozalaybey and Savage* [1995] modified the same inversion approach used in SC91 by inverting single SKS/teleseismic S waveform with high SNR to determine splitting parameters

of the two layers. Some studies have extended the 2-layer modeling approach to characterize the behavior of apparent splitting parameters in multiple-layered anisotropic media [e.g. *Rumpker and Silver*, 1998, *Walker and Wookey*, 2012]. *Rumpker and Silver* [1998] pointed out that the splitting parameters are biased towards the topmost layer anisotropy if short-period ($T/\delta t \leq 1$) S-waves vertically propagate through a medium with smoothly varying anisotropic alignments in each layer.

3.5.2 Frequency dependent splitting

Frequency dependent S-wave splitting has been considered as another phenomenon that can be used to describe the complexities in anisotropic structure beneath a station. *Favier and Chevrot* [2003] adapted a finite-frequency approach, assuming the transverse component observed in SKS splitting represents the anisotropic perturbations of the elastic parameters and derived sensitivity kernels for δt (or splitting intensity). They demonstrated that the size of the sensitivity kernel is proportional to the period of the S-wave (i.e., it is inversely proportional to the S-wave frequency) in an anisotropic medium. Thus, if the medium consists of variable anisotropic structure with different length-scales, S-waves with different periods will extract different information about the anisotropic medium because those phases sample the medium differently.

As mentioned in the section 3.5.1, apparent splitting parameters from high frequency S-phases are mostly sensitive the topmost layer of anisotropy in the layered anisotropic medium. Thus, the higher frequency phases allow us to investigate the properties of the shallow anisotropic media (e.g. crustal anisotropy). Especially when utilizing deep high-frequency S-phases, it is important to clarify whether the splitting parameters actually agree with the deep anisotropy or if they represent the shallow structure. Recent studies investigating anisotropic structures beneath

subduction zones found frequency dependent splitting [Wirth and Long, 2010, Long, 2010] and interpreted it as possibly multi-layer anisotropy in subduction tectonic settings [Huang *et al.*, 2011a, Greve and Savage, 2009].

Marson-Pidgeon and Savage [1997] provide a set of splitting observation from SKS/ScS phases recorded in Wellington region, New Zealand that indicates frequency dependent shear-wave splitting. They noticed a clear increase in δt with the periods of the S-waves and explained it in terms of three phenomena that might cause frequency dependent splitting: 1) the previously discussed dependency of size of the sensitivity kernel (splitting resolution) on the period; 2) multiple layers of anisotropy beneath the station; and/or 3) scattering of seismic waves by heterogeneous crustal structures. Frequency dependent splitting due to crustal structures have been investigated by several studies (e.g. Liu *et al.* [2003], Werner and Shapiro [1999], and Tod and Liu [2002]). As suggested by Marson-Pidgeon and Savage [1997] and Liu *et al.* [2003], frequency dependent splitting can be attributed to scattering of S-waves by large-scale crustal heterogeneities.

According to Sato and Fehler [1997], anisotropic heterogeneities can be examined as effective anisotropy only if the scale-lengths of heterogeneities are small enough to preclude scattering, so that the medium behaves as a quasi-homogeneous medium. In that case, an empirical relationship between wave-number (k ; $k = 2\pi/\lambda$, where λ is wavelength) and the scale of heterogeneities ($\mathbf{a}$) is given by: $k\mathbf{a} < 0.01$ [Wu and Aki, 1988]. For instance, a waveform with 5 s period (typical ScS period in our data set) requires heterogeneities with scale-lengths less than ~ 38 m to observe effective anisotropy without scattering. For local S phases (period of 1 s), the maximum scale length is around 5.2 m. Thus, the large-scale heterogeneities will affect δt due to scattering and can be seen as an effect of frequency dependency [Marson-Pidgeon and Savage, 1997]. We also noticed frequency-dependent splitting from both local and teleseismic S-phases and these results are discussed in sections 4.4.2, 6.2.1, and appendix C.5.

3.5.3 Dipping anisotropic layers

Dipping anisotropic layers have been first investigated in central Europe by *Babuska et al.* [1984] from P-wave residuals analysis. The study found directional dependence of P-wave residuals and those directional variations are interpreted as dipping anisotropic structures that are caused by paleosubduction systems. In such domains, shear-wave splitting parameters from SKS phases are expected to show backazimuthal dependence [*Levin et al.*, 1999] (Fig. 3.16) and will produce 2π periodicity with BAZ [*Silver and Savage*, 1994].

Figure 3.16 Modeled splitting parameters in systems that have a single hexagonally anisotropic layer with horizontal symmetry axes (left) and symmetry axes dipping at 45° (right) [after *Levin et al.*, 1999]. Note that the slowly (left) and rapidly (right) varying splitting parameters in two systems.

In the case of dipping anisotropic layers, the anisotropic tensor is inclined (inclined symmetry axis) and, thus, the plane perpendicular to symmetry axis also inclined. In such a system shear-wave splitting from vertical ray paths results in lower effective splitting anisotropy as in the two layer case. *Sileny and Plomerova* [1996] have modified *Silver and Chan* [1991] to measure S-wave splitting parameters 3-dimensionally by calculating them in the plane perpendicular to the ray path. Splitting parameters are measured in the ray coordinate system (LQT), solving for the dip of anisotropic alignment. Here, the fast direction (ψ) that is measured

in the QT plane can be expressed using the fast azimuth (ϕ) in the North-EAST (NE) plane and its inclination (θ) on the vertical plane (Fig 3.17).

Figure 3.17 Splitting parameter estimations in the LQT-coordinate system. The fast polarization direction (ψ) is now measured in the Q-T plane, which is perpendicular to the ray direction (L). ψ can also be expressed by ϕ that is measured in horizontal plane (NE) with geographic coordinates (N-E) and θ , which is a measure of inclination from vertical axis (Z) in the vertical plane (N-Z) [Figure after *Vecsey et al.*, 2008].

The SPLITSHEAR package of *Vecsey et al.* [2008], which is based on the *Sileny and Plomerova* [1996] method, facilitates the measurement of splitting parameters in LQT coordinate system using three different techniques: 1) transverse energy, 2) eigenvalue minimization and 3) cross correlation techniques. The method calculates the misfit function between Q and T components that are corrected for splitting. The minimum of the misfit function, therefore, represents the best splitting parameters. The misfit function of the smaller eigenvalue of the correlation matrix is given by [Vecsey et al., 2008]:

$$\lambda_2 = \frac{1}{2}[(C_{ff} + C_{ss}) - \sqrt{(C_{ff} + C_{ss})^2 + 4C_{fs}^2}] \quad (3.18)$$

where, C_{ff} and C_{ss} are auto-correlation of the rotated components and C_{fs} is the cross-correlation.

3.6 Shear-wave splitting tomography

Local splitting parameters measured at a single station from ray paths with different ray propagation directions (or BAZ) often yield complex splitting measurements. Those complexities can be attributed to spatially and/or depth varying anisotropy that the ray encounters when it traverses from source to station. 2-D delay-time tomography allows us to resolve spatially varying anisotropic structure. However, if the rays are contaminated with depth varying anisotropy (due to either anisotropic layers with variable symmetry axis or different sources of anisotropy) as in subduction zones, 3-D splitting tomography techniques are useful to resolve the complexities in splitting parameters.

3.6.1 2-D delay time tomography and spatial averaging (TESSA)

To examine spatial variations of δt and ϕ associated with local splitting parameters, 2-D delay-time tomography and a spatial averaging technique (TESSA; *Johnson et al.* [2011]) has been utilized. Such an application allows us to obtain approximate constraints on regions with low and high anisotropy. δt from a single measurement is an estimate of the anisotropic structure along the ray path from source to the station and the δt is considered to be proportional to path length (L) of the ray through the anisotropic medium [*Silver*, 1996]. TESSA is based on this simple linear relationship between δt and L .

It assumes that δt observed along the ray path δt_r is the summation of δt calculated for each grid block (δt_b) with variable strength of anisotropy (s_b) in 2-D space. First, straight-line ray tracing between station and events determines the lengths of ray segments (L_{rb}) that lie on each grid block (Fig. 3.18). Then the total anisotropy

Figure 3.18 Schematic shows the calculation of total anisotropy (δt_r) along a given ray path (r) in TESSA. Here, b denotes the block index that is a function of the ray index (r) and the block count (i) from event to station along the ray path, s_b strength of anisotropy in block b , and L_{rb} is length of the ray segment in the block b .

measured at the station for a given event is calculated by,

$$\delta t_r = \sum_{b=1}^n (s_b \times L_{rb}) \quad (3.19)$$

The inversion part of TESSA determines the s_b for resolved blocks and is carried out using the MATLAB medium-scale optimization inversion function that finds the linear least-squares solution of the problem. This solution is then used iteratively to find the model solution for the strength of anisotropy of the whole ray (s_r) and this process is subjected to boundary conditions. As explained in *Johnson et al.* [2011], boundary constraints are applied in such a way that the minimum s_r will always be greater than 0 s/km and the maximum s_r will not exceed $\delta t_{r(max)}/L_b$, where $\delta t_{r(max)}$ is the maximum δt observed along a ray path and L_b is the width of the grid block. Then the weighted model solution (m) is:

$$m = (G^T W_d G)^{-1} G^T W_d d \quad [Stein and Wyession, 2009] \quad (3.20)$$

where G denotes the matrix consisting of L_{rb} , d is the observed δt from each ray path and W_d is the weighting applied to data based on the errors. The covariance matrix (C) of the measurement and inversion errors is used to define the weighting, so that $W_d = C^{-1}$. The diagonal components C_{ii} and off diagonal components C_{ij} of this matrix are given by:

$$\begin{aligned} C_{ii} &= \sigma_d^2(i) + L_b^2 \times \bar{\sigma}_m^2 \times n_b(i)/n \\ C_{ij} &= \bar{\sigma}_m^2 \times L_b^2/n^2 \end{aligned}$$

Here, $\sigma_d^2(i)$, $\bar{\sigma}_m^2$, and $n_b(i)$ indicate the squared standard error of the i^{th} measurement, mean variance of the solution when the data are weighted as $1/\sigma_d(i)$ and the number of boxes that the i^{th} ray propagates through, respectively. Model resolution is constrained from the diagonal elements of the resolution matrix along with checkerboard tests.

In TESSA, spatial averages are calculated by averaging ϕ in a particular grid block once the ϕ calculated from each ray (event-station pair) is assigned to each ray segment in 2-D grid blocks between the event and station. Three different weighting schemes: $1/d$ (where, d is distance from station to grid block), $1/d^2$, and tomographic weighting (ω_{rb}), can be applied to individual ϕ before averaging in order to provide different constraints on the anisotropic structure of the region. For instance, to determine the spatial variations of ϕ close to the station, one can apply $1/d^2$ weighting as it heavily weights the measurements closer to the station. ω_{rb} is suitable if the study aims to weight ϕ considering the delay-time tomography results. We use the tomographic weighting scheme presented in TESSA to determine average ϕ in grid blocks. Tomographic weighting is designed to take into account variations due to heterogeneous anisotropic structure. The tomographic weighting function (ω_{rb})

was the anisotropy strength profile determined for each grid block (s_b) in the δt tomography, so that more weighting is given to the regions with higher strength of anisotropy. ω_{rb} is defined as the strength profile of the each grid block, normalized by δt_r [Johnson *et al.*, 2011],

$$\omega_{rb} = s_b / \delta t_r \quad (3.21)$$

Then, the weighted mean ϕ of the grid block b with n number of rays is given by azimuth ($\bar{\phi}_b$),

$$\bar{\phi}_b = \frac{1}{2} \arctan \left(\frac{\sum_{r=1}^n \cos(2\phi_r) \times \omega_{rb}}{\sum_{r=1}^n \sin(2\phi_r) \times \omega_{rb}} \right) \quad (3.22)$$

where b and r denote the block and ray number, respectively.

3.6.2 3-D splitting inversion: *Abt and Fischer* [2008]

The shear wave splitting tomography (SWST) technique by *Abt and Fischer* [2008] can be used to resolve the 3-dimensionally distributed complex splitting parameters in terms of crystallographic orientations. This approach is useful for examining the geometry of the deep anisotropic structure (e.g. lower crust and upper mantle), where there is a good depth distribution of splitting parameters. The method neglects the anisotropy in the upper-most crust as it is mainly caused by SPO anisotropy and assumes that the observed local splitting is governed solely by the preferred alignment of the intrinsically anisotropic minerals (e.g. Olivine, Pyroxene). We apply this technique to resolve complex splitting parameters observed in our local-shear wave splitting estimation, assuming that the deep ray paths, especially in the two subduction zones, are mainly effected by the mineral alignment.

The method has three main steps: 1) model parametrization, 2) forward modeling of splitting and, 3) inversion of shear-wave splitting parameters. Here, we discuss how

we adapt the above three steps to obtain the final model solution (section 5.3.3).

1. Data and Model parameterization:

We invert the local/regional S-wave splitting obtained from the on-land and off-shore studies. The model space of this data set is defined so that it includes all stations and events that were used to obtain shear-wave splitting parameters. We chose grid blocks with 25 km^3 volume to characterize the 3-D space [i.e., along the plate-boundary (400 km), across plate-boundary (475 km) and maximum depths of 250 km]. A total of 7020 ray paths (2982 events and 56 stations) were used for the inversion (Fig. 3.19). The scale of our study region is comparatively larger than the regions that have been investigated using this technique previously [Abt *et al.*, 2009].

Figure 3.19 2-D (top-left) and 3-D view (bottom and top-right) of the model space (dotted lines), ray paths (blue), events (black circles) and stations (green triangles). Ray-tracing was done using 1-D velocity model of AK135 [Abt and Fischer, 2008].

The orientation of the crystallographic axis and the strength of anisotropy of each block in the model space are determined from the average starting model [see *Abt and Fischer*, 2008, appendix A]. We assume a hexagonal crystal symmetry that is simulated from the orthorhombic single-crystal elastic constants of olivine and ortho-pyroxene as described in *Abt and Fischer* [2008]. Thus, in each block, crystallographic orientations are defined by two angles that represent the horizontal azimuth of the a-axis $[1\ 0\ 0]$ (θ) and the plunge of the a-axis (ψ). The strength of anisotropy (α) that is assigned for each block is a scalar parameter that ranges from 0-100 %. α takes into account the relatively high single crystal anisotropy of minerals (e.g. laboratory measurements) and reduces this high anisotropy to be comparable with δt that is estimated from the in-situ shear-wave splitting.

2. Forward modeling

Forward modeling of splitting parameters (predicted splitting) is done with the particle motion perturbation method [*Abt and Fischer*, 2008] for each ray path in the model. The effect of 3-dimensionally varying anisotropic structure on the rays that travel from earthquakes to stations is determined by taking into account crystallographic orientations and the strength of anisotropy of each block that the ray propagates through. The Christoffel equation (see equation 2.7) is used to determine the polarization directions and the phase velocities (eigenvalues) in each block (as in section 2.2.2) using the initial model parameters. In each block, the two components perpendicular to the ray propagation are time lagged and rotated to add the effects of anisotropy in each block as the ray propagates. Thus, this calculation encounters the effects of S-wave splitting in the layered anisotropic media as the anisotropy in each block can be different from the others.

3. Inversion

Inversion allows us to find the the best-fit model with 3-dimensionally varying

crystallographic orientations and the strength of anisotropy that best explains the observed splitting parameters. The method uses damped least-squares inversion iteratively to minimize the misfit between observed and predicted splitting parameters [e.g. *Tarantola*, 2005]. The improved model (M^{i+1}) after i^{th} iteration by perturbation of model parameters (in our case $\delta\theta$, $\delta\psi$, and $\delta\alpha$) in model M^i is given by the following linear relationship:

$$M^{i+1} = M^i + C_{mm}G^T[GC_{mm}G^T + C_{dd}]^{-1}\delta D^i \quad \text{Abt and Fischer [2008]} \quad (3.23)$$

Here, δD^i ($= [d_{obs} - d_{pred}] = [d_{obs} - GM^i]$) represents the data misfit. The G matrix consists of linear partial derivatives that are computed by finite difference [*Abt and Fischer*, 2008, equation 5]. This explains the variation in splitting parameters upon the perturbation of model parameters (5° for θ , ψ and 5% for α). During inversion, it recalculates the partial derivatives of the G matrix only if changes in orientations of the model parameters (θ and ψ) between iterations (i and $i+1$) are exceeded by 1° . This allows the linearity of partial derivatives to be kept in the inversion problem and to obtain accurate solution. C_{dd} denotes the diagonal component of the data covariance matrix (d_{obs} and d_{pred}), which represents the sum of the squares of the errors. C_{mm} is the a priori model covariance matrix and the diagonal component of C_{mm} is used to damp the changes to model parameters to maintain the linearity of partial derivatives. Data misfits ($\delta D_{\phi, \delta t}$) are weighted by the both predicted and observed splitting errors [see *Abt and Fischer*, 2008, equation 9].

Model resolution is determined using the resolution matrix ($R_m = C_{mm}G^T[GC_{mm}G^T + C_{dd}]^{-1}G$; *Tarantola* [2005]). The diagonal component of this matrix helps to determine well resolved model parameters. R_m varies between 0 and 1, indicating poorly and well resolved blocks, respectively. This helps to find the regions with lower resolution ($R_m < 0.25$ from the final iteration as defined by [*Abt and Fischer*, 2008]), which is mainly due to low ray-coverage.

In such regions, model resolution is improved by either combining model blocks into a larger block volumes or applying spatial smoothing [*Abt and Fischer, 2008*]. The blocks are combined part-way through the iterations once the solution is stabilized. The inversion also uses damping relaxation to test the model stability after initial model convergence. The relaxing damping will improve the model unless it achieved minimum data misfit. Preliminary results from the 3-D tomographic inversion are discussed in section 5.3.3.

Chapter 4

Seismic anisotropy and lithospheric deformation of the plate-boundary zone in South Island, New Zealand: inferences from local S-wave Splitting

This chapter presents a published manuscript [*Karalliyadda and Savage, 2013*] that includes results of the on-land seismic anisotropy study. Some parts in the introduction and methodology sections may contain some repetitive sections from the previous chapters. A sub-study, which was completed after this publication is included in the section 5.3.1.

Abstract

Insights into plate-boundary deformation are gained from the shear-wave splitting of local S phases that originate within the lithosphere of the South Island, New Zealand. Analysis of the splitting parameters from land stations reveals changes in both delay times and fast azimuths with earthquake depth, earthquake-station back-azimuth and initial polarization azimuth, suggesting both laterally and depth varying anisotropy. When the average results are examined as a whole via tomographic inversion and spatial averaging, consistent patterns in delay times and fast azimuths exist. Spatially averaged fast azimuths reveal a localized high strain zone in the southern central region of the South Island. Based on fast azimuths observed above 100 km depth, we suggest that the plate-boundary sub-parallel anisotropy that is produced by pervasive shear is mainly distributed within a zone extending ~ 130 km SE of the Alpine fault in southern South Island and is widely distributed in northern South Island. Average station delay times of $\sim 0.1 - 0.4$ s compared to 1.7 s SKS delay times from previous studies in South Island further suggests a deeper seated

anisotropic zone in some areas, but in other regions frequency dependent anisotropy could explain the observed splitting.

4.1 Introduction

Anisotropy beneath plate-boundary zones provides indirect insights into the tectonic deformation associated with plate interactions. Among geophysical techniques, shear wave splitting (S-wave splitting) has been a well established and commonly used technique in exploring seismic anisotropy [*Silver and Chan, 1991, Teanby et al., 2004b, Wustefeld et al., 2008*]. Splitting tomography is another method that has been utilized to resolve the heterogeneous anisotropy structure from initial splitting parameters and to provide constraints on the distribution of seismic anisotropy [*Abt and Fischer, 2008, Johnson et al., 2011*]. Usage of such techniques is important in local and regional splitting studies, because it incorporates the changes in splitting due to various factors (e.g. back azimuth, depth, orientation of crystallographic axes) that could result in complex splitting. However, such methods are highly dependent on the station coverage and the earthquake distribution.

The crustal contributions of anisotropy are usually ignored by splitting analysis that investigates mantle anisotropy using SKS phases [*Savage, 1999*]. Yet crustal contributions might have significant effects on SKS measurements and result in complex patterns and small-scale lateral variations [*Long and Silver, 2009, Mattatall and Fouch, 2007, Rumpker and Silver, 1998*]. Thus, S-wave splitting of local and regional S-phases is vital for understanding the crustal and subcrustal contribution to total anisotropy in a given region. S-wave splitting using local events observed along the ray paths in the lithosphere is attributed to different sources of anisotropy that are developed from both shape preferred orientation and lattice preferred orientation. Seismic anisotropy at crustal depths is assumed to result from either stress-induced anisotropy or structural anisotropy [e.g., *Boness and Zoback, 2006*].

In the upper crust, fluid-filled microcracks or fractures and structural anisotropy (e.g. crustal faults, rock foliations or beddings) may affect S-wave splitting whereas in the lower crust and the upper mantle, splitting is mainly influenced by intrinsic anisotropy resulting from rock fabrics and preferred orientation of rock-forming minerals such as olivine in the upper mantle and mica and amphibole in the lower crust [Mainprice and Nicolas, 1989, Crampin *et al.*, 2004, Silver, 1996]. As suggested by Wilson *et al.* [2004], sub-horizontal foliations with lineations (e.g. schist fabric) developed at lower crustal depths due to pervasive shear in strike-slip fault zones could also act as source of anisotropy.

The interpretation of local splitting is challenging because of the different factors and sources of anisotropy (e.g. tectonic processes) that control splitting. Lattice preferred orientation is mainly a result of finite strain [Nicolas and Christensen, 1987]. In the lower crust, mineral alignment is predicted to be parallel to the flow or shear direction and in the upper mantle, the [100] axis of olivine is fast and is parallel to the extension (maximum principal stretch) direction or to the flow direction if significant strain has developed in a dry regime [Nicolas and Christensen, 1987, Savage, 1999, Silver, 1996]. At shallow crustal depths, differential stress results in preferential closure of micro cracks perpendicular to the maximum stress direction creating an anisotropic medium [Crampin *et al.*, 2004]. The fast azimuth of a split S-wave in such a stress-aligned anisotropic medium is parallel to the cracks that remain open, hence, parallel to maximum principal stress direction [Nur and Simmons, 1969, Boness and Zoback, 2006].

Plate interactions at the plate-boundary zone in South Island, New Zealand are expected to result in deformation induced seismic anisotropy. Westward movement of the Pacific plate with respect to the Australian plate results in an obliquely convergent boundary across New Zealand (Fig. 4.1). The main surface manifestation of this plate-boundary, the Alpine fault, extends for about 460 km, serving as an oblique transform fault that connects Puysegur and Hikurangi subduction zones with

opposite polarities (Fig. 4.1). The structural evolution of South Island is marked by different tectonic episodes. Some of the key processes include rifting and the breakup of Gondwana in the late Cretaceous, evolution of the two subduction systems and the plate-boundary, the transition of the southern subduction system (Puysegur subduction) into the central Alpine fault zone during the Eocene to Quaternary periods and southward migration of the Hikurangi subduction system increasing the transpressional character of the central collisional (Alpine fault) system [Sutherland *et al.*, 2000, Furlong, 2007, Cox and Sutherland, 2007]. Seismic anisotropy could remain from tectonic deformation during any of these episodes.

Figure 4.1 (a) Map of the tectonic setting of the South Island, New Zealand and the main structural features associated with the Australian-Pacific plate-boundary. Approximate location of the schist belt with high grade metamorphism is marked by the dark-gray shaded region. The black arrow denotes the relative plate motion direction (~ 38 mm/yr) between the Australian and Pacific plates [Walcott, 1998].

Figure 4.1 (b) Schematic diagrams (no vertical exaggeration) showing the proposed models of deformation for central South Island (dashed-line box in the inset of Figs. 4.1a). (i) Pervasive shear or thin viscous sheet model proposed by *Molnar et al.* [1999]. (ii) Localized deformation model. **LAB** - Lithosphere-Asthenosphere Boundary.

Present deformation along the plate-boundary can also influence anisotropy. In the northern part of South Island, the relative plate motion (~ 38 mm/yr) between the Australian and Pacific plates is partitioned into a plate-boundary/Alpine fault parallel component (37 mm/yr strike-slip movement) and a plate convergent component (11 mm/yr) orthogonal to the Alpine fault, resulting in significant transpression [*Norris et al.*, 1990, *Walcott*, 1998]. Northern and north-eastern parts of this region are dominated by crustal faults (Fig. 4.1) that indicate transition from the Hikurangi subduction zone to the central collisional zone. Continental collision in the central region is marked by structural features such as the Southern Alps mountain range, the Alpine Fault and a crustal root (~ 37 -42 km) and the thickened mantle lithosphere (~ 150 km) beneath the Southern Alps [*Scherwath et al.*, 2002, *Okaya et al.*, 2007]. In southern South Island, the Australian plate is being subducted obliquely underneath Fiordland, creating a relatively young subduction zone extending up to maximum depths of ~ 150 km [*Davey and Smith*, 1983].

There have been two opposing hypotheses to explain the deformation in the central South Island: the pervasive shear model and the narrow/localized deformation model (Fig. 4.1b). Structural and geodetic data suggest the partition of total plate

displacement to 480 km of right lateral slip on the Alpine fault and ~ 200 km wide distributed deformation to the east of the fault [Norris *et al.*, 1990]. Based on the consistent patterns of fast azimuths observed from SKS studies in South Island, Klosko *et al.* [1999] suggested the distributed deformation of right-lateral shear (~ 200 km) within a 40-400 km depth range. Molnar *et al.* [1999] and Stern *et al.* [2000] argue that this distributed deformation zone is restricted to the first 100 kilometers of the upper mantle and extends in a ~ 100 -150 km wide region either side of the Alpine fault. They proposed a thin viscous sheet model [Molnar *et al.*, 1999, Wilson *et al.*, 2004, Moore *et al.*, 2002] that agrees with the pervasive shear model [Bourne *et al.*, 1998] (Fig. 4.1b-*i*) to explain such deformation behavior in the presence of weak lower crust. Conversely, the localized deformation model (Fig. 4.1b-*ii*) assumes the presence of a weak upper mantle instead of a weak lower crust [Meade and Hager, 2005] and suggests that the strain is localized along the faults within the crustal and upper-mantle depths and is more distributed in the asthenosphere and/or deep mantle due to asthenospheric flow or shear [Avouac and Tapponnier, 1993, Devs *et al.*, 2011].

Here we focus on splitting of local S waves that originate within the lithosphere of the South Island to determine how anisotropy is distributed laterally and vertically in and around the plate-boundary zone across the South Island. We consider the possible sources of anisotropy and the possible mechanisms/processes associated with the formation of anisotropy. In comparison to previous local splitting studies (section 4.4.4) in the northern most region of the South Island [Audoine *et al.*, 2000, Balfour *et al.*, 2005], we use a larger data set, covering the entire South Island and the complete depth range available, and apply a 2-D delay-time tomography/spatial averaging technique to resolve anisotropy. We compare our results with the available SKS splitting [Klosko *et al.*, 1999, Savage *et al.*, 2007a], finite strain calculations from GPS data and the structural geology data [Little and Mortimer, 2001, Little *et al.*, 2002b] to examine how the upper lithospheric deformation varies from the

asthenospheric deformation and to test the deformation models proposed for South Island.

4.2 Data and method

We use both local (epicentral distance $\leq \sim 100$ km) and regional ($100 \text{ km} < \text{epicentral distance} < 250$ km) earthquake data recorded on twenty six GeoNet permanent three component seismographs for the period of 2004-2010 and an eleven month deployment of four temporary broad band stations in the northern part of South Island to measure S-wave splitting parameters (Fig. 4.2). These four temporary PASSCAL seismographs were deployed by Victoria University of Wellington in collaboration with the University of Colorado to enhance the seismic coverage. The complete region has been divided into three geographic units: northern, central and southern South Island. Events in each block are collected from the stations in that particular block (Fig. 4.2b). Stations that are close to the borders (WKZ, MSZ, ODZ, CASS and LTZ) of a block provided events recorded from either side. This criterion is used to control the epicentral distances (see 3.1.1). To obtain good signal-to-noise ratio, we use a threshold magnitude of 3.5 in northern and southern South Island. Due to the lack of seismicity and deep events in central South Island, the threshold magnitude of 3.0 is chosen to obtain more events. We utilize an automatic S-wave splitting (Mfast) technique [Wessel, 2010, Savage et al., 2010], which was developed on the basis of the eigen-value minimization method [Silver and Chan, 1991] and the cluster analysis method [Teanby et al., 2004b].

Figure 4.2 (a) and (b) maps showing the distribution of earthquakes that yield A and B quality splitting measurements in South Island (SI). The earthquakes are color coded according to earthquake depth. (a) displays events recorded from all possible incidence angles and (b) shows the events that are within the incidence angle (i_c) of 35° . Earthquakes within three blocks (northern SI, central SI and southern SI) are analyzed only using the station within that block, but the stations at the block boundaries (MSZ, WKZ, ODZ, CASS and LTZ) provided the events recorded on both sides. Squares and triangles display the temporary stations and permanent stations (GeoNet network) respectively.

Splitting of shear waves into two quasi S-phases upon propagating through an anisotropic medium and the directional dependence of velocities of these split phases are the principal phenomena associated with the S-wave splitting method. The delay time (δt) between fast and slow split components and the polarization azimuth of the fast split phase (ϕ) are the two main parameters that are determined during the splitting analysis. These parameters provide important information about the extent and degree of anisotropy and the orientation of the anisotropic medium and the past and/or present deformational processes [Silver and Chan, 1991, Crampin, 1994]. In an isotropic medium, the focal mechanism controls the S-wave particle motion, which is planar [e.g., Aki and Richards, 2002]. On entering an anisotropic medium, S waves obtain an elliptical particle motion (Fig. 4.3). Silver and Chan [1991] introduce an inverse splitting operator that removes the splitting from the S-waves to find the linearized particle motion. A grid search over all the possible pairs of fast azimuths (ranging from -90° to 90° with 1° increments) and delay times (ranging from 0 to 1 s at 0.01 s increments) is carried out. The values of ϕ and δt are those that give the most linear incoming particle motion, as measured by minimizing the smaller eigenvalue (λ_2) of the corrected two dimensional time-domain covariance matrix between the two horizontal components in the reference coordinate system (Fig. 4.3). Once the smaller eigenvalue is minimized, the corrected particle motion is linear and the eigenvector of the corresponding larger eigenvalue provides an approximate azimuth of the initial polarization or incoming polarization (ϕ_p) of the unsplit S-wave [Vidale, 1986]. This method uses the F-test to constrain the uncertainties of the splitting parameters by calculating the 95% confidence limits of the ϕ and δt that correspond to the optimum solutions from the grid search analysis.

Figure 4.3 (a) Outputs from the S-wave splitting analysis using Mfast [Savage *et al.*, 2010]. Top-left diagram shows the manual S pick (black line) on three components (e, n and z) of the seismograph. Dashed lines indicate the minimum starting time and maximum end time of the different S-wave analysis windows. Top-right figure displays the rotated horizontal components before correction for splitting (p and $p_{\perp}$) and after correction for splitting (corrected p and corrected $p_{\perp}$). Gray shaded regions in top figures indicate the selected S-wave analysis window. Dashed lines indicate minimum and maximum starting and ending windows. Two middle graphs show the splitting parameters (best is the blue cross) obtained from the *Teanby et al.* [2004b] cluster analysis. Bottom-left indicates the particle motion plots before and after correction for S-wave splitting. Here the gray region is the region not used in the analysis. Bottom-right plot is the contour diagram indicating the energy surfaces. The solution with the smallest second eigenvalue (λ_2) is marked by blue cross and the 95% confidence limit of this solution is denoted by the dark black line.

Figure 4.3 (b) Same as figure caption in Fig. 4.3(a), but with an A grade measurement with higher δt (see Figure 4.8).

Figure 4.3 (c) Same as figure caption in Fig. 4.3(a), but showing a B grade splitting measurement.

Before processing data with the automatic method, all the events are visually inspected and S arrivals are hand picked. The automatic method involves searching over a given set of 14 band pass filters, which are two-pole Butterworth filters with frequency band-widths within the range of 0.4 Hz to 10 Hz, and calculates

signal-to-noise ratio (SNR) to find filters that give the highest SNR-bandwidth product (≥ 3) for each earthquake-station pair. Then the signals obtained from the selected filters are used to estimate the splitting parameters. The dominant period of each S-phase is determined using the frequency that corresponds to the highest amplitude of the spectrum calculated from the three second window directly after the S-pick. Cluster analysis avoids uncertainties in the splitting parameters caused by the selection of S-wave analysis windows [Teanby *et al.*, 2004b]. This method selects the analysis window that gives the most stable splitting parameters with small errors by performing splitting measurements through a given set of analysis windows (Fig. 4.3). We chose the splitting measurements that agree with the A and/or B grading criteria (here after AB grade) as good quality splitting measurements [Savage *et al.*, 2010]. In Mfast, measurement attains AB grading if the measurement's $\text{SNR} > 3$, $\delta t < 0.8 * tlag_{max}$ ($tlag_{max}$ is the maximum δt for grid search), maximum error in $\phi < 25^\circ$, if it is not a null measurement, and if the measurement graded as a A or B cluster in the cluster grading (section 3.2.3). Note that the A grade criteria (a. $\text{SNR} > 4$, b. $\delta t < 0.8 * tlag_{max}$, c. not null and $d\phi < 10^\circ$) are a subset of AB grade criteria, which use more narrow limits for SNR and $d\phi$ compared to AB grade criteria. Null measurements are those for which ϕ_p is within 20° of parallel or perpendicular to the fast azimuth, which can lead to incorrect δt . The splitting measurements from each event that meet the above quality scheme are again checked considering their uncertainties and errors for different filters. This final quality criterion extracts the measurements that give a similar answer, i.e. if the time difference $< tlag_{max}/8$ or if the angular distance $< \pi/8$, in each filter (refers to 'best filter measurements' in later sections). From the ~ 1600 AB quality splitting measurements [Savage *et al.*, 2010] we selected the ~ 850 measurements that are within the incidence angle of 35° (see section 4.2.1) for further analysis (Fig. 4.2b). The number of good splitting measurements made at each station varies depending on the seismicity distribution (Fig. 4.2) and the quality criteria used in the analysis.

4.2.1 Limitations and statistical analysis of the measurements

The polarization of an S-wave incident at the free surface is governed by many factors such as the angle of incidence, frequency of the incident wave, the source depth, and the layering of the crust [Booth and Crampin, 1985]. To minimize the effect of incidence angle on S-wave polarization, splitting analysis is usually carried out inside the shear-wave window (i.e. a few degrees less than critical incidence angle) [Evans, 1984, Nuttli, 1961]. Wave fronts with incident angles beyond the critical angle ($i_c = \sin^{-1}(V_s/V_p)$, where V_s and V_p are S and P velocities respectively.) are likely to display complex particle motions due to phase conversions at the free surface. This will cause particle motion to be non-linear even in the absence of anisotropy [Nuttli, 1961] and make the interpretation of S-wave polarization complicated. To accurately calculate the incidence angle, we created a velocity model with average velocities that obtained from velocity-structure studies in South Island [Eberhart-Phillips and Bannister, 2010, Eberhart-Phillips and Reyners, 2001]. This average velocity model has an upper layer with V_s (S wave velocity) and V_p (P wave velocity) values of 2.6 km/s and 4.4 km/s respectively. The low velocity surface layer allows earthquake waves from a broad epicentral range to approach the surface within the S-wave window [Booth and Crampin, 1985]. In South Island, most of the seismicity is limited to the plate-boundary zone and this narrow distribution significantly reduces the number of events that can be used to measure the crustal anisotropy far away from the plate-boundary. Therefore, some of the stations in the eastern margin of the central and southern South Island (e.g. OPZ, ODZ, and SYZ) are only used for deep events (depth $\gtrsim 60$ km) (Fig. 4.4b).

In addition to calculating uncertainties and to grading individual splitting measurements, statistical parameters (e.g. mean, standard deviation, standard error) of the single station measurements are estimated (Table 4.1). We use directional statistics

Figure 4.4 Straight-line ray paths showing the coverage in the subduction zones in northern South Island (a) and southern South Island (b). Right side maps show the cross-sections (depth profiles) along AA' and BB' lines that are marked on the left side figures. Events and ray paths corresponding to each station are color coded using the same color. The approximate slab positions are colored in grey.

discussed in *Mardia* [1972] and *Gerst and Savage* [2004] for ϕ . Directional statistics are necessary to determine the statistics of ϕ and to estimate errors and accuracy of the mean ϕ [Roman *et al.*, 2011]. Because of the 180° ambiguity of angular data, directional statistics are different from normal statistics. Mean ϕ , the length of the mean resultant vector of ϕ (R), circular standard deviation (CSD) and standard error (SE) are the statistical parameters that we estimate. R is a quantitative estimate of circular spread that is calculated from the mean angular azimuth (ϕ_s). SE of ϕ_s is calculated using R and the concentration parameter (κ), which is a parameter of a Von Mises distribution ($VM_{(\phi_s, \kappa)}$), and SE indicates the accuracy of the ϕ_s . $VM_{(\phi_s, \kappa)}$

is a continuous distribution on the range of $0 - 360^\circ$ and it is a probability density function that varies with ϕ_s and κ . Thus, $VM_{(\phi_s, \kappa)}$ is considered to be a circular equivalent of the normal distribution that measures the circular spread [Mardia, 1972, Philipp, 2009]. Table 4.1(a) and Table 4.1(b) summarize the above statistical parameters that are calculated using splitting measurements made at each station and section 4.3 explains these results. We have also conducted a detailed analysis of splitting measurements considering the possible relationships between splitting parameters and other variables such as earthquake depth, earthquake-station back azimuth, average maximum frequency (or an average period) in the shear-wave analysis window and ϕ_p to provide constraints on both lateral distribution and depth extent of anisotropy (see section 4.4.1).

4.2.2 Resolving the spatial heterogeneity of splitting measurements

The heterogeneous anisotropic structure observed is further evaluated using 2-D delay-time tomography and a spatial averaging technique (TESSA) introduced by Johnson *et al.* [2011]. This allows us to obtain approximate constraints on regions with low and high anisotropy. A δt from a single measurement is a quantitative estimation of the anisotropic structure along the ray path from source to the station and the δt is taken to be proportional to path length (L) of the ray through the anisotropic medium [Silver, 1996]. TESSA is based on a simplified and linear relationship between δt and (L). It assumes that the δt_r observed along the ray path is the summation of δt calculated for each grid block (δt_b) with variable strength of anisotropy (s_b) in 2-D space. First, straight-line ray tracing between station and events determines the lengths of ray segments (L_{rb}) that lie on each grid block (Fig. 3.18). Then the total anisotropy measured at the station for a given event is

calculated by,

$$\delta t_r = \sum_{b=1}^n (s_b \times L_{rb}) \quad (4.1)$$

The inversion part of TESSA determines the s_b for resolved blocks and is carried out using the MATLAB medium-scale optimization inversion function that finds the linear least-square solution of the problem. This solution is then used iteratively to find the model solution for the strength of anisotropy of the whole ray (s_r) and this process is subjected to boundary conditions. As explained in *Johnson et al.* [2011], boundary constraints are applied in such a way that the minimum s_r will always be greater than 0 s/km and the maximum s_r will not exceed $\delta t_{r(max)}/L_b$, where $\delta t_{r(max)}$ is the maximum δt observed along a ray path and L_b is the width of the grid block. Tomography results discussed in section 4.4.3 are obtained using regular gridding with 25 km square blocks. The minimum number of rays per block is set up to be 6 to obtain reliable results. Tomographic analysis only uses the AB grade measurements that are within the incidence angle of 35° and that satisfy the best filter criteria. This enables us to avoid ambiguous splitting measurements as well as statistical uncertainties that could affect the final solution of the tomography and spatial averages.

We use the tomographic weighting scheme presented in TESSA to determine average ϕ in grid blocks. Tomographic weighting is designed to take into account variations due to heterogeneous anisotropic structure. The tomographic weighting function (ω_{rb}) that is applied in this averaging weights the ϕ considering the anisotropy strength profile determined for each grid block (s_b) in the δt tomography, so that more weighting is given to the regions with higher strength of anisotropy. ω_{rb} is defined as the strength profile of the each grid block, normalized by δt_r [*Johnson et al.*, 2011],

$$\omega_{rb} = s_b / \delta t_r \quad (4.2)$$

Then, the weighted mean ϕ of the grid block b with n number of rays is given by

azimuth ($\bar{\phi}_b$),

$$\bar{\phi}_b = \frac{1}{2} \arctan \left(\frac{\sum_{r=1}^n \cos(2\phi_r) \times \omega_{rb}}{\sum_{r=1}^n \sin(2\phi_r) \times \omega_{rb}} \right) \quad (4.3)$$

where b and r denote the block and ray number respectively. Both tomography and spatial averaging results are given in the section 4.3.

4.3 Results

Single-station splitting parameters are mainly discussed with regard to two depth units; shallow (≤ 100 km) and deep (>100 km). The 100 km depth contour roughly represents the minimum depth limit of the lithosphere-asthenosphere boundary (LAB) in South Island [Stern *et al.*, 2000, Scherwath *et al.*, 2002]. Figs. 4.5(a) and 4.5(b)/(ci/ii) show the normalized circular histograms of fast azimuths (ϕ) measured at each station using shallow and deep local earthquakes. These spatial maps represent ϕ of the high quality, A grade (section 4.2), measurements that were obtained from multiple filters (a maximum of three filters with $\text{SNR} \geq 3$ for each event). Events that have the same ϕ for different filters are visually more prominent than those with varying azimuths for multiple filters. Table 4.1(a) and Table 4.1(b) summarize the average splitting parameters obtained at each station using deep and shallow earthquakes respectively. To calculate single station averages (Table 4.1a/b), we use an A-grade measurements that satisfy the best filter criterion (section 4.2) for each station-earthquake pair.

Some stations (e.g. MSZ, NNZ, KHZ, CROE) exhibit either large scattering or bimodal distribution of ϕ (or high standard error) (Fig. 4.5a and Table 4.1a/b). We attribute these complex patterns of ϕ to spatially varying anisotropy and depth dependent ϕ as seen in the detailed analysis of splitting parameters (Fig. 4.6). In general, ϕ observed from both shallow and deep events indicate significant lateral

Table 4.1 (a) Summary statistics (see p105) of the splitting parameters from shallow (≤ 100 km) events.

station	no	ϕ_s^a	SE^b of ϕ_s	CSD^c of ϕ	R^d	mean δt	SE of δt	SD^e of δt
DSZ	49	60.526	11.701	48.148	0.244	0.172	0.016	0.115
MQZ	20	-26.623	14.977	44.727	0.296	0.180	0.019	0.086
KHZ	68	-49.314	16.878	56.321	0.145	0.196	0.015	0.127
LTZ	79	-72.066	5.310	38.273	0.410	0.188	0.017	0.148
NNZ	111	-69.658	8.798	50.158	0.216	0.197	0.013	0.138
QRZ	66	38.922	8.153	44.532	0.299	0.184	0.011	0.086
FOZ	15	-84.337	7.642	28.637	0.607	0.103	0.020	0.079
THZ	58	11.330	17.411	55.619	0.152	0.285	0.019	0.146
CRLZ	43	-34.399	3.838	25.109	0.681	0.147	0.010	0.066
WKZ	94	32.646	7.506	46.163	0.273	0.302	0.016	0.158
JCZ	61	-74.061	14.429	53.199	0.178	0.203	0.015	0.116
LBZ	79	38.882	9.187	48.098	0.244	0.206	0.015	0.129
DCZ	139	-9.234	3.246	34.081	0.493	0.159	0.008	0.089
MSZ	99	12.399	13.200	55.479	0.153	0.218	0.017	0.171
EAZ	95	8.267	8.333	48.010	0.246	0.222	0.012	0.114
MLZ	120	41.762	7.044	47.155	0.258	0.132	0.007	0.076
TUZ	86	-9.634	4.820	37.210	0.430	0.203	0.011	0.099
WHZ	83	-26.898	6.790	43.319	0.319	0.210	0.013	0.121
ODZ	38	-25.905	4.077	25.077	0.682	0.269	0.029	0.177
OXZ	15	-32.126	9.739	33.780	0.499	0.232	0.041	0.160
APZ	38	-83.567	6.099	33.713	0.500	0.263	0.020	0.124
OPZ	13	15.006	19.602	45.660	0.281	0.182	0.022	0.080
PYZ	24	22.666	16.154	47.578	0.252	0.229	0.025	0.121
SYZ	26	-9.093	11.187	41.845	0.344	0.108	0.010	0.053
CASS	6	66.843	10.572	25.730	0.668	0.139	0.027	0.066
CROE	51	-57.412	8.198	42.323	0.336	0.153	0.012	0.086
FREW	10	-75.721	22.599	45.851	0.278	0.209	0.046	0.145
KELY	107	8.748	6.321	44.303	0.302	0.176	0.011	0.114
RPZ	29	25.981	11.223	42.899	0.326	0.177	0.016	0.088

^amean fast azimuth^bstandard error^ccircular standard deviation^dthe length of the mean resultant vector ϕ ^estandard deviation

Table 4.1 (b) Summary statistics of the splitting parameters from deep (>100 km) events.

station	no	ϕ_s	SE of ϕ_s	CSD ϕ	R	δt	SE of δt	SD δt
DSZ	172	-89.818	4.740	43.407	0.317	0.190	0.008	0.101
KHZ	131	42.588	17.024	60.993	0.104	0.185	0.009	0.101
LTZ	91	-58.693	12.466	54.010	0.169	0.190	0.014	0.129
NNZ	137	22.979	20.417	63.669	0.085	0.181	0.009	0.105
QRZ	131	-89.716	5.395	43.286	0.319	0.239	0.010	0.118
THZ	103	7.699	5.271	40.653	0.365	0.185	0.012	0.122
WKZ	36	-8.316	12.335	46.449	0.269	0.259	0.027	0.161
JCZ	43	-60.998	2.761	18.326	0.815	0.219	0.009	0.059
LBZ	124	38.022	1.786	20.145	0.781	0.165	0.006	0.064
DCZ	91	-12.607	4.413	36.018	0.454	0.162	0.010	0.096
MSZ	84	-67.490	11.236	51.816	0.195	0.257	0.013	0.115
EAZ	53	-11.328	22.251	58.472	0.125	0.259	0.024	0.173
MLZ	84	59.590	9.536	49.211	0.229	0.193	0.013	0.119
TUZ	74	-41.503	5.457	38.171	0.412	0.206	0.018	0.154
WHZ	79	-9.397	11.475	51.666	0.197	0.178	0.015	0.129
ODZ	36	-20.020	8.517	39.788	0.381	0.228	0.026	0.156
OXZ	20	-34.412	20.972	50.346	0.213	0.308	0.027	0.120
APZ	23	-86.429	7.795	33.596	0.503	0.191	0.034	0.164
OPZ	9	86.165	64.184	60.834	0.105	0.182	0.028	0.084
PYZ	20	-19.624	11.095	39.224	0.392	0.199	0.030	0.136
SYZ	52	32.961	11.443	48.271	0.242	0.129	0.010	0.073
CASS	3	49.081	35.524	43.222	0.320	0.230	0.084	0.146
CROE	48	-24.622	44.491	66.860	0.066	0.208	0.020	0.135
FREW	23	21.890	18.091	49.093	0.230	0.247	0.037	0.178
KELY	83	13.813	11.706	52.357	0.188	0.209	0.013	0.117
RPZ	76	-19.418	5.329	37.968	0.416	0.245	0.017	0.147

variations compared to that of SKS azimuths (section 4.4.4), but measurements at some closely spaced stations tend to show regional consistencies over short distances suggesting coherent anisotropy in small regions (Fig. 4.5). Shallow events processed at stations in the northern South Island display two distinct fast polarization azimuths (Fig. 4.5b): sub-parallel to the Alpine fault and the Marlborough fault system, mainly in the western and central part of northern South Island (QRZ, KELY, DSZ and THZ); and approximate EW alignment at a high angle to the Alpine fault trace in the eastern side of the northern South Island (LTZ, CRLZ, OXZ and MQZ). ϕ also becomes approximately EW oriented from deeper earthquakes (>100 km) observed at coastal stations in northern South Island (DSZ, QRZ and

KHZ) and the western part of southern South Island (MSZ and JCZ). For shallow earthquakes, NE oriented (plate-boundary sub-parallel) ϕ are observed at most of the stations in northern South Island and the central region of central South Island and southern South Island (e.g. CASS, RPZ, LBZ, WKZ, MLZ). However, the stations that are very close to the Alpine fault in central South Island (JCZ and FOZ) show either bimodal distributions or WNW-ESE fast azimuths at high angle to the fault (Fig. 4.5). This nearly fault-perpendicular azimuth also appears to be the prominent azimuth at depths >100 km as seen in JCZ and MSZ (Figs. 4.5d and 4.6b), indicating a strong WNW-ESE alignment in anisotropy at subcrustal and/or upper mantle depths in the western side of southern South Island.

Figure 4.5 (a) The spatial distribution of ϕ (A grade) obtained from all events is plotted as circular histograms (rose diagrams) at the stations. (b) and (c) i/ii show the A grade ϕ measured from shallow and deep earthquakes respectively. Blue line on the Fig. 4.5(b) outlines the approximate limit of the plate-boundary parallel anisotropy, which we interpreted as the limit of pervasive simple shear. (d) Rose diagrams of ϕ from the stations on the Alpine fault. The consistent WNW-ESE alignment is significant at deeper levels (rose diagrams on the right side of dashed-line). Top-right number denotes the number of measurements at each depth level (bottom-right value)

4.4 Interpretations and discussion

4.4.1 Depth extent and backazimuthal variations of splitting parameters

An analysis of the splitting parameters with respect to depth and backazimuth is carried out in order to understand the complex patterns of ϕ observed from the rose diagrams. This analysis is limited to northern and southern South Island where there is a good earthquake distribution with depth due to subducting-slab earthquakes (Fig. 4.4). Stations that sample different regions of the two subduction zones with considerable numbers of splitting measurements exhibit changes in ϕ with depth (e.g. DSZ, QRZ in Figs. 4.6a and JCZ, MLZ in Figs. 4.6b). These depth variations sometimes appear to be combined with lateral variation, indicated by a complex patterns of ϕ (Fig. 4.6).

We also search for variation in delay times (δt) to determine the extent and the degree of the anisotropy. In the presence of homogeneous or slowly varying anisotropy, δt accumulates along the earthquake-station ray path as a result of the difference in travel velocities of two split shear waves [Rumpker and Silver, 1998]. At most of the stations, δt are within the range of ~ 0.15 - 0.30 s, but some stations recorded high δt (>0.4 s) with ϕ sub-parallel to the Alpine fault. The west coast close to the Alpine fault in southern South Island is also characterized by varying anisotropic structure as shown in Fig. 4.6b (JCZ, WKZ and EAZ), with inconsistent patterns of ϕ and high δt . Such variations imply a complex tectonic setting in the Fiordland region. However, some clusters of closely spaced earthquakes give consistent readings at each station (e.g. Figs. 4.6 and A.2).

Figure 4.6 (a) Changes in ϕ with depth in northern South Island along the profile AA' (see Fig. 4.4a). Station measurements displayed in the upper, middle and bottom diagrams are located in western, central and eastern regions of the northern South Island respectively. Depth boundary where there is a visible change in ϕ with depth is marked with a black dashed line (i.e DSZ and THZ) and the colors of the bars denote the individual δt measurements. Note that the ϕ are determined in the horizontal plane and are thus plotted relative to N as marked on the figures. The blue cross indicates the reference point of each profile that is used to calculate the relative distance to the station. The approximate position of the slab (gray line), possible depth range for the location of lithosphere-asthenosphere boundary (LAB - shaded gray region), and the location of the Alpine/Wairau (AF/Wr) and Murchison-basin (Mu) faults are shown in each profile. The direction of the black bar next to each fault label indicates the approximate strike direction of the fault in map view (not the dip).

Figure 4.6 (b) Changes in splitting parameters with depth in southern South Island along the profile BB' (see Fig. 4.4b). Note the higher δt and complex patterns of ϕ observed at the stations JCZ, WKZ and EAZ that are located in the northern region of the southern South Island close to the Alpine fault. Black dashed line (e.g. JCZ) indicates the depth boundary where there is a clear change in ϕ (colors of the bars denote the individual δt measurements).

ϕ perpendicular to the strike of the Alpine fault observed from the intermediate (50-100 km) to deep measurements (>100 km) that are made on some stations (e.g. JCZ, MSZ and FOZ) right above and close to the fault (Fig. 4.5d) could be explained by two alternative mechanisms. If the upper-most mantle (~ 30 km thick upper-mantle layer) contributes most of the anisotropy, it is likely a product of deformation of olivine crystals under different physical and chemical conditions. As proposed by *Karato et al.* [2008], deformational fabrics of olivine could be influenced by stress, temperature and the water content of the material resulting in changes of olivine slip directions from typical [100] to [010] (B-type olivine). The subduction to collision tectonic setting in the western edge of the central and southern South Island could be a favorable place for high pressure and/or water rich environment that could change the orientation of ϕ from the expected fault sub-parallel ϕ and result the ϕ to be orthogonal to the plate-boundary and the Alpine fault. The other possibility is that the deep high-frequency phases are sensitive to crustal anisotropic structures [*Audoine et al.*, 2000]. The prominent WNW-ESE direction at the above stations is parallel to the principal compressive direction ($\sim 115^\circ$) of this region [*Boese et al.*, 2012, *Townend et al.*, 2012], suggesting the contribution of crustal stress controlled mechanisms.

Apparent spatial variations of ϕ can also be caused by the geometry of the anisotropic unit. Models of vertically varying structure with two [*Silver and Savage*, 1994] or more layers [*Rumpker and Silver*, 1998] of anisotropy that have different azimuths of horizontal symmetry or have plunging symmetry axes [*Babuska et al.*, 1984, *Babuska and Plomerova*, 1993] yield complex but systematic patterns of splitting parameters (see section 4.4.2). However, modeling of these effects has been carried out mainly on SKS splitting studies focusing on the anisotropic structure of the upper mantle and models have limited applications for the high-frequency local S-wave splitting because of the crustal heterogeneities and scattering that also affect the splitting. Lack of strong correlation between splitting parameters and BAZ/incidence angle

(Fig. A.2) in our splitting data set suggests splitting parameters are not affected solely by single dipping layers of anisotropy.

Anisotropy in the two subduction systems

According to the ray geometry, seismic anisotropy sampled by stations LTZ and KHZ in northern South Island and MSZ and DCZ in southern South Island are most likely coming from the slab and sub-slab regions of the Hikurangi subduction zone (southern most part) and the Fiordland subduction zone respectively (Figs. 4.4 and 4.7). LTZ and DCZ show approximately the same azimuths for most of the slab events (Fig. 4.6 and 4.5-a). MSZ and DCZ indicate a significant increase in δt with depth (Fig. 4.7). Assuming 4.5 km/s average S-wave velocity for a single anisotropic layer, we have calculated an approximate percent velocity anisotropy [*Savage, 1999*] in the full depth range (~ 0 -240 km in Hikurangi subduction zone and ~ 0 -120 km Fiordland subduction zone) of the slab region (Fig. 4.7) using the highest quality splitting measurements (A grade, best filter and $\lambda_{max} > 8$; section 3.2.3). Most of the δt used for this analysis are less than 0.3 s, but a very few measurements are within the range of 0.3 - 0.5 s. These higher δt may indicate either a strong crustal contribution of anisotropy in the slab or the sub-slab contribution of anisotropy. Our calculations yield about average 0.14% ($\pm 0.09\%$) shear-wave velocity anisotropy for the Hikurangi subducting slab region (Fig. 4.7a-*ii*) and 0.40% ($\pm 0.135\%$) velocity anisotropy for the Fiordland subducting slab (Fig. 4.7b-*ii*). Slopes of the MSZ and the combined plots of MSZ and DCZ (Fig. 4.7b-*i/ii*) are statistically significant (table 4.2). Velocity anisotropy calculated for the Hikurangi subduction zone is within the error range of velocity anisotropy ($0.23\% \pm 0.34\%$) found by *Audoine et al. [2000]* for the northern most region of northern South Island. Yet, they both are small compared to higher shear-wave velocity anisotropy ($3.1\% \pm 2.2\%$) observed from 70 km deep zone in the Wellington region that is located further north in the Hikurangi subduction zone [*Matcham et al., 2000*]. The observed low velocity

anisotropy implies the re-splitting of S-phases or sensitivity of S-phases to the uppermost crustal anisotropy. The ray paths suggest that deep slab events (>70 km) (Fig. 4.7) in northern South Island are most likely sampling the sub-slab region and only shallow events recorded on these stations should be sensitive to shallow slab anisotropy. When we only consider shallow slab events (≤ 70 km), our calculation yields $1.2 \pm 0.63\%$ velocity anisotropy (Fig. A.3), within the error range of *Matcham et al.* [2000]. The anisotropy measured could be under estimated if the re-splitting has occurred (section 4.4.2 and 4.4.4), in which case the anisotropic region is smaller than the whole path.

Table 4.2 Regression coefficients and uncertainties of the Figure 4.7

Plot	Slope (s/km)	95% C.I. of slope (s/km)	Intercept (s)	95% C.I. of intercept (s)	Vel.aniso.% *
a(i) LTZ	0.0003	-0.0004, 0.0009	0.12**	0.0361, 0.2003	0.14(± 0.135)
a(ii) LTZ+KHZ	0.0003	-0.0001, 0.0007	0.12**	0.0654, 0.1646	0.14(± 0.090)
b(i) MSZ	0.0019**	0.0005, 0.0033	0.02	-0.1032, 0.1458	0.86(± 0.315)
b(ii) MSZ+DCZ	0.0009**	0.0003, 0.0016	0.09**	0.0311, 0.1410	0.40(± 0.135)

*Velocity anisotropy is determined from the slope assuming 4.5 km/s S-wave velocity. The error of percent velocity anisotropy is calculated from the standard error (SE) of the slope.

**statistically significant regression coefficients.

The steepness of the Fiordland subducting slab results in a narrow spatial extent of the subduction zone. Therefore, complexities due to anisotropy contributions from the mantle wedge and sub-slab zones are limited in this region. The velocity anisotropy calculated for the Fiordland subducting slab is higher than that calculated for southern most Hikurangi subducting slab (Fig. 4.7 and table 4.2). Shallow slab events (≤ 100 km) recorded on MSZ yield Alpine fault sub-parallel ϕ (Fig. 4.5b), suggesting fault induced anisotropy at crustal depths in Fiordland. Fairly low percent velocity anisotropy and ϕ that are parallel to structures (e.g. strike of the slab and faults) obtained for both subduction zones suggest that the anisotropy in the subducting slab could be overprinted by crustal anisotropy. The δt measured at MSZ and LTZ are also sensitive to the average period of the shear-wave (Fig. 4.9).

Figure 4.7 Anisotropy in the two subduction systems. Left and right hand plots show ray paths (a and b) and δt measured from events originating in Hikurangi and Fiordland subducting slabs respectively. Possible depth range at which the Moho and lithosphere-asthenosphere boundary (LAB) are located is shown by the gray shaded region. Here most of the ray paths sample the subducting slab region. Yellow (LTZ and MSZ) and blue (KHZ and DCZ) dots indicate δt and the black line shows the linear regression fit that is used to calculate the velocity anisotropy of the two subduction zones. Regression coefficient uncertainties are given in the table 4.2.

Therefore, the measured anisotropy from the slab events could be governed by more than a single source of anisotropy, such as crustal anisotropy or else the slab itself may consist of more than a single source of anisotropy (e.g. aligned micro-cracks,

bending-induced faulting and mineral alignment), that could result in re-splitting of S-phases or frequency dependent splitting.

4.4.2 Evidence for heterogeneous anisotropic structure

Scatter in δt and spatially varying ϕ imply complex anisotropic structure beneath South Island. Complex and/or multiple layered anisotropy can be investigated by studying changes in splitting parameters as a function of backazimuth (BAZ) and ϕ_p [Silver and Savage, 1994, Rumpker and Silver, 1998, Long and Silver, 2009] or with dominant period (T) of the S-wave. ϕ_p is the polarization azimuth of the unsplit shear-wave before entering the anisotropic medium. In the case of re-splitting, this polarization azimuth is equivalent to the azimuth of the fast shear-wave that resulted from the preceding anisotropic medium. At DSZ in the western edge of the northern South Island, δt varies with ϕ_p (Fig. 4.8b), and varies with depth (Fig. 4.8c), but other variables failed to yield strong correlation (e.g. backazimuth, magnitude, period of the S-wave).

Multiple-layered anisotropy with horizontal symmetry axes produce a characteristic pattern between ϕ_p and apparent splitting parameters with periodicity of $\pi/2$ [Silver and Savage, 1994]. However, the patterns are only detectable with long-period ($T/\delta t > 5$) waves [Rumpker and Silver, 1998] from a wide range of ϕ_p . $T/\delta t$ values obtained from the local and regional S-phases used in this analysis are within the range of ~ 0.3 -20. Therefore, seismograms that have long period waveforms ($\sim 11\%$ have $T/\delta t > 5$) are likely yielding apparent splitting, but those with smaller ratios of $T/\delta t \leq 1$ ($\sim 18\%$) probably measuring splitting parameters corresponding to the upper layer. We do not see a $\pi/2$ periodic pattern in splitting parameters with respect to ϕ_p as observed in long period waveform studies [Silver and Savage, 1994], but instead we see a jump in δt at about $\phi_p = 80^\circ$ (Fig. 4.8b). Higher δt tend to have ϕ_p between 20° - 80° (blue) and the lower δt have ϕ_p in the range of 80° - 170° (red). At depths $> \sim 160$

km both sets of ϕ_p approximately give the same E-W ϕ . But, at depths ≤ 160 km, initial polarizations between 20° - 80° yield N-S or NW-SE ϕ , whereas ϕ with 80° - 170° initial polarizations are mostly NE (see rose diagram of ϕ in Fig. 4.8c). Changes in both δt with ϕ_p and ϕ with depth may result from a localized subcrustal anisotropic zone. *Wannamaker et al.* [2009] found that the Murchison basin area fault zone and Marlborough faults are characterized by moderately low resistivity structures from ~ 25 - 50 km. They infer that these structures are produced by fluid expelled from the deep (~ 100 km) subduction interface and that they are dominated by fluid filled cracks and shear fabrics (Fig. 4.8e-*i*, *Wannamaker et al.* 2009), which should yield anisotropy parallel to the plate-boundary. Moderately high V_p/V_s areas identified via tomographic inversion of local earthquake travel-times by *Eberhart-Phillips and Bannister* [2010] in the same region also suggest high fluid pressure (Fig. 4.8e-*ii*). Northwest of the major fault systems, low V_p/V_s and high V_p [*Eberhart-Phillips and Bannister*, 2010] in the upper mantle is interpreted to be strong Australian lithosphere, which may not be deformed by present processes [*Eberhart-Phillips and Bannister*, 2010]. Thus, the ray paths passing through different zones will measure different ϕ .

We suggest that the waves from earthquakes deeper than 160 km, which travel on the paths mostly from the NW of DSZ and display mainly E-W and WWN-EES ϕ (Figs. 4.8c and 4.8d), are passing through an anisotropic zone with E-W and/or WWN-EES fast orientation. This zone could be either in the mantle lithosphere or crust; if it is in the mantle lithosphere it could be left from the Gondwana Cretaceous deformation [*Eberhart-Phillips and Bannister*, 2010] and if it is in the crust it could be caused by the present stress orientation [*Audoine et al.*, 2000]. A third possibility is that the E-W/WWN-EES alignment is a near source effect, because the earthquakes are within a region of high V_p/V_s ratios at the deep slab-edge (Fig. 4.8e-*ii*, *Eberhart-Phillips and Bannister* 2010). We favor the crustal stress interpretation as the ϕ is nearly parallel to the crustal stress direction (yellow bars in

Figure 4.8 Changes in δt with ϕ_p (b) and changes in ϕ with depth from events recorded at DSZ in Murcheson basin area (Fig. 4.1) in the western northern South Island. (a) map view of the events recorded. (c) depth profile of the ϕ along the line AA'. ϕ and the events are color coded according to ϕ_p and the dashed line indicates the change in ϕ with depth. Here the rose diagram shows the histogram of ϕ obtained from events ≤ 160 km and color coded according to ϕ_p . (d) schematic diagram illustrating the different anisotropic zones (P is a region with E-W aligned anisotropy, Q is a layer with NW-SE anisotropy and R shows anisotropic patches with NE-SW orientation), which result in the observed patterns of ϕ in Fig. 4.8(c). (e-i) 2D inversion model of electrical resistivity for northern South Island [Wannamaker et al., 2009]. The low resistivity body in Murcheson basin area and north of the Marlborough faults are marked by C and B respectively. (e-ii) Depth profile of the V_p/V_s model for central northern South Island [Eberhart-Phillips and Bannister, 2010]. Dashed line box shows the rough location of the electrical resistivity model explained in Fig. 4.8(e-i).

Fig. 4.11, *Townend et al.* 2012), and short period waveforms are often most effected by the uppermost path. The shallower ray paths are in the lithospheric mantle and the crust to the SE of station DSZ, and pass through the Marlborough fault system, which has patches (zone c in Fig 4.8e-*i*) of high fluid content. These rays pass through a region in the lower crust/upper mantle that has roughly NE/SW oriented fast anisotropy induced by plate-boundary parallel shearing (zone S in Fig. 4.8d), and then pass through another layer with roughly NW/SE fast anisotropy (zone Q in Fig. 4.8d), which re-splits them. This NW/SE anisotropy could be generated by asthenospheric counterflow perpendicular to the plate-boundary or by some other processes. Some waves continue to DSZ without traveling through any other regions. They have ϕ between N and NW, δt of about 0.3 s and NE/SW incoming polarizations obtained as they were traveling through the lowest layer as fast waves. Other waves pass through a third patch of NE/SW oriented ϕ (patchy area marked by R in Fig. 4.8d). They have NE/SW ϕ with shorter δt of ~ 0.1 s, and NW/SE incoming polarizations (Fig. 4.8b and c). Perhaps these shallow patches are in the fault zone or are regions of schist. Cycle skipping could also produce a dependence between δt and ϕ_p [*Walsh*, 2012]. However, the contour plots of the large δt with ~ 0.3 -4 s (Figs. 4.3b) suggest this is not occurring.

Other than some stations (e.g. MSZ) that sample slab anisotropy, values of δt show no clear increase with depth. Instead, there is a scatter in δt that is measured on stations in the central northern South Island (e.g. THZ, KELY) and a few stations (e.g. WHZ, APZ) in southern South Island (Fig. A.4). Both scatter in δt and changes in ϕ at nearby stations imply the presence of heterogeneous anisotropic structure. Regions where changes occur over short distances laterally and vertically are likely characterized by variable sources of anisotropy (e.g. shear-induced structures, micro-cracks, mineral alignment due to flow and large-scale structural fabric).

Figure 4.9 Examples of stations that show changes in δt with period of the S-waves in northern and southern South Island. Black line indicates the least squares fit of the data. Two plots with gray circles show the combined plots of individual station data in the southern and northern South Island. Here the correlation coefficient (CCEF) and 95 % confidence interval of the CCEF for each station are marked on the corresponding station plot. Note the moderately high CCEF (>0.4) calculated at the southern South Island stations.

Some studies suggest that different S-wave frequencies sample different scale-lengths of anisotropy, hence produce variable results depending on heterogeneity of the medium [Marson-Pidgeon and Savage, 1997, Wirth and Long, 2010]. We test this hypothesis by analyzing the highest quality splitting parameters (A grade, best filter and $\lambda_{max} > 8$; section 3.2.3) at each station (Fig. 4.9). The correlation coefficient (CCEF) between δt and period determined at northern South Island stations (KELY, LTZ and NNZ) are significant if they are combined (Fig. 4.9-top). In southern South Island, the individual CCEF values determined at stations MSZ, WHZ, APZ, and EAZ and CCEF of the pooled data set are significant (Fig. 4.9-bottom). These stations and EAZ station in South Island show moderately high CCEF between period and δt , i.e., frequency dependent splitting (e.g. CCEF > 0.4). The dependency of δt on the period of the S-wave may reflect a heterogeneous anisotropic structure in the Fiordland subduction zone, and SKS splitting can be explained by shallow anisotropy in this area. But the lack of frequency dependent splitting at the other stations suggests the large SKS δt with consistent ϕ [Klosko *et al.*, 1999] can not solely be explained by frequency dependence and therefore require a deep seated source.

4.4.3 Interpretations from 2-D delay-time tomography and spatial averages

The inconsistency of measurements from closely spaced earthquakes at different stations suggests that lateral variations in anisotropy occur. This in turn suggests that the 2-D delay-time tomographic approach might be useful to determine the average structure. Fig. 4.10 illustrates the 2-D delay-time tomography results obtained using AB quality splitting parameters calculated from 2615 event-station pairs. Because of the depth dependence of splitting parameters, we produce 2-D distribution maps for two depth sections (depth ≤ 100 km and depth > 100 km)

and these maps provide average spatial variations of δt . In Fig. 4.10 (shallow/deep), warm colors (yellow to red) indicate the regions with high strength of anisotropy ($\gtrsim 0.004$ s/km) compared to blue colors with weak strength of anisotropy ($\leq \sim 0.002$ s/km). The shaded regions Fig. 4.10 indicate the regions with low resolution and/or large residuals. Anisotropy strength profiles that are within the higher resolution region reasonably agree with the possible anisotropic sources in South Island. For depths less than 100 km, crustal fault zones and the areas with schist rock (compare to Fig. 4.1) appear to be the regions with strong anisotropy. This is clear from the high anisotropy strength yields around the Alpine/Wairau faults, Marlborough fault zone, Murchison basin area faults in the northern South Island (Fig. 4.1) and some crustal faults in the southern South Island (Fig. 4.10-shallow). For the deep earthquakes (>100 km), the region with high strength of anisotropy in the northern part of the southern South Island (SE of the Alpine fault) coincides quite well with the region that has large crustal thickness (~ 48 km) [Stern *et al.*, 2000, Bourguignon *et al.*, 2007a] and large negative Bouguer anomaly in and around the Southern Alps [Davey *et al.*, 2007]. Deformation associated with crustal shortening and mantle lithosphere (upper ~ 150 km) thickening as a result of continental collision could be the main mechanisms that governs the anisotropic structure in this region. As suggested by Norris *et al.* [1990], this region (SE of the Alpine fault) might be dominated by distributed deformation that results from the plate motion mechanisms.

Weighted mean ϕ obtained for each grid block are shown in Fig. 4.11. Due to lack of spatial resolution of deep measurements (Fig. 4.4), spatial averaging is limited to depths ≤ 100 km. Even at shallow depths of eastern-southern South Island, measurements are made in a limited region as a result of sparse distribution of ray paths and therefore, spatial averages are not well constrained at some locations (see Fig. 4.4 and Fig. 4.11). Spatial averages of ϕ obtained from shallow events display consistent regional patterns. ϕ in the red shaded regions are either sub-

parallel to the plate-boundary or aligned parallel to the translational faults (e.g. Marlborough fault zone) in northern South Island. The consistent pattern of ϕ with NS and NNW alignment in the region surrounding the red-shaded zone in southern South Island agree with the SKS measurements that were made in those regions (red bars in Fig. 4.11, *Savage et al.* 2007a). Thus, it is possible that most of the anisotropy measured in this region is coming from the mantle lithosphere or from coupling between the crustal and the mantle-lithospheric anisotropy. But, the NE-SW alignment in the southern red-shaded zone shows a sharp change in ϕ from the general trend in this region and is aligned parallel to the P_n anisotropy fast azimuth (blue bars in Fig. 4.11, *Scherwath et al.* 2002), the plate-boundary and the strike of the high topography (e.g. Southern Alps) of the region. This zone suggests an abrupt changes in strain field and is characterized by relatively high strength of anisotropy as seen in Fig. 4.10 (shallow).

Spatial averages in the northern part of the central South Island are poorly constrained due to lack of measurements. Standard error associated with most of the spatial averages in eastern and central part of the central South Island are greater than 25° , indicating more than a single mode of ϕ . Although the prominent trend (WNW-ESE aligned gray bars in Fig. 4.11) is consistent with the horizontal maximum compressive direction (SH_{max}) of this region (yellow bars in Fig. 4.11, *Townend et al.* 2012), spatial averages do not provide robust estimations due to large errors. Single-station measurements at RPZ, LBZ and CASS (Figs. 4.5(b)) show that there is a plate-boundary sub-parallel trend in the central region, however this trend is masked by the WNW-ESE ϕ parallel to SH_{max} when considering spatial averages. Therefore, the contour line connecting the two red-shaded regions is determined comparing both Figs. 4.5(b) and Fig. 4.11 (see Fig. 4.11).

Figure 4.10 2-D delay-time tomography from TESSA [Johnson *et al.*, 2011]. Tomography maps are created for shallow (≤ 100 km) and deep (> 100 km) levels. Warm colors (red to yellowish green) in the maps indicate region with higher strength of anisotropy compared to weak zone (blue color) and the shaded region shows the regions with low resolution and/or large residuals.

Figure 4.11 Spatially (25 km grid blocks) averaged fast azimuths ($\bar{\phi}_b$, equation 4.3) using TESSA. Map displays the spatial averages from shallow (≤ 100 km) events. Black color bars denote the $\bar{\phi}_b$ with standard error (SE) $\leq \pm 25^\circ$ and the gray bars indicate the SE of $\bar{\phi}_b > \pm 25^\circ$. Regions with plate-boundary sub-parallel $\bar{\phi}_b$ are denoted by red shaded regions (note that the sharp change in ϕ from NW-SE trend to NE-SW trend in southern South Island is marked by the shaded red region in the south). Blue dashed line marks the boundary limits of the NE-SW trend that is aligned with strike of plate-boundary and the associated faults. Both individual station measurements (Fig. 4.5b) and spatial averages have been used to constrained these boundary limits. SKS [Savage *et al.*, 2007a, Klosko *et al.*, 1999], P_n anisotropy [Bourguignon *et al.*, 2007a] and SH_{max} [Townend *et al.*, 2012] directions from previous studies are marked by red, blue and yellow bars respectively. White irregular lines indicate the fault structures in South Island. The question mark denotes the possible boundary limit of pervasive shear in the central South Island as noted by station ϕ .

4.4.4 Comparison with previous anisotropy studies and implications for plate-boundary deformation

Anisotropy in the crust and the lithospheric mantle in the Marlborough fault zone and the upper most region of the northern South Island was previously investigated by *Balfour et al.* [2005] and *Audoine et al.* [2000], but the spatial and the depth resolution of these studies are limited. *Balfour et al.* [2005] argued that the anisotropy in Marlborough is not controlled by the regional stress and both *Balfour et al.* [2005] and *Audoine et al.* [2000] found that ϕ in those regions are roughly parallel to the plate-boundary (NE-SW). In agreement with the above studies, we find a similar average ϕ (spatial) for the same regions in northern South Island (Fig. 4.11) from shallow events. Most of our deep events Figs. 4.6a yield the E-W/WNW-ESE trend presented by *Audoine et al.* [2000] from deep events at some stations. As suggested in section 4.4.1, this E-W/WNW-ESE trend may result from either a deep anisotropic source (e.g. LPO fabrics in the upper mantle or lower crust) or a shallow stress-controlled source or the combination of both. The NE-SW trend is likely caused by pervasive shear associated with the plate-boundary deformation (Fig. 4.11).

As highlighted in New Zealand's terrane map (Fig. 1.4(b)), the basement terranes of South Island are aligned with the orientation of the ancient Gondwana margin (Fig. 1.3), however close to the plate-boundary these terranes have been bent and shifted by the present-day oblique strike-slip movement of the Alpine fault. In general most of the basement terrane rocks are weakly metamorphosed, but some Gneissic fabrics that indicate high-grade metamorphism is common in some western province rocks [*Cox and Sutherland*, 2007]. Overprinted metamorphic fabrics on the terranes adjacent to the plate-boundary (e.g. some parts of Rakaia terrane), increase in metamorphic grade and the bending of terranes towards the boundary in the eastern side of the Alpine fault reflect the increased deformation and change

in state-of-deformation due to current plate-boundary activity. As explained in section 1.4, overprinted Haast-schist terrane consists of three distinct schist units (Otago, Alpine, and Marlborough Schists) with different deformation histories and structural orientations. Fabrics within those schist units and/or terranes manifest different deformation histories. In some regions of South Island, the individual station splittings, delay-time tomography and spatial averages of ϕ (e.g. Figs. 4.5, 4.10, and 4.11) correlate well with the above structural orientations and fabrics. Southeast of the South Island, about 60 km away from the Alpine fault, terrane boundaries and structures have a prominent NW-SE alignment, whereas close to the fault terranes have been truncated by the Alpine fault, resulting in bending and changing structural and terrane orientations to N-S, NNE-NNW and/or NE-SW (Fig. 1.4(b); *Cox and Sutherland* [2007]). Prominent NW-SE ϕ alignment in the southeast of central and southern South Island is in agreement with the structural trend in the southeast South Island (Fig. 4.5). The high strain zone with an abrupt change in ϕ from NW-SE to NE-SW in Fig. 4.11 coincides well with the region of dextral-bending of terranes occurs (Fig. 1.4(b)). The high strength of anisotropy detected from the delay-time tomography in the same region (around lat:-45°, long:168.2° in Fig. 4.10) could also be attributed to significant straining in rock fabrics and heterogeneous anisotropic structure associated with the oroclinal bending associated with Cenozoic deformation.

When compared to SKS and P_n anisotropy studies in South Island [*Scherwath et al.*, 2002, *Klosko et al.*, 1999], ϕ obtained from local/regional data are variable. But, the spatially averaged ϕ from earthquakes with depths ≤ 100 km (Fig. 4.11) are mostly consistent with the SKS and P_n fast azimuths observed for the southern central South Island (regions where WKZ and LBZ are located), the eastern part of the southern South Island and northern South Island (compare red and black bars in Fig. 4.11). Our observations are also consistent with the crustal finite strain estimations from bending related geological markers in South Island [*Little et al.*,

2002b]. As suggested by *Little et al.* [2002b], the consistency indicates coupling of the upper mantle deformation (as inferred from SKS results) with crustal deformation more towards the Alpine fault and weak coupling in the eastern part of the central and southern South Island (Fig. 4.5). Based on the ϕ from shallow events (≤ 100 km) (Figs. 4.5b and 4.11), we infer that the width of the plate-boundary sub-parallel ϕ that indicates pervasive simple shear [*Molnar et al.*, 1999, *Little et al.*, 2002b] is limited to about 130 km southeast from the Alpine fault in the southern part of central South Island (regions where WKZ and LBZ are located) and the northern region of southern South Island. A region with high geodetic shear strain rates in South Island as proposed by *Beavan et al.* [1999] also lies within the estimated width of the plate-boundary induced-deformation in our study. The western boundary limit of the near offshore region of northern and central South Island are not yet known. Also, the lack of deep events in the central South Island does not allow accurate constraints on the depth extent of this trend.

As proposed by *Molnar et al.* [1999], if the total strain produced by the pervasive shear is confined to the top 100 km of the upper mantle, a large portion of SKS δt anisotropy should be accommodated in the lithosphere. But for local/regional events, even though the ray paths travel through the same region of the upper mantle, the observed station δt average of local/regional splitting measurements is about 9-20 % of that observed by SKS studies. Thus, small average station δt ($\sim 0.1 - 0.4$ s) suggest either: (i) the majority of SKS δt are from asthenosphere and/or deeper part of the lithosphere or (ii) SKS and local phases sample the same structure, but the latter have higher frequencies and are seeing smaller δt due to frequency dependent splitting (Fig.4.9 and section 4.4.2) or (iii) higher frequency phases are more susceptible to re-splitting from small scale structures that are present at the last layer of a layered anisotropic media (e.g. Fig 4.8). Possibility (ii) appears to be a localized effect (section 4.4.2) and therefore either (i) and/or (iii) can be used to explain the δt discrepancies between local (this study) and SKS (previous work)

splitting.

4.5 Conclusions

Using both local and regional earthquake data, we have obtained a S-wave splitting data set to provide constraints on the lateral and vertical distribution of seismic anisotropy in the lithosphere of the South Island. The application of quality criteria and statistical analysis allow us to extract well resolved splitting parameters and to provide more confident results from the complex splitting associated with local/regional S-wave splitting analysis. Therefore, the spatial and depth variations observed from the detailed splitting analysis and results from the tomography provide a good approximation for the spatial distribution of anisotropy (sections 4.4.1, 4.4.2, and 4.4.3). In general, splitting measurements in South Island show regional consistencies over short distances, which imply localized coherent anisotropy. Spatial and depth variations in splitting parameters suggest the existence of several deformation domains (e.g. Fig. 4.11 and Fig. 4.8d). Anisotropy in these domains appears to be controlled by different source mechanisms: shear induced fabrics related to present-day plate-boundary deformation (in northern and central-southern South Island), stress controlled anisotropy linked with prevailing stress (in eastern part of the central South Island) and anisotropy deeper than 50 km due to the mantle anisotropy within the lithosphere and/or asthenosphere that reflects the paleo-anisotropy from previous tectonic episodes or present asthenospheric flow (in southern and eastern parts of the Southern South Island).

2-D delay-time tomography and spatial averaging of ϕ give consistent patterns on a regional scale and suggest that the schist rock may also have a significant contribution to the total splitting ($\sim 0.1 - 0.4$ s) measured. A NE-SW trend of the spatial averages of northern South Island consistent with the strike of the plate-boundary and the strikes of the crustal faults, suggests that the dominant source of

anisotropy is plate-boundary induced structures. An abrupt variation in ϕ from NW-SE to NE-SW in southern South Island (Fig. 4.11) suggests a high strain structure at crustal or subcrustal depths. This zone also indicates fairly high strength of anisotropy (~ 0.006 s/km) and resides in the region with high crustal thickness (~ 48 km) and inferred thick lithosphere [*Bourguignon et al.*, 2007a, *Stern et al.*, 2000]. The origin of this anisotropic structure may be related to pervasive shear from plate-boundary deformation.

Assuming that the plate-boundary sub-parallel ϕ is a product of distributed shear associated with plate-boundary deformation, we suggest that the zone of shear deformation extends ~ 130 km from the Alpine fault in the northern part of the southern South Island (eastern boundary limit of the red shaded region in South), but in northern and central South Island this zone is more distributed than in southern South Island. Unless there is re-splitting due to misaligned anisotropic media, and/or frequency dependent splitting affecting the local/regional splitting, small δt observed from local S-wave splitting compared to large δt (~ 1.7 s) observed from SKS splitting suggest that most of splitting from SKS is accommodated either in the asthenosphere or in the deeper part of thickened mantle lithosphere (depths $\sim 150 - 250$ km) where we do not have enough depth coverage of splitting measurements. When considering upper lithospheric depths (~ 80 -120 km thick lithosphere) in the central and upper-most southern South Island, our splitting results fit best with the localized deformation model (Fig. 4.1b).

Chapter 5

Shear-wave splitting anisotropy in the offshore plate-boundary zone, South Island

This chapter is in the initial stage of manuscript preparation and this may contain some repetitive sections from the Introduction and Methodology chapters (1 and 3).

5.1 Introduction

The study of internal deformation at plate-boundaries enables us to understand how lithospheric blocks interact with each other. An oblique-convergent boundary between the Australian-Pacific (AUS-PAC) plates has evolved since Eocene (~ 40 Ma) in New Zealand (NZ), changing its tectonic character [*Cox and Sutherland, 2007*]. The plate-boundary was initiated as a passive margin and it was later transformed to a large strike-slip fault (Alpine fault) [*Sutherland, 1999*]. The Alpine Fault is the main structure in the AUS-PAC boundary in SI, NZ. The fault has accumulated an offset of 450 km over the last 25 My. During this time the fault movement has changed from pure strike-slip to oblique strike-slip in response to the changes in plate-boundary orientation. Consequently, the present-day plate-boundary configuration in the SI is characterized by the transpressional Alpine fault that interlinks the Hikurangi and Puysegur subduction zones (Fig. 5.1a). The large-scale strike-slip faulting, transpressional behavior and relatively young tectonic history are some of the key features that make this plate-boundary an exceptional place to explore lithospheric deformation associated with continental margins.

5.1.1 Constraining plate-boundary deformation via shear-wave splitting

Lattice preferred orientation (LPO) of minerals creates an anisotropic medium with directional variations of its physical properties. This has been considered as the main source of anisotropy in the upper mantle, asthenosphere and perhaps in the lower-crust [Savage, 1999]. Anisotropy in the upper-mid crust is largely due to stress controlled sources such as aligned micro-cracks, fluid-filled cracks; structural sources like macro-faults and the alignment of bedding planes and foliations [Boness and Zoback, 2004]. Anisotropy is an excellent indicator of strain (or deformation) distribution at the plate-boundaries [Skemer *et al.*, 2012] and it helps to delineate deformation kinematics to understand the rheology of the system [Karato *et al.*, 1989].

Vaucher and Nicolas [1991] hypothesized that large-scale displacement in plate-boundary faults is likely associated with deformation in the whole lithosphere, and is characterized by broadly distributed shear deformation on either sides of the fault. Depth distribution of deformation, however, depends on the mechanical decoupling between layers [Flesch *et al.*, 2005]. Shearing of intrinsically anisotropic minerals (e.g. olivine, pyroxene) in the lithosphere and asthenosphere in response to the fault displacement results in LPO anisotropy with horizontal mineral alignment on vertical flow planes (or steep foliations) that are parallel to the plate-boundary [Nicolas and Christensen, 1987]. The width of this LPO anisotropy zone is, therefore, a measure of distributed shear deformation associated with the transcurrent fault movement. As suggested by Otsuki [1978], 500 to 1000 km of fault displacement can produce deformation of a few kilometers to ~ 40 km width in the lithospheric mantle and mid lower-crust. If this deformation is partitioned into several major shear zones in the mid-lower crustal depths, such zones will have a single large-scale shear zone in the mantle lithosphere and, therefore, distributed deformation ($\sim > 100$

km) around the plate-boundary [Nicolas, 1993].

Previous studies of P_n anisotropy [Scherwath *et al.*, 2002, Baldock and Stern, 2005], P-wave delays [Molnar *et al.*, 1999, Stern *et al.*, 2000] and SKS splitting [Klosko *et al.*, 1999] suggest that the mantle shear zone that is associated with the Alpine fault is distributed in a ~ 200 -300 km broad area around the plate-boundary and may extend into offshore regions of the SI. However, geodetic estimations and finite-element modeling suggest that the width of deformation is distributed only about 80-100 km zone from the fault [Beavan *et al.*, 1999, Ellis *et al.*, 2006, Moore *et al.*, 2002]. Finite strain estimations from geological markers [Little *et al.*, 2002b] also found more strain accumulation towards the plate-boundary (east to west), indicating crust-mantle coupling towards the fault. Because of the lack of depth resolution of the SKS splitting, such a study would not allow constraints on the lithospheric anisotropy using SKS phases. For this reason, Karalliyadda and Savage [2013] carried out a local shear-wave (S-wave) splitting investigation of on-land SI to examine lithospheric anisotropy. Despite the complex nature of splitting from high-frequency local S-waves, the study found that the mantle shear zone due to plate-boundary deformation is distributed within ~ 130 km from the boundary in the central SI and may be more distributed in the northern SI than in the central SI. The study also noticed that high-frequency local S-phases are more sensitive to crustal stress in some regions in SI.

Klosko *et al.* [1999] found little anisotropy at Chatham Island station in the eastern off-shore SI compared to 1.7 s SKS δt onland (Fig. 5.1-a). Based on the limit of low P_n velocity ($\sim 7.95 \pm 0.05$ km/s) across the central SI, Baldock and Stern [2005] suggested the width of distributed shear is less than 235 km east of the plate-boundary. However, P_n anisotropy has a limited depth constraint (i.e. just below Moho depth). Comparing the above findings with petrophysical analysis, Duclos *et al.* [2005] interpreted this anisotropic zone as an asymmetric body that fans out of the plate-boundary, so that the upper-limit of this body is shallower in the western

side in agreement with *Scherwath et al.* [2002] and deeper in the eastern side to explain 0% P_n anisotropy in the central-east offshore SI [*Baldock and Stern*, 2005]. Both local S-wave splitting delays (maximum ~ 0.3 s station average and 0.6-0.7 s individual delay times) [*Karalliyadda and Savage*, 2013] and petrophysical results [*Duclos et al.*, 2005, *Godfrey et al.*, 2000] have suggested that the depth extent of this anisotropic body is not only limited to the shallowest upper-mantle as argued by *Molnar et al.* [1999], but extends into the deep upper-mantle, possibly down to the asthenosphere. This has also been suggested by the recent geodetic studies by *Houlié and Stern* [2012], at least in the southern-central and southern SI.

We investigate seismic anisotropy in the previously unexplored offshore regions of SI using local/regional S-wave splitting. S-wave birefringence (i.e. splitting of S-wave into two quasi S-phases) is a characteristic property of S-waves propagating in an anisotropic medium. Splitting occurs in such a way that the polarization direction of the faster S-wave (fast direction- ϕ) is governed by the alignment of the anisotropic medium and the slower S-wave is orthogonal to the polarization direction of the fast S-wave. The time separation (delay time- δt) acquired by the two quasi S-phases upon propagating in the anisotropic medium gives information on the strength and spatial extent of anisotropy. We estimate these S-wave splitting parameters (i.e. ϕ and δt) from regional and local events recorded on the ocean-bottom seismographs to constrain the spatial distribution of splitting parameters. We also aim to address the question of whether the local/regional splitting in this region is entirely governed by the plate-boundary deformation or by other means.

Figure 5.1 (a) Offshore study region of the SI and circular histograms of fast azimuth plotted at the OBS station locations (Fig. (b)). Bold and thin arrows indicate the absolute and relative plate motion directions, respectively. The green bar on the Alpine fault denotes the approximate SH_{max} direction in the central SI ($\sim 115^\circ$; Boese *et al.* [2012]). (b) Ray path distribution and event locations of the measurements that satisfy the best filter and highest grade quality criteria (see section 5.2.1). Individual splitting measurements (gray bars) are plotted at one-quarter of the epicentral distance that is measured from the station. Red bars indicate the average station δt and ϕ azimuths (FA). Both red and gray bars are scaled by δt .

5.2 Splitting parameters from OBS data

5.2.1 Data and method

Twenty eight broadband Ocean Bottom Seismographs (OBS) that were deployed in eastern and western offshore regions of the SI with the MOANA (Marin Observation of Anisotropy Near Aotearoa) experiment [Collins *et al.*, 2010, Yang *et al.*, 2012] were used to retrieve S-phases from local and regional events. OBS on the western side were distributed in both the shallow (challenger Plateau) and deep ocean (Tasman sea). Eastern side OBS were located on the Campbell Plateau and Chatham Rise (Fig. 5.1-a). Seismicity in the area was mostly distributed outside of the OBS network, roughly centering between the western and eastern OBS deployments (open circles in Fig. 5.1-b show the events that yield good quality splitting measurements). We used the GeoNet catalog for the period from February 2009 - January 2010 to determine events and to extract the events from OBS records. Waveforms recorded on OBS stations were corrected for the OBS rotations using the rotation angles estimations by Stachnik *et al.* [2012]. We examined S-wave splitting on ~ 2000 S-phases from events with magnitude ≥ 3.0 in the eastern and ≥ 3.5 in the western side, respectively. Hand picked S-phases were analyzed for splitting with the automated S-wave splitting technique of Mfast [Wessel, 2010, Savage *et al.*, 2010]. Mfast uses the eigenvalue minimization and cluster analysis methods introduced by Silver and Chan [1991] and Teanby *et al.* [2004b], respectively. S-waves were subjected to filtering using a range of two-pole Butterworth bandpass filters (frequency band-widths between 0.4 Hz and 10 Hz frequencies). Filters were selected so that the S-wave analysis window had the highest signal-to-noise (SNR) ratio bandwidth product. We eliminated poor quality events and/or measurements in three steps: (i) events with ambiguous and/or distorted S-wave impulses were avoided prior to splitting analysis, (ii) signals with low SNR (≤ 2) were avoided during the analysis and, (iii) poor quality measurements were eliminated by either visual inspection of splitting

parameters on diagnostic plots (e.g. particle motion plots, contour diagrams and wave-form plots) or applying the highest quality grading criteria that comes with the Mfast program (section 3.2.3). We noticed that waveforms and/or splitting measurements of the farthest stations (NZ05, NZ06, and NZ 26) do not satisfy the above quality criteria and, thus, those stations do not yield any acceptable splitting measurements.

Volti et al. [2005] emphasized that the S-wave window (SWW) used for splitting in the sea-floor can be deviated from the limits of usual SWW (incidence angle $\leq 35^\circ$) that applies in the onshore splitting analysis. As a result of the low-velocity unconsolidated sediment layers in the sea floor [*Tsuji et al.*, 2011], high-frequency local and regional S-waves attain almost vertical incidence as they reach the sea-floor [*Volti et al.*, 2005] and, therefore, splitting measurements are unaffected by the distorted S-arrivals owing to phase conversions [*Booth and Crampin*, 1985, *Nuttli*, 1961] at the sea-floor. With careful inspection of splitting measurements from S-waves that have both large ($>35^\circ$) and small ($\leq 35^\circ$) incidence, we adapted the same approach to extract an unambiguous set of splitting parameters that have been determined from the earthquakes with a large range of epicentral distances. Long rays may attain horizontal propagation in some parts of the ray path, depending on the ray geometry. Such a horizontal propagation (especially shallow ray paths) could be sensitive to transversely isotropic media with vertical symmetry axis (e.g. horizontal foliations and sedimentary bedding). These types of media most likely to occur in the eastern offshore, where there is an oceanic-plateau with shallow bathymetry, thick continental crust and less tectonic activity [*Davy and Uruski*, 2002]. As found by *Paulssen* [2004], the fast-split S-wave of a wave with horizontal propagation is usually horizontally polarized in the transversely isotropic media and thus, the horizontal projections of ray path and ϕ are orthogonal to each other.

5.2.2 Results and interpretations

About 1540 measurements were graded as AB quality, but the spatial variations of splitting parameters were evaluated from 1300 measurements, which satisfied the high quality grading criteria (table 5.1). The selected measurements were non-null measurements with relatively dense error contours ($\lambda_{max} > 5$) and small splitting errors (see grading in section 3.2.3). In Fig. 5.1-a, fast azimuths (ϕ) measured at stations in the western and eastern offshore regions are represented as rose diagrams. We allowed the program to calculate at least three splitting measurements for each event to examine the sensitivity of splitting parameters to different filter bandwidths. Individual rose diagrams are weighted by the number of measurements with that particular azimuth and therefore, the longer the rose segment the higher the number of measurements in that direction. The rose diagrams are not normalized according to the number of measurements at each station, but information can be found in tables 5.2 and 5.1. Because of the configuration of the OBS network and earthquake distribution, splitting measurements on western and eastern OBS mainly sample the anisotropy in the Australian and Pacific lithosphere respectively (Fig. 5.1-a).

Table 5.1 The number of measurements from different selection criteria.

criteria*	no.measurements
AB grade/all**	2076
AB grade/ $\lambda_{max} > 5 / \delta t \leq 0.7$ s	1300
AB grade/best-filter/ $\delta t \leq 0.7$ s	1390
AB grade/best-filter/ $\lambda_{max} > 5 / \delta t \leq 0.7$ s	547

**measurements from all filters with SNR > 3.

*see appendix B.2 for more information on the selection criteria.

We extracted a single pair of splitting parameters for each event by using the best filter criterion. The best filter criterion compares the splitting measurements from different filters for each event and chooses a single set of splitting parameters if the splitting measurements are similar for all different filters [Savage *et al.*, 2010]. Figures 5.1, 5.2-A/B and table 5.2 show these best filter splitting measurements and their station averages. In general, individual splitting delay times (δt) from both sides range from $\sim 0.02 - 0.5$ s (Fig. 5.2-B) with an average value of 0.25 s. ϕ are not spatially consistent throughout the whole region, but patches of spatial consistencies can be seen in the western offshore with a rapid change in ϕ from plate-boundary parallel in the far west to plate-boundary perpendicular at about 150 km from the plate-boundary (Fig. 5.1a/b). Section 5.2.2 discusses the splitting estimations and gives possible interpretations for the observed patterns in the western and eastern offshore regions.

Figure 5.2 ϕ (bottom-B) and δt (top-A) variations with distance from the Alpine fault (gray open circles and open boxes). Mean splitting parameters at each station are marked by the black circles. Black triangles denote the average δt and ϕ estimated at the inland stations [Karalliyadda and Savage, 2013], which are located in the central and northern SI. Error bars indicate the standard errors of the mean splitting parameters at each station.

Table 5.2 Measurements and statistics of offshore splitting measurements shown in figure 5.2.

station	ST-AF (km)*	no	$\bar{\phi}$	$SE_{\bar{\phi}}$	$CS D_{\bar{\phi}}$	$error_{\phi}^{**}$	R	δt	$SE_{\delta t}$	$SD_{\delta t}$	$error_{\delta t}^{**}$
NZ02	116.79	81	16	8	46	2.6	0.283	0.18	0.02	0.16	0.004
NZ03	224.52	42	54	16	52	3.9	0.195	0.23	0.03	0.18	0.007
NZ04	338.50	21	19	14	44	1.9	0.313	0.15	0.02	0.08	0.003
NZ07	387.18	2	48	4	5	0.1	0.985	0.38	0.02	0.03	0.005
NZ08	296.54	6	8	17	36	1.4	0.464	0.22	0.09	0.21	0.004
NZ09	263.04	4	78	4	8	0.7	0.961	0.26	0.02	0.03	0.007
NZ10	111.47	97	-4	7	45	2.9	0.290	0.15	0.01	0.11	0.004
NZ11	65.34	79	-9	6	39	2.0	0.389	0.17	0.01	0.11	0.004
NZ12	61.13	15	-4	21	48	2.8	0.248	0.23	0.04	0.18	0.013
NZ13	134.98	47	-64	10	45	2.7	0.285	0.17	0.02	0.14	0.003
NZ14	77.28	14	-56	19	46	1.3	0.273	0.23	0.04	0.17	0.005
NZ15	101.86	8	-42	18	40	2.5	0.376	0.16	0.03	0.10	0.006
NZ16	192.30	11	54	35	54	3.1	0.174	0.10	0.03	0.09	0.003
NZ18	226.76	15	25	23	49	2.6	0.226	0.26	0.05	0.18	0.005
NZ19	193.44	25	68	41	62	2.5	0.098	0.20	0.03	0.16	0.005
NZ20	248.38	21	69	14	44	2.2	0.306	0.24	0.03	0.13	0.004
NZ21	349.08	2	16	10	14	0.2	0.891	0.29	0.24	0.34	0.006
NZ22	276.66	15	54	23	50	4.3	0.222	0.26	0.06	0.22	0.011
NZ23	374.95	6	88	12	28	4.7	0.611	0.13	0.07	0.21	0.003
NZ24	463.27	6	73	6	14	1.0	0.884	0.08	0.02	0.05	0.003
NZ27	421.65	4	18	20	35	3.4	0.465	0.16	0.06	0.11	0.005
NZ28	305.93	5	13	23	40	0.7	0.372	0.18	0.03	0.06	0.007
NZ29	263.41	17	-20	10	36	1.9	0.455	0.22	0.03	0.12	0.006
NZ30	343.98	4	84	17	32	1.6	0.533	0.26	0.08	0.17	0.009

*distance from the Alpine Fault to station.
**average errors of the individual ϕ and δt errors [Silver and Chan, 1991] at each station.

Splitting estimation in western offshore

Western OBS are distributed up to ~ 350 km distance from the Alpine fault. Stations are located in three distinct zones (Fig. 5.1): 1) deep ocean basin (Tasman sea) with oceanic crust close to the Puysegur subduction (NZ02, NZ03, and NZ04), 2) central region (southernmost part of the Challenger plateau) with continental-continental collisional zone near the Alpine fault (where NZ11, NZ10, NZ12, NZ13, NZ14, and NZ15 are located), and 3) northeast and central parts of the Challenger Plateau with paleo-rift systems [Wood, 1991]. In the northeast Challenger Plateau (NZ18, NZ20, NZ21, and NZ07), ϕ show prominent plate-boundary parallel ($30-70^\circ$) alignment compared to ϕ perpendicular to the plate-boundary in the central offshore region. OBS in the deep sea and near by regions (NZ02, NZ03, NZ04, and NZ10) yield bimodal distribution of ϕ with one mode parallel to the NE-SW alignment. This azimuth is also consistent with the ϕ at stations in the northeast Challenger Plateau (NZ18, NZ07, NZ20). The other mode is roughly parallel to the NW-SE or NNW-SSE direction and it is consistent with the predominant azimuth of the two-modes observed at the central stations (NZ10, NZ11, NZ13, NZ14, NZ15). NNW-SSE correlates well with the *absolute plate motion* (APM) in the Tasman sea region, indicating the possibility of asthenospheric flow-governed anisotropy in the region.

It is not clear if the NW-SE ϕ mode in the central stations (NZ15, NZ14, NZ13, NZ12) is a product of asthenospheric flow or has a stress controlled source (e.g. fluid-filled cracks and/or grain-boundary aligned consolidation). The maximum horizontal compressive stress (SH_{max}) direction of the adjacent onshore region is $\sim 115^\circ$ (Fig. 5.1-a) [Boese *et al.*, 2012], which is consistent with the NW-SE trend. If the brittle deformation (e.g. cracks and fractures) associated with the Alpine fault extends into offshore regions, relatively high compressive stress in this region may have an effect on crustal anisotropy. A rather diffuse, but fault parallel mode

(NE-SW) observed at NZ12, NZ13, NZ15 and NZ14 in central offshore suggests that S-wave splitting for some paths in this region is also sensitive to anisotropy due to plate-boundary parallel shear. However, high frequency local/regional S-phases tend to be more biased towards shallow-stress controlled anisotropy. NE-SW ϕ at stations in the northeast Challenger Plateau agrees well with ϕ patterns in the onshore regions of the northern SI (see section 5.2.3; *Karalliyadda and Savage [2013]*). This implies that the deformation around the transform-translational fault zone in the northern SI is distributed in a broader region. The other possible explanation for NE-SW ϕ in this region is that fast anisotropic directions align parallel to the paleo-rifting (parallel to rift axis; *Vaucher and Nicolas [1991]*), which are assumed to be preserved throughout the Challenger Plateau [*Wood, 1991*]. The average δt varies between 0.1-0.4 s in western offshore regions (Fig. 5.2-A). When considering the δt variations across the Alpine fault, there is a slight increase in average station δt (from about 0.17 ± 0.02 s to 0.26 ± 0.04 ; table 5.2) after the 150 km boundary-limit from the Alpine fault (Fig. 5.2-B). This could be due to accumulation of the δt along the ray path [*Silver and Chan, 1991, Rumpker and Silver, 1998*] and/or sampling of a more pronounced zone of anisotropy at depth by steeply curved ray paths. Except for NZ08, ϕ display a NE-SW azimuth that is parallel to the plate-boundary (Fig. 5.2-A) beyond the 150 km boundary limit.

Splitting parameter estimation in eastern offshore

OBS on the eastern side are mainly located on the Chatham Rise and northernmost part of the Campbell Plateau; these regions are considered as continental extensions of the eastern SI [*Sutherland, 1999*]. Stations in the eastern offshore are far away from most of the earthquakes and, thus, only a few S-phases yield acceptable quality splitting parameters (Fig. 5.1). Splitting analysis of SKS phases on Chatham Island in the Chatham Rise inferred that the region is almost isotropic [*Klosko et al., 1999*].

Figure 5.3 Zoomed-in view of the individual measurements (good quality) in eastern offshore SI. Measurements are plotted at one-quarter of the epicentral distance from the station to avoid overlapping. Histogram show the number of measurements versus an acute-angle between BAZ and ϕ (FD). Note that about 40% of the ϕ are perpendicular or high-angle ($70^\circ - 90^\circ$) to the horizontal projection of the ray paths (gray lines).

The station δt averages range from 0.1 - 0.3 s (Fig. 5.2-B) and these values are most likely representing a shallow crustal source. NZ29 and NZ28 show NW-SE ϕ that are consistent with the ϕ from adjacent onshore stations (section 5.2.3). S-waves could also be sensitive to anisotropy due to horizontal layering if the ray paths attain a longer horizontal propagation due to ray-curvature from source to station [Paulssen, 2004]. ϕ perpendicular to ray paths as observed on some event-station pairs in the eastern offshore (Fig. 5.3) may indicate such an effect ($\sim 40\%$ of the total ϕ within are $70^\circ - 90^\circ$ of the *back azimuth* (BAZ), Fig. 5.3). Chatham Rise stations (NZ23 and NZ24) have smaller average delays (~ 0.1 s or less) compared to those measured at other stations (Fig. 5.1-b). Most of the measurements made at these stations have ϕ nearly perpendicular to the ray paths. This suggests that they may sample the local/shallow anisotropy that represents a crustal source such

as shallow sedimentary layers. Therefore, complex fast azimuth patterns in eastern offshore (Fig. 5.3) most likely reflect varying anisotropic sources or geometry (i.e. changing symmetry axis orientation) of the anisotropy. Low δt (~ 0.2 s) can be attributed to either cancellation of anisotropy due to changes in the orientation of the symmetry axis, and flipping between fast and slow waves along the ray path or to crustal re-splitting owing to shallow upper-crustal anisotropy.

5.2.3 Comparison with SKS splitting measurements

We compare fast split azimuths with the recent SKS splitting analysis by *Zietlow et al.* [2013]. This was a parallel study that focused on the deep anisotropy in the off-shore region using the same OBS network, and also re-analyzed on-land SKS splitting measurements using a recent data set. We infer that the small observed δt may be caused by the re-splitting due to multi-layer anisotropy or upper-most layer crustal anisotropy and/or frequency-dependent splitting (see 3.5). If the observed anisotropy is due to multi-layer anisotropy and the medium close to the station is nearly isotropic, the observed ϕ may represent the anisotropic alignment of the most prominent source along the ray path. SKS measurements show clear a deflection in the alignment of ϕ from fault-parallel (NE-SW) near the fault to N-S or NW-SE (at high angle to the fault) alignment in the central-east and southeast of the SI (Fig. 5.4-bottom). This change reflects the change in orientation of the deep anisotropy. A similar trend can be observed from the local/regional splitting, but only in the eastern side of the plate-boundary. The prominent ϕ from the local events observed at stations NZ30, NZ29 and eastern on-land SI stations (WHZ, EAZ, ODZ, TUZ, SYZ) display APM sub-parallel ϕ and they are also consistent with the SKS azimuths. This implies that the long ray paths of the S-waves may have propagated through the lithosphere and/or asthenosphere and captured the anisotropy due to flow in the asthenosphere or fossilized anisotropy in the deep-

Figure 5.4 A comparison of local/regional splitting (top) and SKS splitting measurements [Zietlow *et al.*, 2013] on-land (re-analysis) and offshore (bottom). Green and yellow rose diagrams in the top figure indicate the on-land and offshore splitting measurements, respectively. Thick arrows indicate APM direction and thin arrows indicate the RPM directions. Green bars in the bottom figure display the SKS fast azimuths, and the size of the bars are proportional to the δt . The length of the pink bars and circular uncertainties of gray wedges denote δt and ϕ errors respectively. Crosses and dots represent the null measurements and stations with no measurement, respectively.

lithosphere. SKS ϕ azimuths attain plate-boundary parallel/subparallel direction (also RPM direction) within ~ 80 -90 km distance east of the boundary, indicating anisotropy due to plate-boundary parallel shear anisotropy. Although this is not clear from the local ϕ at stations on the Alpine fault due to multi-mode ϕ , this change is obvious at MLZ, WKZ, LBZ, RPZ stations and is consistent with SKS ϕ . This shear-zone is considered to be much broader in northern and northern-central SI than in the southern SI [Karalliyadda and Savage, 2013, Houlié and Stern, 2012, Wilson *et al.*, 2004]. This is clear offshore in the west as well, from the prominent local ϕ at stations NZ18, NZ07, and NZ20 that show RPM and SKS parallel fast azimuths (except NZ19). In the central west-offshore, local ϕ are different from the SKS fast-azimuth, but NZ12 shows one ϕ mode that is parallel to the fault. This mode can be attributed to either structure or to shear induced anisotropy. The prominent ϕ at other stations in central-offshore (NZ14, NZ13, and NZ11) are oblique to the SKS fast trends at those stations.

5.3 Integration of on-shore and offshore splitting measurements

We combined the splitting measurements from offshore OBS stations (this study), and on-shore local splitting measurements from Karalliyadda and Savage [2013], and an additional study (below) in central SI to examine the spatial patterns of ϕ in the entire plate-boundary zone of the South Island.

5.3.1 Additional land measurements: splitting measurements in the schist belt of Central SI

The work included in this section was carried out at on-land stations to improve the spatial constraints on splitting parameters in the areas dominated by schist rock in the central SI. This study was completed after the on-land manuscript publication (chapter 4).

The events analyzed for the splitting of S-phases in the central SI in previous study (chapter 4) had threshold magnitude of 3.0 and the maximum earthquake depth of 90 km. The measurements used in chapter 4 were also restricted to (section 4.2) the incidence angle (i_c) of 35° . The above event and measurements selection criteria significantly reduced the number of measurements close to the station and, therefore, measurements presented in chapter 4 had a limited splitting resolution at shallow depths close to the station (Fig. 4.2). Because of the lack of measurements to provide information on the in-situ anisotropy of the schist rock in the central SI, splitting measurements were made using shallow (<40 km) and smaller earthquakes (threshold magnitude of 2.0) than in the earlier study. This data set enables us to characterize the shallow anisotropy close to the station (within 60 km epicentral distance) with an enhanced data set. Event selection criteria can be found in section 3.1.1. Splitting measurements displayed in figure 5.5 and table 5.3 are AB grade and no-null measurements that were made from 1-5 Hz bandpass filtered data.

Circular histograms in Fig. 5.5 display fast azimuths (ϕ) measured at stations. The prominent ϕ observed at stations MLZ, EAZ, and LBZ show a clear change in orientation compared to ϕ observed from events with larger spatial and depth distribution (compare with Fig. 4.5). This suggests that the measured anisotropy at those stations is mainly controlled by a shallow source (crustal contribution) close to the station. These prominent ϕ are also apparent in the circular histograms of Fig. 5.5 as secondary-modes, suggesting depth varying or spatial effects of anisotropy.

ϕ of the off-fault stations (except WKZ) show no clear alignment with the plate-boundary (Fig. 5.5), suggesting that shallow anisotropy (depth <40 km) about 100 km away from the boundary is not controlled by the plate-boundary shear. ϕ at ODZ and LBZ are aligned with the strike of the geologic contact and fault structures in the area. The prominent ϕ measured at the station in the southeast of central SI are consistent with NW-SE aligned basement terranes (Fig. 5.5) and the structures within those terranes [Cox and Sutherland, 2007] suggesting the crustal origin of anisotropy. WKZ shows NNE-SSW ϕ that are parallel to local faulting and sub-parallel to the plate-boundary. It is possible that some of the anisotropy measured at WKZ is sampling the western side (as the maximum epicentral distance is 60 km), so that S-phases are sensitive to plate-boundary shear. Nevertheless, it is also possible that the observed anisotropy at WKZ is biased towards the near-surface structural anisotropy as it is located on top of the NE-SW aligned fault (Fig. 5.5). Alternatively a series of NE-SW striking (sub-parallel to the plate-boundary) faults in this region could have well-developed shear-zones beneath them, and if those shear zones are linked at depth with the plate-boundary shear-zone, producing a single large shear-zone [Otsuki, 1978]. If that is the case, ϕ at WKZ likely represent the alignment of both structural and shear-zone anisotropy. LBZ shows plate-boundary parallel ϕ from the splitting of S-phases with depths ≤ 100 km (Fig. 4.5-b), but not with depths ≤ 40 km (Fig. 5.5). This implies that the plate-boundary shear-zone at depths 40 km and shallower than 40 km is not as broad as at depths 40-100 km. An absence of plate-boundary parallel ϕ at both LBZ and EAZ at shallow depths (Fig. 5.5) suggests that shear-deformation is likely distributed within at least ~ 30 km (distance from plate-boundary to WKZ is about 90 km and maximum epicentral distance is 60 km) distance from the plate-boundary at shallow depths (<40 km). The maximum limit of the distributed shear could be ~ 90 km from the fault at depths ≤ 40 km. Beyond that distance limit, anisotropy may reflect structural-induced anisotropy. The stations that are adjacent to plate-boundary (JCZ, FOZ and WVZ) show a bimodal distribution of fast azimuth, but at WVZ both modes

have approximately similar azimuths (NNW-SSE and NW-SE). One fast mode at FOZ and JCZ is plate-boundary sub-parallel, which could be attributed to localized shear at sub-crustal depths and/or due to fault structure. NNW-SSE and NW-SE modes at FOZ and WVZ are consistent with the SH_{max} orientation in the region (compare with Fig. 4.11). Table 5.3 summarizes the average splitting measurements and their statistics as discussed in section 3.4.

Figure 5.5 ϕ from the shallow S-phases recorded at the stations in schist rock of the central SI. ϕ at stations display as circular histograms. Inset figure shows the geology of the area and events used for this analysis. Purple to pink colors marks schist rock with increase in metamorphism.

Table 5.3 All measurements and statistics of the shallow (≤ 40 km) splitting measurements in central SI.

station	no.	mean δt	error δt	mean ϕ	error ϕ	SE of ϕ	CSD of ϕ
WKZ	56	0.20	0.007	26	2.6	9	44
JCZ**	339	0.16	0.005	43	2.5	39	76
LBZ	41	0.15	0.005	-37	1.9	15	50
WVZ	127	0.14	0.007	-24	2.6	3	34
EAZ	8	0.14	0.006	81	1.1	25	45
MLZ	19	0.10	0.006	-57	3.0	14	43
TUZ	7	0.12	0.010	86	2.8	28	47
ODZ	32	0.10	0.004	-17	2.5	5	29
FOZ**	42	0.15	0.005	79	2.0	33	62

**stations with clear double-modes of ϕ

Delay time variations across the plate-boundary zone and Schist rock in the central SI are analyzed from the AB graded (no-null) splitting measurements with dense error contours ($\lambda_{max} \geq 5$; section 3.2.3) and $\delta t \leq 0.6$ s. Figure 5.6 shows the individual (top-gray circles) and average (bottom/top - black circles) δt that are calculated from those measurements as a function of distance from the Alpine Fault (dashed line in Figure 5.6). The two stations that are located further south, away from the central plate-boundary zone are marked in black squares. Except MLZ that is located further south and does not lie on the schist rock, the other stations show an increase in average δt towards the plate-boundary (the Alpine fault). Although individual measurements are a bit scattered, this trend is also visible in the individual station measurements (Fig. 5.6-bottom). Individual measurements also show an increase in scatter towards the plate-boundary, implying a more complex anisotropic structure towards the plate-boundary. A similar increase in velocity anisotropy had been determined by the petrophysical studies in the same schist terrane of the central SI by *Okaya et al.* [1995], and this was attributed to an increase in metamorphic grade towards the Alpine fault (section 1.5). Local S-phases recorded on the stations that are close to the fault (e.g. JCZ, FOZ) could sample anisotropy as high as ~ 0.5 s and as low as ~ 0.05 s. These splitting measurements were obtained from high-frequency S-phases (filtered at 1-5 Hz), thus

Figure 5.6 Top: The Averaged (black) and individual (gray-open) δt (measurements with $\lambda_{max} \geq 5$ and $\delta t \leq 0.6$ s) variations across the schist rock in central SI from shallow S-waves (< 40 km). Distance to station is measured from the Alpine Fault (dashed line). Error bars in both plots denote the standard error of the mean station δt . Note that MLZ and TUZ (black squares) are located in the southernmost region of Central SI. **Bottom:** Same as top figure, but zoomed-in view of the average δt variations across the schist rock in central SI. Note a moderate increase in δt at the central SI stations from east to west.

we could minimize the effects due to frequency-dependence, but smaller delays could be caused by the re-split S-phases from upper-most crustal anisotropy. *Pulford et al.* [2003] found very little splitting anisotropy (0 - 0.08 s) from high-frequency S-phases from active source experiment across the Alpine Fault. As determined by the petrophysical studies in the same area, variable delays can also be caused by changing ray propagation with respect to the symmetry of the anisotropic system [*Godfrey et al.*, 2000] or the orientation of the foliations planes [*Barruol*, 1993]. According to figure 1.6, such a variation can produce a range of δt (~ 0 -0.9 s), as seen in our in-situ data set. Thus, unless there is crustal re-splitting we favor the

proposed idea by *Godfrey et al.* [2000] to explain the range of delays from shallow local-splitting. The increase in δt from Tuz to JCZ possibly reflects the increased metamorphic grade, finite strain, and hence, the strength of anisotropy towards the plate-boundary in agreement with the petrophysical studies [*Okaya et al.*, 1995] and finite strain estimations from bending related geological markers [*Little et al.*, 2002b]. Strain may accumulate as crustal shear-zones with foliations parallel to the plate boundary. Thus, the splitting at shallow depths in the Hasst schist of the central SI suggests that such a shear-zone is likely extends ~ 30 -90 km distances from the plate-boundary at shallow depths.

5.3.2 2-D δt tomography and spatial averages of ϕ

Karalliyadda and Savage [2013] attributed variable splitting parameters from the on-shore experiment to stress and mantle-shear induced anisotropy in different regions. However, the spatial extent of these anisotropic sources is not well resolved. Thus, we attempted to constrain the boundary limits of the observed patterns using the enhanced data set, which includes OBS measurements and additional central SI measurements (appendix B.4). Because of this enhanced data set, we expect some discrepancies between the solutions obtained from the on-land tomography/spatial averaging and this combined analysis. As there are limited measurements in offshore, this analysis assumes a single anisotropic layer at depth, but this assumption may not be true especially in on-land as the depth varying anisotropy is obvious throughout the region [*Karalliyadda and Savage*, 2013]. Thus this study will give us a first-order approximation on the spatial patterns of anisotropy.

We used three runs (processing cycles) with different parameters as summarized in table 5.4 and *run₃* is used to determine both tomography and spatial averages. The results from *run₃* are discussed in this section. *run₁* is only used to obtain the spatial averages of ϕ (Fig. 5.9) using quadtree gridding with maximum 50 km grid-sizes

Table 5.4 Parameters used for the delay-time tomography and/or spatial averaging

run	block size (km)	grid	no paths	weighting	grade
<i>run₁</i>	50	quadtree	7035	no-weighting	AB-grade/best-filter
<i>run₂</i>	50	regular	4838	no-weighting	AB-grade/good*/best-filter
<i>run₃</i>	25	regular	4838	tomo.weighting	AB-grade/good*/best-filter

*best filter measurements with dense error contours ($\lambda_{max} > 5$). See appendix B.4.

and to compare with the spatial averages from *run₃* (Fig. 5.8) with 25 km regular grids.

The final solution from *run₃* is constrained from 4838 ray paths, 2130 events and 56 stations from 25 km regular gridding. Log resolution contours displayed in figure 5.7(a) are used to determined the well constrained solution. By adding OBS splitting measurements and enhancing the splitting data set in the central SI, we have increased the spatial resolution of the tomographic solution and minimized uncertainties of the spatial averages in the central SI data. Comparing to figure 4.10, most of the features resolved by the on-land tomography remain consistent, but combined analysis has a better spatial constraint on those features. For example, high strength of anisotropy (SA) along the central part of the Alpine fault is more resolved than that in figure 4.10. This feature does not show a broadly distributed character within the depth resolution ($\lesssim 60$ km) of the central SI data. Relatively high SA in the southern west-offshore SI is likely associated with the oceanic crust (Fig. 5.1-a). In the northern SI, high strength of anisotropy in both on-land and adjacent offshore (northeast) regions appears to correlate with the complexities associated with the subduction-collision transition. The region where the Alpine Fault terminates into the Marlborough fault zone (transition faults) appears to have a high strength of anisotropy (about lat:-42° and lon:171.5°). Note that the high strength of anisotropy observed in some regions (e.g. far-west offshore) exists in

the regions with low resolution (Fig. 5.7-a; $\log_{10}$ resolution $\lesssim 4$) and may indicate artifacts in the solution.

Both regular (Fig. 5.8-left) and quadtree (Fig. 5.9-bottom) gridding with tomographic weighting and without weighting respectively yield similar spatial patterns. According to spatial averages of ϕ (Fig. 5.8-left), the spatial distribution of anisotropic structure appears to be characterized by three distinct anisotropic sources. The region with yellow bars: ϕ are parallel/sub-parallel to the relative plate motion (RPM = $\sim 70^\circ$; *DeMets et al.* [1994b]) direction (thin arrows in Fig. 5.8) and and strike of the Alpine Fault ($\sim 55^\circ$); the region with blue bars: ϕ (outermost- $\sim 330^\circ \pm 45^\circ$ are consistent with the absolute plate motion (APM; *Gripp and Gordon* [2002]) direction (bold blue arrows in Fig. 5.8). The zone with red bars: ϕ ($\sim 115^\circ \pm 45^\circ$) are consistent with SH_{max} azimuth (bold-gray bars; *Townend et al.* [2012]). Except in the SH_{max} governed zone, local/regional ϕ are mostly aligned with the SKS fast azimuths (gray bars with closed circles in Fig. 5.8) in other regions, implying either a deep source of anisotropy or crust-mantle coupled anisotropy. Yellow bars are distributed in a wide region across the upper-central and northern SI (Fig. 5.9-bottom). However, in the central SI, the plate-boundary parallel zone tends to be masked by a zone with a plate-boundary oblique azimuth (zone with red bars) that is aligned with the SH_{max} direction in that region [*Townend et al.*, 2012] or the axis of shortening [*Houlié and Stern*, 2012]. The outer-most zone with blue bars has a very diffuse pattern, but it is likely controlled by the present-day asthenospheric flow or preserved fabrics in the deep upper-lithosphere (Fig. 5.9-bottom).

Adding the OBS splitting measurements and additional inland measurement to the previous inland data set, we have extended delay time tomographic solution in Fig.4.10 into near offshore regions and the quality of the solution in central SI is improved. The offshore portion of the tomographic solution is only well constrained on the western side within ~ 120 km of the plate-boundary. In the region with high

Figure 5.7 Delay time tomography of both offshore and on-land splitting. Resolution matrix and tomography results are shown in **a** and **b** respectively. The darker the region, the higher the resolution. The regions with the higher strength of anisotropy (warm colors in **b**) are considered to be well constrained if those areas have higher resolution ($\log_{10}$ resolution $\gtrsim 4$ in Fig.a).

Figure 5.8 Left: Spatial averages of ϕ ($\bar{\phi}_n$) of offshore and onshore splitting measurements using 25 km grid blocks. ϕ were weighted by the tomographic weighting function (section 3.6.1). Bars are color coded according to the regional consistencies that are likely aligned with the RPM/Alpine Fault (yellow), APM (blue) and SH_{max} (red) directions. $\bar{\phi}_b$ that are not consistent with those regional trends and/or anti-correlated (angular difference $> 45^\circ$) with the expected regional trends are denoted by black bars. Long and short colored bars indicate the well (angular difference $\leq 15^\circ$) and moderately ($15^\circ < \text{angular difference} \leq 45^\circ$) correlated $\bar{\phi}_b$. The bars with black circles denote $\bar{\phi}_b$ with standard error (SE) $> 25^\circ$. Note gray color symbols (see the left-top legend) represent SH_{max} [Tounend *et al.*, 2012], SKS ϕ and P_n anisotropy [Savage *et al.*, 2007a] measurements from previous studies. **Right:** Ray paths, events, and station distribution that were used to determine spatial averages and tomography.

Figure 5.9 The colors and symbols are the same as figure 5.8(left), but AB-grade best-filter (run_1 in table 5.4) measurements (top-ray paths) are spatially averaged (bottom) from quadtree gridding with 50 km maximum grid size.

resolution, the patches of relatively high strength of anisotropy can be detected in the two-subduction zones (see Fig. 5.7). The inland spatial patterns of ϕ with the improved data set (Fig. 5.8) are better constrained compared to the patterns observed from the initial data set (see section 4.4.3). For instance, the zones of plate-boundary parallel ϕ detected in the northern and southern central SI (Fig. 4.11) are likely linked at depth as a single shear zone, which is produced by the present-day plate-boundary deformation. This zone is broader in the northern SI and extends into offshore regions than in the southern South Island (see Fig. 5.8). In addition to this 2-D spatial analysis, we used 3-D tomographic analysis (SWST) that allows us to characterize the 3-D distribution of anisotropy in terms of mineral alignment especially in the deeper part ($\gtrsim 12$ km). The methodology used and the preliminary results of this sub-study are given in sections 3.6.2 and section 5.3.3 respectively.

5.3.3 Preliminary results from 3-D tomography

We used the AB-grade (best filter) splitting data set that was used in the combined analysis of 2-D delay time tomography (section 5.3.2) for a 3-D inversion. Tomography code (SWST) by *Abt and Fischer* [2008] is used for the analysis (section 3.6.2). The model space is defined so that the x -axis is aligned with the plate-boundary and the y -axis is perpendicular to the plate-boundary. Because of the extreme computation time (approximately 10 days for 100 iterations), we have limited this analysis to one main run (run_3). We performed some test runs prior to the main run in order to select the inversion and model parameters. The table 5.5 summarizes the model-space, model and inversion parameters that were used to obtain the inversion solution.

Abt and Fischer [2008] noticed that the final inversion solution depends on the starting model. In their test runs with splitting from known structures and different

starting models, the final model obtained from the average starting model produced the splitting parameters that fit well with the observations in both well-resolved areas and under-sampled regions. Final models determined using the other starting models did not yield a better-fit in the under-sampled regions because of the non-linearity of the problem. Therefore, we used the average starting model for this analysis. The average starting model is determined from averaging the observed splitting parameters in the 3-D model space after ray tracing through the model blocks [see *Abt and Fischer, 2008*, appendix A].

We used the plotting options given in the 3-D tomography code to illustrate the final model with the resolved a-axis azimuths (fast direction). Figures 5.10 and 5.11 display the 2-D views (map views) of the 3-D models with a-axis azimuths. In figure 5.11, the independently well-resolved model (Fig. 10(b)) is divided in to several depth-layers.

Table 5.5 Model-space, model and inversion parameters

Model-space setup		
parameter	value	variable ^a
block size	25 km	MTP(8)
no. of blocks	12160	nblks
blocks per layer	1216	nblyr
no. events	2982	nevents
no. rays	7020	nrys
Model		
parameter	value	variable ^a
run no.	3	run
crystal symmetry	hexagonal	MTP(2)=2
Inversion		
parameter	value	variable ^a
type	real	RTP(1)=0
iterations	100	ITP(3)
starting model	average	ITP(2)=1
no model parameters to solve	$\text{np}=\theta, \psi, \alpha$	MTP(1)=3
larger-volumes	yes	ITP(5)=70
smoothing	no	ITP(7)=0

^aas assigned in the 3-D tomography code (SWST)

Main features in the independently resolved model

This section discusses the main features that have been identified in the independently well-resolved model (Figs. 5.10 and 5.11-d, e, & f). Note that *Abt and Fischer [2008]* use a-axis orientation (olivine) to explain the alignment of fast direction (ϕ). The overall well-resolved model suggests (Fig. 5.10(b)) that the plate-boundary parallel shear dominates close to the plate-boundary. ϕ are either parallel (red) or sub-parallel (magenta) in the near by regions of the plate boundary. Well resolved plate-boundary parallel azimuths in the south-east of the Alpine fault agree well with the schist-band in the region, suggesting a high-strained zone possibly linked with the plate-boundary shear. Most of the resolved structures also agree with the 2-D spatial averages of ϕ (compare Fig. 5.10(b) with 5.8 and Fig.5.9). Further investigation into inversion modeling of this data set would involve exploring smoothing and damping parameters, comparison with models obtained assuming orthorhombic symmetry and testing the stability of the inversion solution with different starting models. The final model solution may vary with the selection of different starting models and crystal symmetry (or elastic coefficients). Thus, the tectonic interpretations based on the final model solution given here, which is constrained from the average starting model with hexagonal symmetry, could be preliminary. The following section summarizes main features resolved at different depths of the independently well-resolved model (Fig. 5.10(b)).

(a) a-axis azimuths solved from all rays without considering resolution

(b) Azimuths from the well-resolved individual blocks.

(c) Azimuths from the well-resolved combined blocks.

Figure 5.10 3D-tomographic solutions. 2D view (map view) of the 3-D model with a-axis azimuths. Bars are color-coded according to the a-axis orientation with respect to the plate boundary (green line). A gradual change in plate-boundary parallel to plate-boundary perpendicular azimuths are indicated by color changes from red to blue color, respectively. Azimuths in Fig. 5.10-(a&b) are scaled by the diagonal element of the resolution matrix (RM_{dia}). The thicker the bars, the higher the resolution. The bar-length denotes the strength of anisotropy. Note that the alignments with high strength of anisotropy observed in most of the regions in Fig. 5.10-(a) have been eliminated in the well resolved model (Fig. 5.10-(b)).

(a) all: depth ≤ 30 km(b) all: $30 \text{ km} < \text{depth} \leq 100$ km(c) all: depth > 100 km (d) resolved: depth ≤ 30 km(e) resolved: $30 \text{ km} < \text{depth} \leq 100$ km(f) resolved: depth > 100 km

Figure 5.11 2-D depth layers through the 3-D model

Resolved features:

- **Fig. 5.11(d):** ϕ orientations at depths ≤ 30 km suggest that the plate-boundary parallel/sub-parallel ($\sim 30^\circ$ from the plate-boundary) shear could occur in a ~ 40 -80 km wide zone around the plate-boundary. This zone is narrower (~ 40 km) towards southern SI compared to northern SI and is characterized by a high strength of anisotropy compared to other well-resolved features. Consistent ϕ with low-strength of anisotropy that is oblique to the plate-boundary is significant in the eastern offshore and south of the Fiordland subduction zone. This feature could be related to the preserved anisotropic fabrics from the ancient Gondwana-subduction or be representative of the deeper flow in the asthenosphere (APM direction in Fig. 5.8).
- **Fig. 5.11(e):** depths $30 \text{ km} < \text{depth} \leq 100 \text{ km}$, the well-resolved features are mainly located in the southern SI. A broad plate-boundary parallel zone seems to correlate with the schist-band in the southern SI (Fig. 1.4). We do not know the depth extent of this schist-band, thus this alignment could represent the straining in the schist-band itself and/or deep extension of the plate-boundary shear zone that fans out from the plate-boundary.
- **Fig. 5.11(f):** Anisotropic alignment in the two subduction zones is well resolved, indicating the fore-arc mantle wedge in both Hikurangi and Fiordland subduction zones are characterized by the trench-parallel flow or shear at depths > 100 km. The resolution in the central SI is poor, therefore we could not constrain the deep extension of the plate-boundary shear.

5.4 Discussion and conclusions

Local/regional events recorded on the OBS network on offshore SI enabled us to enhance the spatial resolution of anisotropic structure in the AUS-PAC plate-boundary zone in SI. Both long and short S-wave paths used in this analysis are likely sensitive to different structures at depth. Schematic 5.12 shows the possible ray paths that could sample different sources of anisotropy in the central SI plate-boundary zone. If we assume a broad zone of an asymmetric lithospheric shear-zone parallel to the plate-boundary as predicted by *Duclos et al.* [2005] (possibly extending down to the asthenosphere) (Fig. 5.12), the majority of S-waves of deep or long ray paths are likely to sample the anisotropy of this shear-zone (red ray paths in Fig. 5.12). Thus, they will yield plate-boundary parallel ϕ unless anisotropy at shallow depth is strong. Near vertical shear-planes, which are in the vicinity of the plate-boundary, can be tilted to produce dipping anisotropy or can be horizontal away from the plate-boundary. In such cases the strength of splitting may vary. This may also be the case if ray segments are tilted (i.e. not near vertical) in vertically aligned media (Fig. 1.6). In the regions where this shear is not distributed or strong, S-phases of long and/or deep paths will be sensitive to present-day asthenospheric flow-derived or preserved anisotropy from the asthenospheric flow in the deep lithosphere, resulting in APM parallel/subparallel anisotropy. In the southern SI especially in offshore and eastern on-land this could be the case (Fig. 5.9). S-phases from shallow and shorter ray paths mostly reside in the shallow crust and they will be sensitive to crustal anisotropy (blue ray paths in Fig. 5.12). In the central SI, the majority of events occur at shallow crustal depths. Thus, those S-phases can be sensitive to both shallow shear-induced anisotropy and crustal stress or/structure induced anisotropy, depending on the geometry of ray propagation.

We have identified a zone with SH_{max} parallel ϕ in the northern-central/central SI, where we do not have a good depth resolution (section 5.3.2). High-stress conditions

Figure 5.12 Schematic explaining the ray configuration and sensitivity of ray paths to different anisotropic zones. Red-colored ray paths likely sample the plate-boundary parallel shear-zone, whereas blue ray paths may be sensitive to more shallow sources (e.g. crustal anisotropy).

in the shallow depths of the Alpine fault zone may have a strong control over the anisotropy at shallow depths. This effect diminishes in the southern-central and the southern SI, and plate-boundary parallel ϕ or APM sub-parallel ϕ become dominant (Fig. 5.8). This observation could be partly due to presence of deep ray paths that could sample deep anisotropy. It is possible that there is a plate-boundary induced shear zone beneath the shallow stress-induced anisotropic zone in the central SI. This zone is likely distributed in the west-offshore at lithospheric depths and not extended into east-offshore at lithospheric depths (Fig. 5.12). This type of an asymmetric anisotropic shear is possible due to the oblique strike-slip nature of the plate-boundary. However, it is also possible that this zone is broadly distributed in the deep-lithosphere (deeper than 90 km), where we do not have ray-coverage, and extends into the asthenosphere. Nevertheless, SKS splitting suggests that such a shear-induced body is absent in the eastern offshore regions at deeper depths and is mainly governed by the present-day asthenospheric flow (Zietlow *et al.* [2013] and Fig. 5.4). The northern SI is mainly characterized by a broad zone (~ 200 km wide) of shear beneath the plate-boundary and the transitional fault zone.

We think that long S-wave paths possibly encounter more than a single layer of anisotropy, or plunging anisotropic layers, when they reach the station. Even if

the anisotropy is characterized by a horizontal symmetry axis, the S-phases may not sample the full effect of the properties of the anisotropic medium because of the ray geometry (e.g. curved ray paths). At least some parts of the long and shallow ray paths are likely sensitive to transverse isotropy with vertical symmetry axis. In that case, observed variable ϕ can be attributed to the effects of both transverse isotropy with vertical symmetry and horizontal symmetry (i.e. azimuthally anisotropy). Depending on how rays propagate in the layered media and the anisotropic symmetry, splitting may reflect an integrated effect at the receiver. Measured low δt , therefore, probably represents either the anisotropy of the upper most layer (crustal source) or the integrated effect of the variable anisotropy along the ray path. This integrated effect can be addressed by 3D splitting tomography (section 5.3.3).

It is hard to conclude whether this entire region agrees with either localized deformation or the thin-viscous sheet model based only on the observed patterns of ϕ and δt of the local/regional S-wave splitting. These measurements, however, suggest that at least in the northern SI the shear-zones associated with the plate-boundary zone is broad (~ 100 - 150 km), as predicted by the thin-viscous sheet model. In the northern-central SI, this deformation is not well constrained due to lack of depth resolution and shallow stress-induced anisotropy. S-wave splitting from shallow earthquakes (< 40 km) in the central SI suggests that plate-boundary shear is most likely to occur up to 30-90 km distance east of the plate-boundary within the shallow depths (section 5.3.1). Thus, in the southern SI and the lower-central SI, shear induced deformation is narrower ($\lesssim 80$ km) east of the Alpine fault than the northern SI. In the rest of the southern SI, anisotropy is governed by the asthenospheric flow at deeper levels.

Chapter 6

Deep anisotropic structure beneath the Hikurangi subduction zone

This chapter presents the results from a deep anisotropy investigation in the southern North Island (NI) from SKS, ScS, and teleseismic S (Tel. S) splittings on the SAHKE transect (Fig. 3.6). Previous work around the study area is discussed in section 1.6. Data and techniques that have been used during this analysis are explained in sections 3.1.3 and 3.2.2. SKS splitting was determined from the transverse energy minimization method and ScS/Tel. S splitting was estimated from the eigenvalue minimization method (Fig. 6.1). Splitting parameters are then analyzed for frequency-dependent (section 6.2.1) splitting to extract more information about the anisotropic structure in the region. All the good/fair quality splitting measurements were also re-analyzed with the method by *Vecsey et al.* [2008] to examine splitting parameters in the ray-coordinate (LQT) system. Finally, we combined and compared our splitting measurements with the previously published data in order to fully understand the lateral anisotropic structure in the southernmost Hikurangi subduction zone.

6.1 Splitting parameters from SKS, ScS and Tel. S phases

We assessed 95 well resolved SKS, ScS and Tel. S splitting measurements from 27 events (see Appendix C.2) to understand the mantle anisotropy in the east of the southern NI. Teleseisms with magnitudes ≥ 6 were used to extract SKS, ScS and teleseismic S arrivals (see table 3.2 for event selection criteria). The events of selected ScS and Tel. S splitting measurements have a minimum source-side

Figure 6.1 Examples of the splitting measurements using transverse component minimization with SKS phases, and eigenvalue minimization technique with ScS and Tel. S splitting. (a) seismogram with S-pick and the selected S-wave analysis window. (b) waveform plots of rotated (top) and corrected horizontal components (bottom) (c) contour plots of errors and the best splitting parameter (blue cross). The bold line indicates the 95% confidence interval (d) comparison of waveform and particle motion plots before and after correction for splitting.

Figure 6.2 SAHKE network, tectonic settings and splitting parameters. The double headed arrows that are plotted at 100 km ray piercing points denote the ϕ from ScS (blue), Tel.S (yellow) and SKS (red) phases and their lengths indicate the δt . The gray (80.2° NE) and white (52.8° NW) arrows indicate the relative (RPM; *DeMets et al.* [1994b]) and absolute plate motion (APM relative to hotspot reference frame; *Gripp and Gordon* [2002]) directions. The shallow slab model [*Ansell and Bannister*, 1996] is denoted by the dashed-white lines. LTW1-3 and LTN1-6 (squares) are the Tararua array [*Stuart et al.*, 1995].

depth of 400 km and 450 km, respectively (table 3.2). The study region is located in the southern most extension of the Hikurangi subduction zone (Fig. 6.2). Splitting measurements are determined from the events recorded on both temporary stations that are located on the NW-SE oriented transect and one permanent station (WEL) in Wellington region (section 3.1.3). Splitting measurements at Wellington station provide a complementary measurement set. Event and measurement information are given in table 6.1 (good and fair measurement) and table C.3 (null measurements). We considered the measurement as null if $12^\circ \geq |\phi - \phi_p|$ and $80^\circ \leq |(\phi + 90^\circ) - \phi_p|$, where, ϕ_p is initial polarization direction. All SKS and ScS phases have almost vertical incidence ($i_c \leq 12^\circ$), and i_c of the Tel. seismic waves range between $16 - 22^\circ$ (Fig. 6.3 and section 6.2.3).

Figure 6.3 2-D cross-section of the slab, station locations, and ray paths along the SE-NW profile in figure 6.2. The stations along the profile are denoted by inverted triangles. Ray paths are obtained from the ray trace [Abt and Fischer, 2008] through the 1-D (AK135) velocity model without encountering the velocity gradient in the slab. Note the near vertical incidence of most of the rays and Tel. S arrivals with slightly higher i_c angles. Also, note the vertical exaggeration.

About 93% of the measurements display NE-SW ϕ , which is aligned with the strike of the Hikurangi trench, the subducting slab and the faulting in the overriding Australian plate. NE-SW ϕ varies between $5 - 85^\circ$ and δt values ranges from ~ 0.5

Figure 6.4 δt versus BAZ (a) and Fast azimuth (ϕ) versus BAZ. Fair quality measurements (non-null, but error $\delta t > 0.5$ s and/or error $\phi > 20^\circ$) are marked in black symbols.

- 3.0 s (Figs. 6.2 and 6.4). Compared to SKS and Tel. S, ScS phases yield NNE-SSW and low δt (ranging from 0.5 to 1.5 s) at most of the stations. Measurements from Tel. S and ScS are limited to 280 - 340° and 10 - 30° BAZ, respectively, and sample the fore-arc and/or mantle wedge region (Fig. 6.4). Although most of the ϕ are NE-SW aligned parallel to the trench, ϕ of a few well resolved measurements from ScS and Tel. S are oriented oblique (-20 to 60°) to the trench and consistent with the absolute plate motion (APM direction) (Fig. 6.2). It is hard to provide accurate constraints on the backazimuthal variations of splitting parameters from ScS and Tel. S phases as the arrivals are limited to a small range of azimuths. SKS phases, have relatively good coverage, but show no clear BAZ dependency of splitting parameters (Fig. 6.4 and section 6.2.3). However, both ScS and some SKS phases from 0° - 30° BAZ tend to show relatively smaller δt (ranging from 0.5 - 1.5) compared to that from other BAZ (Fig. 6.4).

In general, the estimated splitting parameters from the three different phases are in agreement with the previous splitting anisotropy studies in the western side (Taraua array and LBLU) and in Wellington [Marson-Pidgeon and Savage, 1997, Marson-Pidgeon et al., 1999, Gledhill and Gubbins, 1996]. However, they show a wide range of values as found by Marson-Pidgeon and Savage [2004b], implying lateral variations in the anisotropic structure in the south of Hikurangi subduction zone.

Table 6.1 Measurement information of the splitting of SKS ScS and Tel. S phases on SAHKE transect.

station	event lat.°	event lon.°	phase	δt (s)	error δt (s)	ϕ°	error ϕ°	grade ^s
LE4	-169.848	52.876	SKS	1.050	0.544	37	18.000	fair
T004	-169.848	52.876	SKS	1.238	0.919	39	21.500	fair
T004	-97.782	16.396	SKS	1.312	0.694	12	15.750	fair
T004	-175.870	51.451	SKS	1.087	1.003	34	20.500	fair
T004	63.951	28.777	SKS	1.538	0.628	42	17.250	fair
T004	96.548	33.165	SKS	1.725	1.378	17	23.250	fair
T007	-68.328	-22.227	SKS	1.762	0.450	62	8.750	good
T010	-175.870	51.451	SKS	1.650	0.525	66	12.750	fair
T010	93.471	11.132	SKS	2.175	1.322	68	23.250	fair
T014	-97.782	16.396	SKS	1.440	0.340	43	10.500	good
T014	-175.870	51.451	SKS	2.812	0.309	29	3.250	good
T014	-109.159	24.690	SKS	2.925	0.403	42	4.000	good
T016	-171.836	52.050	SKS	1.462	0.319	64	11.000	good
T016	159.843	52.498	SKS	2.438	0.591	61	6.750	fair
T016	-97.782	16.396	SKS	1.913	0.506	52	9.000	fair
T016	-175.870	51.451	SKS	1.050	0.178	53	11.000	good
T016	-178.649	51.423	SKS	1.050	0.375	53	15.750	good
T016	93.471	11.132	SKS	2.400	0.900	47	15.500	fair
T020	-171.836	52.050	SKS	1.875	0.562	31	11.500	fair
T020	-175.361	51.374	SKS	2.475	0.553	23	5.250	fair
T020	-97.782	16.396	SKS	1.762	0.628	54	10.000	fair
T020	-178.649	51.423	SKS	1.762	0.806	21	9.750	fair
T020	92.831	13.667	SKS	1.087	0.338	45	12.750	good
T022	-178.649	51.423	SKS	2.812	0.562	79	3.750	fair
T022	92.831	13.667	SKS	3.263	0.487	83	3.500	good
LE4	123.519	6.218	Tel.S	2.075	0.294	23	16.500	good
LE4	122.974	4.195	Tel.S	1.825	0.287	6	14.500	good
LE4	123.039	4.077	Tel.S	1.725	0.319	-32	22.000	fair
LE4	123.467	6.486	Tel.S	1.850	0.662	43	24.250	fair
LE4	116.720	-6.873	Tel.S	2.350	0.275	-2	11.000	good
T004	123.519	6.218	Tel.S	2.250	0.275	35	15.500	good
T004	122.974	4.195	Tel.S	2.200	0.487	26	24.000	fair
T004	116.720	-6.873	Tel.S	2.425	0.244	-2	10.000	good
T004	139.869	26.978	Tel.S	2.725	0.562	37	4.750	fair
T016	139.154	28.360	Tel.S	3.400	0.431	30	5.000	good
T016	123.519	6.218	Tel.S	2.150	0.119	31	5.500	good
T016	123.247	6.528	Tel.S	2.450	0.156	-20	8.000	good
T016	122.974	4.195	Tel.S	1.850	0.269	37	4.250	good
T016	123.039	4.077	Tel.S	1.975	0.069	27	4.250	good
T016	116.720	-6.873	Tel.S	3.000	0.263	46	16.500	good
T016	139.869	26.978	Tel.S	2.600	0.569	35	7.500	fair
T016	123.409	6.718	Tel.S	1.875	0.656	33	18.000	fair
T016	123.467	6.486	Tel.S	1.775	0.163	47	9.250	good
T020	123.519	6.218	Tel.S	2.300	0.325	55	9.750	good
T020	123.247	6.528	Tel.S	2.575	0.244	-45	8.000	good
T020	122.974	4.195	Tel.S	1.950	0.487	46	12.000	good

Continued on next page...

Table 6.1 – Continued

station	event lat.°	event lon.°	phase	δt (s)	error δt (s)	ϕ°	error ϕ°	grade [§]
T020	123.039	4.077	Tel.S	1.725	0.244	60	11.000	good
T020	116.720	-6.873	Tel.S	2.675	0.444	41	27.500	fair
T020	123.467	6.486	Tel.S	2.150	0.225	65	5.000	good
T022	123.519	6.218	Tel.S	2.375	0.431	30	14.500	good
T022	123.247	6.528	Tel.S	2.375	0.175	-64	7.500	good
T022	123.467	6.486	Tel.S	2.075	0.188	42	4.000	good
LE4	179.116	-23.307	ScS	1.325	0.044	32	7.500	good
LE4	179.760	-23.784	ScS	0.825	0.188	6	6.750	good
LE4	-178.425	-17.794	ScS	0.675	0.094	-5	11.250	good
LE4	178.394	-26.142	ScS	1.125	0.031	31	3.500	good
LE4	-178.826	-20.799	ScS	1.475	0.219	-27	10.000	good
T004	178.394	-26.142	ScS	1.150	0.044	28	3.000	good
T004	-178.826	-20.799	ScS	0.850	0.188	-22	15.250	good
T014	179.116	-23.307	ScS	1.350	0.056	12	4.250	good
T016	179.116	-23.307	ScS	1.050	0.225	48	16.000	good
T016	178.394	-26.142	ScS	1.150	0.087	10	7.250	good
T016	-178.826	-20.799	ScS	0.975	0.113	21	6.000	good
T020	179.116	-23.307	ScS	1.550	0.100	3	8.000	good
T020	178.394	-26.142	ScS	0.950	0.025	16	3.000	good
T020	-178.826	-20.799	ScS	1.000	0.106	34	9.500	good
T022	-178.826	-20.799	ScS	0.575	0.169	-3	17.000	good

Lateral variations of splitting in the northern Hikurangi subduction zone across the CVR (Fig.1.1) had been investigated by *Greve and Savage* [2009] via finite difference and analytical modeling, assuming an anisotropic model with a constant average isotropic velocity. The study found a model with a highly anisotropic mantle-wedge (15 % S-wave anisotropy), subslab region with 3.5 % anisotropy, and isotropic slab and back-arc region as the best-fit (forward) model that explains the observed measurements. High anisotropy in the mantle wedge is attributed to aligned-melt bands/cracks together with mineral alignment. Because of the shallow subduction below our array in the southern NI and the absence of arc activity, we do not expect to see such strong wedge contribution of anisotropy as in the model proposed for

[§]grade of the non-null measurements. good: error $\delta t \leq 0.5$ s and error $\phi \leq 20^\circ$.

the northern NI. However, lateral variations in splitting parameters could be caused by other phenomena (e.g. isotropic velocity variations, the upper-plate structures, properties in the shallow-wedge, and properties within the slab). To investigate whether the scatter in splitting parameters is due to any spatial effect we analyzed splitting parameters along the profile across the subduction zone, combining our splitting measurements with the previous splitting data. The detailed description of the lateral variations is given in section 6.2.2.

6.2 Synthesis of splitting parameters

The most recent studies across Tararua array and modeling of splitting parameters at SNZO in the Wellington region suggested laterally varying anisotropy and complex geometry beneath the SNZO station [*Marson-Pidgeon and Savage, 2004a,b*]. The main findings of these studies was the NE-SW aligned anisotropic domains beneath the Tararua array (NE of SNZO) that possibly extends towards the Wellington region. The SAHKE transect is aligned with the Tararua array, but extends further SE from the SE end of the NW-SE aligned arm of the Tararua array (Fig. 6.2). The NE-SW aligned anisotropic domain that was constrained by *Marson-Pidgeon and Savage* [2004b] lies in the northwest end of the Tararua array and our splitting data set does not provide any constraints on that zone, but we compare it to see whether that zone is just a localized feature across the Hikurangi subduction zone and how anisotropy in the southwest region varies compared to northeast along the profile (Fig. 6.2).

Since we use three different phases for the anisotropy investigation, it is important to check whether these phases are sensitive to the same anisotropic zones at depth. This allows us to determine lateral variations with more confidence. *Marson-Pidgeon and Savage* [1997] found heterogeneous anisotropic media as inferred by frequency dependent splitting of SKS, ScS and S phases in the Wellington region SW of the

SAHKE transect. Therefore, we check for the sensitivity of splitting parameters to the period of the waves before discussing variations along the transect.

6.2.1 Frequency dependence and evidence for multi-layer anisotropy

Periods of the ScS, Tel. S, and SKS phases used in this study vary within the ranges of ~ 3 -7 s, ~ 8 -13 s, and ~ 9 -15 s, respectively. Only the SKS and Tel. S consist of long period energy. Filter bandwidths used to filter SKS and Tel. S range from 0.04 - 0.1 Hz, whereas ScS phases are band-passed at 0.1 and 0.3 Hz. *Marson-Pidgeon and Savage* [1997] noticed that ScS splitting recorded in Wellington region are independent of the filter. Because ScS phases have little long period energy we cannot use the same long period filters that we use for SKS/Tel.S phases.

Figure 6.5b-*i/ii* shows the δt variations with respect to the period of the S-phases recorded at SAHKE transect stations separately, and figure 6.5a-*i/ii* shows transect measurements together with the measurements at WEL station (Fig. 6.2). In both cases, δt of the good quality ScS and Tel. S phases show a clear increase in δt with periods up to 11 s. ϕ show no significant variation with the period (not shown). Frequency dependency is also not sensitive to the BAZ of the phases (Fig. 6.5a/b-*ii*). Tel.S phases with period $\gtrsim 11$ s and SKS phases show no clear evidence for frequency dependent splitting, suggesting that ScS ($T \lesssim 7$ s) and Tel.S ($T \lesssim 10$ s) phases with relatively short period energy are either sensitive to different length-scales of anisotropy or scattering by large-scale crustal heterogeneities (section 3.5.2; *Wu and Aki* [1988]). Varying anisotropic length-scales can distribute laterally and/or at depths. Modeling of the multi-layer anisotropy can be used to extract information on depth varying anisotropic layers, but this can be complicated by the lateral variations in the anisotropic structure. Fresnel zone arguments have been used by many studies (e.g. *Wirth and Long* [2010], *Rumpker et al.* [2003], *Marson-Pidgeon and Savage*

Table 6.2 Fresnel zone (FZ) approximations for the average frequencies of ScS, Tel. S and SKS phases at different depths (DP).

phase	ScS	Tel. S	SKS
avg.freq. [†] (Hz)	0.2	0.1	0.07
depth (km)	FZ width (km)		
5	15	25	32
10	24	36	46
100	70	100	122

[†]average frequencies of the ScS, Tel. S and SKS phases (Fig. 6.5)

[2004b]) to estimate the highest length-scale of anisotropy at a given depth that could be sampled by S-phases with different frequencies to produce frequency dependency. Fresnel zone (FZ) defines the boundary limits of the area at depth (L) from the receiver, where seismic waves are sensitive to subsurface structures. For plane waves this can be approximated from [Gudmundsson, 1996]: $FZ = \sqrt{2\lambda L + (\lambda)^2/4}$. If the different length-scales of anisotropy are smaller than FZ, then frequency dependent splitting is possible [Eakin and Long, 2013-in revision]. S-waves with short-period energy will be sensitive to smaller area than those with long-period energy, and this provides approximate constraints on the lateral variations in anisotropy. The approximate Fresnel zones widths at different depth for ScS, Tel. S and SKS phases that have been used in this study are given in table 6.2.

ScS phases, which have short period energy, consistently yield NNE-SSW fast orientation with smaller δt at all of the SAHKE stations and WEL station compared to SKS and Tel. S. This result is consistent with the previous observations in the surrounding region (section 6.2.2). It is likely that the fore-arc region of the southernmost Hikurangi subduction zone is characterized by a shallow crustal-anisotropic layer perhaps due to fluid-induced cracks or clay mineralization in the crust/sub-crust of the overriding Australian plate, and short-period ScS phases are sensitive to that structure. Clear frequency dependent anisotropy from both ScS and Tel. S with periods less than ~ 10 s suggests that this crustal anisotropy is most likely

characterized by different length-scales. According to table 6.2, ScS phases can be sensitive to structure with length scales as high as 24 km and Tel. S phases can be sensitive to the maximum length of ~ 36 km at crustal depths. This suggests that medium is most likely characterized by small-scale heterogeneities with maximum length-scales up to ~ 36 km. Nevertheless, the lack of frequency dependence from SKS suggests that those phases are mostly sensitive to the large-scale anisotropic structure (length-scale >100 s of km) that is possibly distributed in the upper-mantle depths and are less effected by the small length-scales of anisotropy.

0.5 - 1.5 s ScS delays are relatively large to be caused solely by a shallow-crustal anisotropic source. As suggested by *Gledhill and Gubbins* [1996], 1 - 2 s δt measured in the region north of Wellington and west of the SHAKE transect (Fig. 6.2) from Tararua array arises from ~ 300 km thick trench-parallel flow-governed anisotropic zone in the sub-slab region. Therefore, some of the ScS phases appear to reflect the typical upper-mantle anisotropy ($\delta t \sim 1 - 1.5$ s) and the smaller delays ($\sim 0.5 - 1$ s) could be due to frequency dependence and/or crustal re-splitting (section 3.5.1). Slight changes in ϕ (from NNE-SSW to NNW-SSE) may also be effects of these two scenarios (Fig. 6.7). If this is the case, large δt (>2 s) from long period SKS and Tel. S phases likely reflect an integrated effect of anisotropy along the ray path. However, the range of δt (1.8 - 3 s) from long period SKS phases implies that not all of the ray paths show the same cumulative effect of anisotropy. This suggests laterally varying or localized anisotropic zone/s perhaps above the sub-slab region (assuming that the slab itself is also governed by lateral variations in anisotropy). The majority of ϕ are oriented NE-SW ($10^\circ - 80^\circ$) throughout the area and thus it is possible that S-waves, which sample the anisotropic layers with approximately similar orientation of axis of symmetry, integrate the δt along the ray path rather than cancel out as a result of the path integration with competing fabric. The lack of frequency dependent anisotropy at long period phases (period >11 s) could also be an effect of this integration. The splitting sensitivity of smaller

periods (period ≤ 11 s) also suggests depth localization of anisotropy (or multi-layer anisotropy). If this is the case, layers with different anisotropic length-scales can be distributed at depths to produce frequency dependent splitting. The resolved multilayer models only have moderate fits to the observed splitting (section C.4), suggesting lateral inconsistencies within the stratified anisotropic structure or a more complex behavior.

Figure 6.5 δt as a function of the period of the different S-phases (stars - ScS, squares - Tel.S, diamonds - SKS). Figure 6.5a(i) and (ii) include the δt measurements at WEL station, whereas figure 6.5b(i) and (ii) exclude the δt measurements at WEL. Figures 6.5 a(i) and b(i) only display the δt variations with periods of the ScS and Tel.S phases. Good quality measurements are marked in black color and uncolored (open) symbols denote fair measurements. δt measurements in figures 6.5 a(ii) and b(ii) are color coded according to BAZ to show that the observed frequency dependent splitting is not a function of BAZ.

6.2.2 Lateral variations of splitting along the transect

The period versus δt analysis in section 6.2.1 (Fig. 6.5) and moderately resolved multilayer models (section C.4) imply stratified anisotropic structure with lateral complexities is plausible. The lack of frequency dependence from long period SKS and a wide range of δt (1.8 - 3 s) even from phases with similar periods also suggest that waves are not only sensitive to the multilayer anisotropy, but also to lateral variations in the anisotropic structure. Therefore, splitting parameters along the transect across the Hikurangi subduction zone were analyzed to detect lateral variations using the measurements made on the SAHKE transect and previous deployments that are within the 30 km distance either side of NW-SE aligned profile (Fig. 6.2).

Figure 6.6 Record sections of the filtered (bandpass at 0.04 - 0.1 Hz) Tel. S (a - event:10205053) and SKS (b - event:10181072) arrivals at different stations across the Hikurangi subduction zone. Phases are rotated into their fast (dashed-line) and slow polarization azimuths to display δt variations. Note that the SKS arrival (red in Fig.b) at T004 displays a smaller lag-time compared to T016 and T014 on the SHAKE transect, however Tel. S arrivals acquire approximately the same high δt along the transect.

In a few cases, the same events recorded at different stations yield good quality splitting measurements. The example record sections of Tel. S and SKS phases that are rotated into their individual fast and slow components are given in Figure 6.6(a&b) for comparison. We do not have an SKS-phase from the same event recorded on all transect stations that we could use to compare the δt variations along the transect in the record section, however Figure 6.6(b) indicates that the time lag between fast (red dashed-line) and slow (red bold-line) are smaller at T004 than at T016 and T020 (also see section 6.2.2). Tel. S phases do not show any lateral variations along the SAHKE transect (Fig. 6.6(a)).

Map views of the measurements made at each station are displayed in Figure 6.7. Measurements from ScS (blue), SKS (red), and Tel. S (yellow) are shown in the separate maps, and previous measurements (SKS-*Gledhill and Gubbins* [1996], ScS/Tel.S-*Marson-Pidgeon et al.* [1999]; *Marson-Pidgeon and Savage* [2004b], and *Greve et al.* [2008]) are marked in purple color. Circular histograms in each plot represent the ϕ measured at all the stations. ScS phases generally show consistent ϕ (NNE-SSW or NE-SW) and small δt at all the stations. The ScS delays also agree with the previous measurements at Tararua array stations (Fig. 6.8-A) and SNZO station (Fig. 6.7a), suggesting ScS phases are sensitive to a coherent anisotropic layer that is spatially distributed throughout the area. But there is a slight change in ϕ between stations in the eastern and western side of the Wairarapa fault. Most of the ScS measurements made in the eastern side are more NNE-SSW compared to NE-SW trend in the western stations (Fig. 6.7a), suggesting only wave-polarizations are slightly effected by the complexities in the region (Fig. 6.8-A).

Because of the consistency in δt and relatively small delays from ScS phases (Fig. 6.8A), we infer ScS splitting represents either background anisotropy or a coherent anisotropic layer in the lithosphere, and it is only slightly affected by the previously suggested lateral variations.

Figure 6.7 Map views of the splitting measurements plotted at the stations. An individual split bar shows the direction of ϕ and length represents the δt . ScS (blue), Tel.S (yellow) and SKS (red) measurements are shown in sub-figures 6.7a, 6.7b, and 6.7c respectively. Circular-histograms in each plot correspond to the estimated ϕ trend from each phase, and the number of measurements from each phase is denoted just above the histogram. In each sub-plot purple color bars indicate the measurements from previous studies (SKS-*Gledhill and Gubbins* [1996], ScS/*Tel.S-Marson-Pidgeon et al.* [1999]; *Marson-Pidgeon and Savage* [2004b] and *Greve et al.* [2008]). Backazimuthal coverage of each phase is shown in the world map.

The sensitivity of ScS phases to a coherent anisotropic layer with the different length-scale of anisotropy is possible due to smaller periods ($\lesssim 6$ s) of the ScS phases compared to SKS and Tel. S phases (Fig. 6.5). However, SKS and Tel. S seem to sample both background anisotropy and spatially varying anisotropic structure producing the range of δt values, which include typical upper mantle values mostly to the west (~ 1 - 1.5 s) and unusually high δt ($\gtrsim 2$ s; Figs. 6.8B/C), from both phases in the central-east. Figures 6.8C and 6.8B display the SKS and Tel. S measurements along the NW-SE aligned profile parallel to the SAHKE transect. Measurements from the Tararua array and LBLU stations (stations within 30 km distance from the transect) (circles in figures 6.8-A,B,C) are also plotted to examine the variations along the line.

Tel.S phases show an abrupt variation in δt (about 1.2 s δt change within ~ 5 km distance) between T022 and LTN6, dividing the profile into two anisotropic domains with distinct anisotropic properties (Fig. 6.8B). Note LTN6 and T022 stations are located in and around the Wairarapa fault (Fig. 6.2). Stations west of this anisotropic transition zone have an average Tel. S δt of $1.38 \text{ s} \pm 0.091 \text{ s}$, and the stations on the eastern side indicate an average Tel. S δt of $2.25 \text{ s} \pm 0.078 \text{ s}$ average Tel.S δt (Fig. 6.10). Both SAHKE and previous measurements from Tel.S phases have the same backazimuthal coverage ($280 - 340^\circ$), which rules out the effect of the BAZ (Fig. 6.8D) and minimizes the path effects. *Marson-Pidgeon and Savage [2004b]* modified the SC91 method to perform splitting from ScS /Tel S phases using transverse minimization method by calculating ϕ_p . They showed that the splitting parameters from both the modified method and eigenvalue minimization method give the same results regardless of the filtering. This confirms that the splitting variation along the transect profile is not due to different techniques used. It is also unlikely that the observed discrepancy between the eastern and western side of the Wairarapa fault is a measuring/processing artifact because: the previous SKS measurements at SNZO and LBLU are consistent with the SKS measurements at

Figure 6.8 Lateral variation in splitting parameters along the transect. Figures 6.8-A,B,C and D display the projected splitting measurement on to the SE-NW oriented profile line (Fig. 6.2) that is approximately parallel to the SAHKE transect and the SE-NW Tararua arm. Stations within 30 km distance from the profile have been used. Figures 6.8-A,B, and C display the variations from ScS, TelS and SKS phases separately. Circles in those plots represent the measurements from previous studies and fair measurements are indicated by the open-symbols. Note an abrupt change in TelS delays around LTN stations. Figure 6.8-D shows all the measurements, color coded according to event BAZ. Relatively small symbols represent the previous measurements.

nearby station WEL and the SAHKE stations on the eastern side (Fig. 6.7-c, Fig. 6.8-C), ScS δt are as same as previous δt estimations (Fig. 6.7-a, Fig. 6.8-A), and our splitting estimations in the LQT coordinate system using a different method yield similar splitting parameters (see section 6.2.3). Most of the broadband sensors used for the SAHKE experiments were CMG-40T seismometers with the standard response of 30 s - 50 Hz and two sites (LTN6 and T004) had CMG-3ESP sensors with the same standard response (table C.1). Previous Tararua array consisted CMG-3T broadband sensors with the standard response of 120 s - 50 Hz. Thus, all the seismometers across the region had a flat response over the 1- 15 s band that is strong in the phases used for this analysis (Fig. 6.5). Both previous and SAHKE sensors are broadband. We do not see a significant change in ScS splitting parameters between previous and SAHKE splitting data sets (Fig. 6.8-A). And also, SKS measurements in the southeastern end of SAHKE transect agree well with the previous SKS measurements in northwest (Fig. 6.8-C), implying that there is no significant effect due to two different instruments of the previous and SAHKE experiments. Therefore, instrument responses cannot explain the abrupt variation in δt , the increase in average station δt of SKS splitting (Fig. 6.9), and the scatter in the individual δt of SKS and Tel. S phases (Fig. 6.8-C&B) towards the NW end of the SAHKE transect, suggesting the presence of laterally varying anisotropic structure.

It is hard to provide accurate constraints on such a boundary zone from SKS phases as there are no δt measurements available at individual LTN stations and there are a limited number of previous measurements in the northwest profile (Fig. 6.8-D). SKS phases recorded on the SAHKE transect arrive at a range of BAZ (Figs. 6.7 and 6.4) and give a wide range of station delays (1 - 3.4 s) especially towards the NW end of the transect (Fig. 6.8-C), where the Wairarapa basin is located. T022, which is located in the vicinity of the Wairarapa fault, yields anomalously high delays from two measurements (2.8 - 3.2 s), but one of those measurements is rated

as marginal by the error criteria. When those measurements are averaged together with Tel. S measurements, the average δt recorded at T022 is about 2.6 s (Fig. 6.10) and this value is quite high to be caused only by a crustal or upper mantle (above and below the slab) source. We think that such a high value can be a product of coupled anisotropy from mantle-flow, lithospheric shear beneath the fault zone and the fault structure itself. Both Wairarapa fault and Wellington fault, which is located to the west of the Wairarapa fault (Fig. 6.2), are large-scale strike-slip faults in the dextral shear-belt of NI. A reflection study by *Henry et al.* [2013] along the same transect line found that the Wairarapa fault structure is more prominent than the Wellington fault. The Wairarapa fault acts as the western boundary limit of the Wairarapa basin. Gravity anomaly detected in Wairarapa region is interpreted as mainly due to density variation between Cenozoic sediments accumulated in the Wairarapa basin and Mesozoic basement rocks [*Hicks and Woodward*, 1978]. As constrained by these models, the thickness of the sediments overlying the Mesozoic basement reaches a maximum of ~ 3.2 km at the western edge of the basin close to Wairarapa fault (i.e., the basin becomes deeper towards the Fault). Well-developed sedimentary beddings within the basin can generate structural anisotropy in the shallow crust. If we assume 1.5 s δt is accumulated within the sub-slab region [*Gledhill and Gubbins*, 1996] and/or the upper-mantle above the slab, then ~ 1.1 s must have come from the anisotropic source that resides in the slab and/or crust. Splitting δt in the crust usually varies between 0.1 - 0.3 s [*Silver*, 1996, *Karalliyadda and Savage*, 2013]. In our case, an additional ~ 0.8 s δt must, therefore, be produced within the crustal depths or in the slab. Crust beneath a large strike-slip fault like the Wairarapa fault can be characterized by steeply dipping foliations, and such vertical foliations, can produce large delays up to 0.5 s (*Barruol et al.* [1997] and *Lloyd et al.* [2009]; section 2.4.1).

Figure 6.9 SKS δt variation along the SAHKE transect. The average station δt are plotted as red circles at the station position in the SE-NW profile (Fig. 6.2). The error bars denote the standard error of average δt at each station. Measurement at LBLU indicates the averages δt of previous study. The black line denotes the linear-fit to the data. Note the slight increase (statistically significant) in δt from SE to NW.

Figure 6.10 δt variation along the profile from Tel.S and SKS phases. SKS and Tel.S station averages are marked in red and yellow circles. The black-dotted line indicates the binned (~ 3.6 km wide) averages of the individual δt along the profile. The error bars denote the standard error of average δt at each station. The mean δt of SKS and SKS+Tel.S in the western and eastern sides of the Wairarapa fault are marked on the plot. The gray shaded area and the gray-dashed line indicate the Wairarapa basin and the fault, respectively.

If the anisotropy contribution from the slab is maximum 0.3 s and the orientation of the axis of symmetry in each layer is approximately similar, then the average δt of ~ 2.6 s at T022 (Fig 6.10) can be well explained by the different possible anisotropic contributions at depths in the region, and can be approximated with the model of vertical coherent deformation [Silver, 1996]. It is also possible that the adjacent Wairarapa basin is characterized by highly foliated structures, sedimentary beddings and/or in-filled materials with different anisotropic properties (e.g. clay minerals) that could produce high crustal contribution of anisotropy. As revealed by Kaviani *et al.* [2011], the basin structure itself and other small scale heterogeneities within the basin can create laterally varying structure (or isotropic velocity) that could affect the split S-waves differently, producing effective anisotropy. To produce effective anisotropy, length-scale of heterogeneities should be less than the dominant S wavelength. Therefore, the observed δt variation could be due to a laterally varying structure (or isotropic velocity) across the study region. Unlike Tel.S, SKS phases do not show a striking δt difference between the west and east side of the transition zone, as inferred by Tel.S splitting. There are not enough SKS splitting measurements available at LTN6 and LTW3 stations to provide robust constraints on such a variation. The slight increase in station averages of SKS δt (Fig. 6.9) and individual measurements from certain BAZ (Fig. 6.8-D) suggest some SKS delays are affected by both backazimuthal variations and the complexities beneath the station.

Figure 6.10 summarizes the average δt measurements at each station along the profile from SKS (red) and Tel. S phases (orange). The black-dotted line indicates the binned (~ 3.6 km wide) averages of the individual δt along the profile. There is a sharp change in δt around the NE-SW aligned Wairarapa fault. From SW to NE along the profile line this change is visible as a sharp increment in δt (~ 1.2 s increment within ~ 3 km distance) and the average δt appears to remain ~ 0.6 s larger east ($2.05 \text{ s} \pm 0.084 \text{ s}$) of the fault than to the west ($1.44 \text{ s} \pm 0.09 \text{ s}$). A similar

type of δt variation has been detected by [Rumpker *et al.*, 2003], across the Avara fault (transform fault) in the Dead Sea region from long-period (5-7 s) SKS phases. They noticed enlarged delays of 0.3-0.5 s within ~ 30 km distance of the fault. Rumpker *et al.* [2003] used the finite-difference approach by Rumpker and Ryberg [2000] to model the 2D anisotropic structure along the profile, inverting the observed wave field and optimizing the fit of observed and synthetic splitting parameters. The best fitting model suggests the variation is assumed to be caused by sub-horizontal mantle flow in a ~ 20 km localized area around the deep extension of the fault. Rumpker and Ryberg [2000] constrained the Fresnel zones (wave-sensitivity range) of teleseismic-waves with various periods at different depths using finite-difference modeling. According to Rumpker and Ryberg [2000], waves with 10 s period would have a wave-sensitivity range of ~ 100 km at 50 km and ~ 175 km at 400 km depth [see Rumpker and Ryberg, 2000, Fig.4]. In our case, the δt variation that is constrained from Tel. S waves ($T \simeq 10$ s) appears to be more abrupt. A change of 1.2 s in δt occurs within a distance as short as ~ 5 km (compare Tel.S δt at T022 and LTN in Fig. 6.10), suggesting a very shallow source of anisotropy and/or small-scale lateral variation in the anisotropic parameters [Favier and Chevrot, 2003].

As pointed out by Kaviani *et al.* [2011], short-scale variation in δt can also be due to spatially varying isotropic velocity and, thus, providing an interpretation based only on variation in anisotropic structure may be misleading. There is evidence for such velocity variations (e.g. high velocity lid on the subducting slab) as found by Gubbins and Snieder [1991] in the Wellington region. Therefore, future work on this study needs to be focused on the finite-difference modeling of the observed δt variations considering both anisotropic structure and isotropic velocity variations in order to provide accurate constraints on the shallow structure in the fore-arc region of the southernmost Hikurangi subduction zone. Also, to obtain robust constraints on these variations, previous data sets need to be re-evaluated as those splitting measurements were obtained from a different set of instruments using a small number

of events.

If this δt variation is solely due to lateral variation in anisotropic structure, it is most likely caused by the shallow (upper most crust) anisotropic structure that lies east of the Wairarapa fault and it may extend further east towards the trench as the observed δt in the east is relatively higher than the western side. Because of the slight increase in delays from SE to NW along the SAHKE line and high delays close to the fault (Fig. 6.9), it is also possible that most of the anisotropy is localized towards the NW end of the profile where the Wairarapa basin and fault are located. As detected by the stacked reflections from active-source first-arrivals recorded on the SAHKE I/II transect [Henry et al., 2013], the Wairarapa fault has been interpreted as a listric strike-slip (strike of N45°) fault with small dips (15°) at deeper levels (~ 15 km), but is exposed at the surface (east of the Tararua range) as a vertically dipping fault (dip $> 60^\circ$). Henry et al. [2013] also noticed several splay faults that are branching out from the deeper part of the Wairarapa fault beneath the Tararua range and are exposed in the Wairarapa basin area and around the east coast. The 10 km shot point separation around the Wairarapa fault, however, did not allow them to characterize detailed structure in the very shallow crust (≤ 5 km). The Wairarapa basin itself also contains several active dextral-faults that are aligned with the trend of the Wairarapa fault [Beanland, 1995].

The observed abrupt δt variation across the forearc Hikurangi subduction zone (Fig. 6.10) also correlates well with the terrane boundary between Pahau and Rakaia terranes, suggesting different deformation histories and/or anisotropic properties associated with the two terranes. This boundary is characterized by Esk Head Mèlange that resulted from intense deformation of accretionary sediments [Wandres and Bradshaw, 2005] and coincides with the dextral Wairarapa fault. Both Pahau and Rakaia terranes consist of Torlesse quartzofelspathic greywacke, but Pahau terrane is a younger Torlesse-greywacke terrane than Rakaia terrane. Pahau terrane mostly consists of Rakaia sediments recycled during the Cenozoic [Wandres and

Bradshaw, 2005]. As explained above, this region is in general characterized by dextral faulting (e.g. Wellington, Wairarapa, Dry river faults) and fault controlled basin structures and uplifted mountain ranges (e.g. Rimutaka and Tarua) (Fig. 6.2). The eastern boundary of the Rimutaka-Tararua ranges is marked by the Wairarapa fault and the ranges are cross-cut by Wellington fault to the west of Wairarapa fault (Fig. 6.2).

Considering all of these facts, it is plausible that the region east of the Wairarapa fault and the basin is governed by shallow structural anisotropy due to faulting and/or uplifted sedimentary bedding. Large scale vertically dipping strike-slip faults like Wairarapa may accommodate intensely developed foliations at crustal depths. Therefore, relatively large δt in the east of the Wairarapa region can be attributed to structural anisotropy due to intense foliations around the fault zones [*Bastow et al.*, 2007], sedimentary bedding and to the fault zones themselves. The scatter in δt in and around the Wairarapa basin may reflect the structural complexities and/or isotropic velocity variations of the material within the basin. However, there could also be other minor effects from deeper contributions such as the nature of the upper-plate and the accretionary sediments above the slab [*Eberhart-Phillips et al.*, 2005], changing slab dip, and bending related faulting structures within the slab. *Eberhart-Phillips et al.* [2005] noticed a sharp change in V_p/V_s ratios around the Wairarapa fault, as well as high V_p/V_s and low Q_p at 0-20 km depths in the eastern side of the fault [see *Eberhart-Phillips et al.*, 2005, Fig. 7d] that they interpreted as over-pressurized subducting sediments. High V_p/V_s ratios may also reflect an intense clay (e.g. talc, chlorite) mineralization, and such clay rich zones can be highly anisotropic (see section 2.7.2). We think that a shallow crustal source with high anisotropy is responsible for the observed rapid change in split delay times across this region. If it's due to clay minerals or sedimentary bedding within the Wairarapa basin, the thickness of this anisotropic zone could be approximately equal the thickness of the sediment infill (~ 3.2 km) of the basin.

6.2.3 Possibility of dipping anisotropic layers?

Splitting parameters have also been estimated in the ray coordinate (LQT) system (section 3.5.3) in order to compare the measurements on the horizontal plane and the inclined plane [Babuska and Plomerova, 1993, Sileny and Plomerova, 1996, Vecsey *et al.*, 2008]. If the geometry of the anisotropic layers are inclined and the ray paths have large incident angles, such a comparison would result in significant variation between two sets of splitting parameters. Modeling of splitting parameters at the SNZO station [Marson-Pidgeon and Savage, 2004a] in the Wellington region suggested the observed splitting fits better with a model with horizontal axis of symmetry that consists of two or more layers of anisotropy. Our comparison of splitting measurements in the LQT with the geographic coordinate system (RTZ) showed no significant variation in splitting parameters from the two methods (Fig. 6.11). The plunge of fast direction of all most all of the measurements are greater than 80° . This implies a nearly horizontal axis of symmetry (Fig. 6.12).

Figure 6.11 Comparison of splitting measurements from ScS (top), Tel.S (middle) and SKS (bottom) phases in RTZ (black) and LQT (gray) coordinate systems. The splitting parameters of ScS and Tel.S phases have been estimated from the eigenvalue minimization method, whereas SKS splitting parameters were determined from transverse minimization method.

Figure 6.12 Splitting measurements in LQT coordinate system are plotted at the earth quake location (ScS -blue, Tel.S - yellow, red - SKS). The common dips (from the vertical) of ϕ with approximately the same event-origin regions are marked in those regions. In all cases the dip varies between ~ 75 - 90° , indicating horizontal symmetry axis.

Song and Kawakatsu [2012] attempted to explain the splitting parameters observed in the Hikurangi subduction zone using slab-coupled asthenospheric flow model. This model determines the possible splitting variations in the presence of dipping anisotropy, presumably oceanic asthenosphere with orthorhombic symmetry. The predictions based on such a model can be more accurate and clear in the shallow subduction zones with low dipping angles ($\leq 15^\circ$). The model that they proposed for NZ (Fig. 6.13b/c) should accurately reproduce the observed splitting in at least the shallow subduction in the northern NI, if such dipping layers exist. We compare our splitting parameters plotted with respect to BAZ and incident angle (Fig. 6.13d) to check whether such a model can be used to explain the δt variations and the trench-orthogonal splitting.

Neither ϕ nor δt agrees with the model predictions, especially from the events that arrive at 30 - 210° BAZ. Previous measurements [*Marson-Pidgeon et al.*, 1999, *Greve and Savage*, 2009] from the same BAZ range also do not fully correlate with the model predictions. ϕ measured from 210 - 30° appear to fit with the predictions, but those directions are also in agreement with the directions that one would expect from the fore-arc and/or sub-slab trench parallel flow models. The few trench oblique ϕ in our splitting data set are unlikely an effect of dipping anisotropy as the observations strongly disagree with the model predictions in those particular regions (Fig. 6.13b). However, to test this proposed model precisely [*Song and Kawakatsu*, 2012], we need more measurements from 90 - 210° BAZs and with higher incidence angles (~ 20 - 35°).

Figure 6.13 Predicted splitting parameters (red bars in **b&c**) from the slab-coupled asthenospheric flow model [Song and Kawakatsu, 2012] and observed splitting (black bars in **b&c**) from Greve and Savage [2009] and Marson-Pidgeon et al. [1999]. Black arrow indicates the APM direction. **(d)** measurements from this study are grouped as the station position along the profile. WEL station data is shown separately. Variation in splitting parameters from SKS (red), Tel.S (yellow) and ScS (blue) phases with respect to BAZ and incidence angle (i_c) are plotted in the equidistant circular diagram. BAZ represents the radial position and i_c represents radius segments of the circular diagram. Note: δt measurements in **d** and **b&c** are scaled differently.

6.3 Conclusions

Shear-wave splitting anisotropy across the Hikurangi subduction zone in the southern NI from SKS, ScS and Tel.S reveal short-scale lateral variations in the anisotropic structure. The general trend of splitting anisotropy, however, agrees well with the previous studies around this area, implying trench parallel flow in the south of the Hikurangi subduction zone. A wide range of splitting δt and small scale inconsistencies in ϕ are evidence for spatially varying anisotropic structure across the fore-arc subduction settings. An increase in average SKS δt along the SE-NW profile and scatter in delays from both SKS and Tel.S phases towards the SW end of the SAHKE profile in the Wairarapa fault and basin area suggest an additional shallow anisotropy source(s) other than sub-slab mantle flow. The proposed model for the northern NI, which consists of highly anisotropic mantle wedge due to presence of melt-bands [Greve and Savage, 2009], cannot be adequately used to explain the observed lateral variations of the Hikurangi subduction zone in the southern NI. The δt values of 2 s - 3 s from Tel. S and SKS phases just east of the CVR in the northern NI (Fig. 1.1) [Greve et al., 2008] is consistent with the δt variations in the NW end of the SAHKE transect in the southeast of the NI.

We think that an enhanced anisotropic zone in crustal and/or sub-crustal depths is responsible for the observations in the southern NI. This zone is likely characterized by a crustal origin and associated with intense foliations around the Wairarapa fault, sedimentary layers within the Wairarapa basin and/or clay mineralization beneath the fault zones and basin area (Fig. 6.14). To produce large delays (2 s), these different sources of anisotropy must be coupled and deform in a coherent manner. The range of delays (~ 1.0 s - 3 s) also suggests the anisotropic body itself may contain heterogeneous anisotropy with different length scales, so that frequency dependent splitting is plausible. Multilayer modeling of the anisotropy suggests that these anisotropic heterogeneities can also be segregated into multiple layers, and

anisotropy in the individual-layer could be enhanced by anisotropy associated with the fault zones in that region. As constrained by Tel.S splitting, an abrupt variation in δt around the Wairarapa fault is presumably caused by such enhanced crustal and sub-crustal anisotropy around the fault, coupling with the multilayer anisotropy (section C.4). This δt change may also reflect the different deformation histories associated with the two different terranes either side of the Fault. Whether this structure is actually characterized by the higher strength of anisotropy compared to surrounding material or isotropic velocity variation due to laterally varying structure is yet to be investigated via finite-difference modeling.

Figure 6.14 Schematic diagram showing the possible anisotropic structure in the region. Here, the crustal faulting and high anisotropy that may be associated with faulting (e.g. shear zone), clay mineralization and/or layered sediment deposits just above the slab are coupled with the sub-slab mantle flow (circled T) to produce high anisotropy in the eastern side of the fault. The Wairarapa fault geometry is obtained from *Henrys et al.* [2013]. Note the vertical exaggeration.

Our study provides no direct evidence for plunging anisotropic structures, and supports the previous anisotropy modeling prediction of axis with horizontal symmetry anisotropy. The slab-coupled asthenospheric flow model [*Song and Kawakatsu*, 2012] is not entirely explained by our measurements in the southern

NI. Thus, the few well-constrained measurements that yield trench oblique (nearly trench orthogonal) directions, which are roughly parallel to APM, are unlikely a result of such a plunging structure. Departure from strong trench parallel splitting may reflect either source-side complexities, or far-field anisotropy that aligns with the APM direction (Fig. 6.2).

Splitting measurements on the SAHKE transect along with previous results suggest trench parallel flow in the sub-slab and perhaps in the fore-arc mantle wedge. NE-SW aligned upper-plate faulting, that is possibly associated with the transpressional deformation in the oblique subduction appears to be coupled with the trench parallel flow in the sub-slab mantle (Fig. 6.14). The NE-SW aligned Wairarapa fault and basin act as a main anisotropic boundary in the region. The fault itself and its splay faults that are distributed in the east of the fault may have a significant effect on the observed high δt by enhancing shallow structural anisotropy and/or shear zones beneath the faulting region.

Chapter 7

Concluding remarks

7.1 South Island on-land and offshore experiment

This study provides an extensive analysis of local/regional S-wave splitting parameters in the upper lithosphere of the transpressional plate-boundary zone in SI and its near offshore regions. The analysis used splitting parameters that were determined from a new seismic data set recorded on an enhanced seismograph network that consisted of both land and offshore seismographs. Estimated splitting measurements and the analysis provide a first-order constraint on the spatially and depth-varying anisotropic structure in to lithospheric depths of the SI.

Despite the complex nature of the local/regional splitting of high-frequency S-phases, grading criteria that have been applied to individual measurements, spatial averaging and delay time tomography allowed us to determine regions with spatial consistencies. These spatially consistent regions can be attributed to coherent sources of anisotropy. However, the limited depth-resolution and inhomogeneous distribution of seismicity and ray paths greatly influence the final solution of the anisotropic patterns. Thus the consistent patterns from spatial averages appear to correlate with the possible depth contribution of anisotropy in the region and the extent of those sources. To minimize such limitations, we have used different event selection criteria and additional studies to obtain information on the anisotropic structure.

Both on-land and offshore splitting measurements suggest that the anisotropy at shallow depths of the plate-boundary zone is not entirely controlled by the shear-induced anisotropy that is associated with the plate-boundary fault at depth, but it is

also governed by structure and stress-induced anisotropy. These different anisotropic sources appear to dominate in tectonically distinct regimes throughout SI, depending on the regional tectonics and depth distribution of seismicity (Fig. 4.11 and Fig. 5.8). In the northern SI, shear-wave splitting anisotropy is primarily controlled by shear-induced fabrics that are aligned parallel/sub-parallel to the plate-boundary and transitional faults, suggesting a broad zone ($\sim 100 - 150$ km) of lithospheric shear at depths as predicted by the thin viscous sheet model [Molnar *et al.*, 1999] and previous studies [Audoine *et al.*, 2000, Balfour *et al.*, 2005]. Because of the limited depth extent of earthquakes, the anisotropic structure resolved at some stations in the central SI was limited to maximum depths of ~ 70 km. Based on the spatially averaged ϕ , which are parallel to SH_{max} , we suggest that the upper-crustal anisotropy in the southern part of the northern SI and part of the central SI is mainly controlled by stress-induced anisotropy. This zone also extends into near offshore regions of the upper-west central SI. However, ϕ from deep S-waves (deeper than crustal depths) associated with Fiordland subduction suggest that the southern-central region is most likely a high-strain zone that is parallel to the plate-boundary (Fig. 4.11). This high strain zone has been interpreted as a lithospheric-shear zone that represents the deep extension of the plate-boundary fault zone [Karalliyadda and Savage, 2013]. If such a shear-zone exists, it should extend northwards, connecting the pervasive-shear zone in the northern SI and, therefore, likely underlies the shallow stress-induced anisotropy in the central SI.

From the spatially averaged ϕ at depths above 100 km, we suggest that this shear-induced zone could extend to a maximum ~ 130 km east of the plate-boundary at depths above 100 km in the central SI and north of the southern SI (4.4.4). As inferred from the seismic anisotropy at shallow depths (<40 km) in schist rock of the central SI, this shear-zone can be as narrow as ~ 30 km at crustal depths, suggesting that the width of this shear zone is most likely fanning out from the plate-boundary at depth. Because of the lack of deep seismicity in the central SI, our results do

not provide constraints on the depth extent of this shear zone. However, an average station δt of $\sim 0.1 - 0.4$ s (maximum ~ 0.7 individual delays) compared to 1.7 s SKS δt from previous studies in SI implies that this shear zone may penetrate down to the asthenosphere.

Varying delays and/or relatively low delays may result from frequency-dependent splitting due to variable length-scale anisotropy, variable anisotropic orientations (complex structure) with respect to S-wave propagation, or re-splitting of S-phases due to multiple-layers anisotropy and/or upper-crustal heterogeneities. We have provided evidence to support some of these proposed ideas, suggesting a complex and heterogeneous anisotropic structure in the region (section 4.4.2 and C.5). In addition to the above scenarios, we also speculate that the change in deformation kinematics to diffusion creep at upper lithospheric depths especially in the regions far from the plate-boundary, where absolute plate motion (APM) governed anisotropy dominates, is another possible explanation for the observed low delays and absence of plate-boundary shear throughout the SI. Dislocation creep typically activates in regimes with high stress/strain, high temperature conditions and large grain sizes. However, the activation of diffusion creep could result in little or no anisotropy. Stress conditions at plate boundaries are usually high (≥ 10 MPa), but if low temperature, low strain conditions prevail and/or the grain size of the medium is small, diffusion creep becomes more favorable [Poirier, 1990, Kameyama *et al.*, 1999].

Beyond the shear and stress induced anisotropic domains at shallow depths of the central South Island, splitting appears to correlate with the structural imprints from the tectonic episodes prior to Cenozoic deformation. Deep upper-lithospheric anisotropy in the southern South Island (both offshore and onshore) is consistent with the APM direction (NW-SE) away from the Fiordland subduction zone, but regions close to the subduction zone show a slight rotation in ϕ with N-S alignment, suggesting anisotropy is likely controlled by the combined effects of trench-parallel

flow/shear and flow in the asthenosphere (Figs. 5.9 and 5.12). In the eastern-part of the SI, splitting reflects asthenospheric flow derived anisotropy (either present-day flow or preserved).

The spatial analysis of OBS splitting measurements along with land measurements suggests that there is a broad zone of pervasive shear in the plate-boundary fault at least in the central-north SI and northern SI. Because of the limited deep earthquakes in the central South Island and possible frequency dependent and/or multilayer effects, the depth extent of this shear zone is not well constrained by both land and inland splitting measurements. Nevertheless, plate boundary parallel ϕ at some stations far way (~ 200 km) from the plate-boundary suggest, steeply curved ray paths from deep events (50 - 90 km) are likely sensitive to a shear zone in the upper lithosphere (above ~ 90 km). This implies that deformation distribution around the plate-boundary zone in South Island agrees with the deformation distribution predicted by the thin viscous sheet model (Fig. 1.2-*i*). However, detailed ray tracing needs to be done in order to fully discriminate between the thin viscous sheet and localized deformation models (Fig. 1.2) proposed for South Island.

Future splitting-anisotropy studies in this region need to consider the affects of dipping anisotropic layers on the observed anisotropy. Comprehensive modeling approaches can be used to demonstrate how ray geometry and the complex orientations of anisotropic structures can influence the in-situ splitting anisotropy. Our data do not provide constraints on the splitting anisotropy deeper than ~ 70 km in the on-land central SI, due to limitations in incidence angles and S-waves with long ray paths. However, a small-scale study can be used to determine the deeper structure in the central SI only using deep S-waves (>100 km) from the two subduction systems unless those phases are sensitive to the uppermost anisotropic structure. If those S-phases consist of both long period and high-frequency energy, such a study will also be able to test the proposed ideas of frequency dependent splitting or re-splitting of S-phases in the layered media. The

model-solution from the preliminary 3-D inversion study (section 5.3.3) can be improved by using different starting models, applying smoothing, and changing damping parameters. To understand the overall anisotropic structure of SI, this model can also be extended, including upper-most crack-induced anisotropy and S-wave splitting from core and regional S-phases.

7.2 Southern North Island experiment

We have explored the deep anisotropic structure in the south-east of the Hikurangi subduction zone in NI using different S-phases that sample the receiver-side anisotropy. The measurements have been integrated with the previous splitting parameters in the same region to examine the lateral variation in anisotropy across the inland fore-arc region of the Hikurangi subduction.

The results suggest that the general trend of ϕ are consistent with trench-parallel flow in the upper-mantle and in agreement with the previous measurements in the surrounding regions. The range of δt ($\sim 0.5 - 3.0$ s) and slightly varying ϕ across the south of NI suggest a laterally varying anisotropic structure. As inferred from the splitting variations across the profile, the upper-plate Wairarapa fault and basin area appear to be characterized by a distinct anisotropic structure that is possibly localized at crustal depths. Abrupt change in delays around this fault zone divides the region in to two distinct domains of eastern and western sides and the average δt in eastern side is ~ 0.6 s higher than that measured in the western side of the Wairarapa fault. This sharp δt change at the Wairarapa fault also coincides well with the terrane boundary between Pahau and Rakaia terranes, suggesting contrasting deformational or structural fabrics in two terranes. Stress and strain around the Wairarapa fault can be distributed asymmetrically owing to contrasting rock fabrics either sides of the fault. Thick sediment infill at crustal depths (≤ 3.2 km) in the western end of the Wairarapa basin in Pahau terrane is the most likely to produce

contrasting anisotropic (or deformational) fabric compared to Mesozoic Greywake basement terrane in Rakaia terrane.

These lateral variations and observed frequency dependence of splitting suggest that the shear wave splitting in the southern NI is governed in part by the laterally varying crustal contribution of anisotropy or velocity variations within the shallow crust. This shallow anisotropy together with the deep anisotropy may reflect multiple layers of anisotropy with small variations in ϕ , but with different length-scales. We find no clear evidence for dipping anisotropic layers within the subduction structure. Deep anisotropic layers may be characterized by lithospheric shear beneath faulting and mantle flow parallel to the trench. The observed large splitting ($\delta t > 2$ s) and similar ϕ imply crust-mantle coupling of anisotropy in the region.

The observed lateral variation needs to be tested in the future using modeling approaches to understand the configuration of the observed anisotropic structure. The 2-D finite-difference method [*Kaviani et al.*, 2011, *Rumpker and Ryberg*, 2000], which can model both variations in anisotropic parameters and isotropic velocity variations in the crust and mantle due to heterogeneous structure, appears to be a good way to obtain quantitative and accurate information (e.g. location, size, strength etc.) on the structure that is responsible for this change. With careful inspection of the data and the correction for instruments responses, teleseismic S-phases recorded on the 3-component short period array/transect across the south of NI can also be used to obtain a denser splitting data set along the transect. Such a dense splitting data set will provide robust constraints on the observed changes and simplify the finite difference modeling.

Appendix A

On-land investigation

A.1 Station information

On-land experiments retrieved data from both the GeoNet permanent network (26 stations) and four temporary deployments (table A.1). All the seismographs used were broadband sensors. Temporary deployments consisted of IRIS-PASSCAL instruments, and they were deployed by VUW in collaboration with the University of Colorado. Raw data was processed with the PASSCAL routine, and event were extracted from the SAC-records using the GeoNet catalog from 2009-2010. Antelope software was used to archive data and this was sent to the IRIS data center.

Problem: Polarity change in the N-component (horizontal) of the DCZ station in the GeoNet network was detected through SKS splitting analysis by *Zietlow et al.* [2013]. Events from this station were reanalyzed to account for this error. We noticed that there was $\sim 40^\circ$ clockwise rotation in the general trend of the uncorrected ϕ . The comparison between before and after correction of the measurements is given in figure A.1.

Figure A.1 The comparison of ϕ change before (error) and after (correct) correction for the polarity flip at DCZ. Note that the general trend of NNW is rotated into NNE ($\sim 40^\circ$ clockwise from the NNW direction) after correcting for the polarity flip.

Table A.1 Station information of the broad-band stations used for the on-land experiment.

deployment type	area	station	lat.°	lon.°	elevation (m)	sensor	geology	acquisition period
GeoNet (permanent stat.)	NSI	THZ	-41.76418	172.90508	747	Guralp CMG-3ESP T3B45	greywacke	2004-10
		QRZ	-40.82725	172.52901	263	Streckeisen STS-2 110017	schist	2004-10
		NNZ	-41.21882	173.37933	130	Kinematics FBA-ES-T 987	stiff-soil	2004-10
		DSZ	-41.74667	171.80449	650	Kinematics FBA-ES-T 999	greywacke	2004-10
		KHZ	-42.41598	173.53897	64	Kinematics FBA-ES-T 1001	greywacke	2004-10
		LTZ	-42.78335	172.27091	648	Guralp CMG-3ESP T3E85	greywacke	2004-10
	GSI	WVZ	-43.07603	170.73665	91	Guralp CMG-3ESP T3P70	granite	2004-10
		MQZ	-43.70774	172.65364	57	Guralp CMG-40T-60S T4636	andesite	2004-11
		CRLZ	-43.57641	172.62310	55	Guralp CMG-3ESP-Z T3738	basalt	2004-11
		OXZ	-43.32590	172.03830	360	Kinematics FBA-ES-T 986	greywacke	2004-11
		RPZ†	-43.71627	171.05376	453	Guralp CMG-3TB T3A63	rhyolite	2004-11
		LBZ	-44.38720	170.18433	438	Guralp CMG-3ESP T3G61	greywacke	2004-10
	SSI	FOZ	-43.53377	169.81539	54	Guralp CMG-3ESP T3G67	greywacke	2004-10
		JCZ	-44.07487	168.78540	1062	Guralp CMG-3ESP T3G78	greywacke	2004-10
		MSZ	-44.67498	167.92634	55	Guralp CMG-3ESP T3E65	granite	2004-10
		WKZ	-44.82866	169.01749	558	Guralp CMG-3ESP T3G77	stiff-soil	2004-10
		DCZ	-45.46634	167.15349	71	Guralp CMG-3ESP T3G80	gneiss/schist	2004-10
		ODZ	-45.04561	170.64453	274	Streckeisen STS-2 110038	schist	2004-10
VUW†		EAZ	-45.23268	169.30818	350	Guralp CMG-3ESP T3G79	schist	2004-10
		MLZ	-45.36817	168.11835	594	Guralp CMG-3ESP T3B46	greywacke	2004-10
		TUZ	-45.95558	169.63107	110	Guralp CMG-3ESP T3B66	schist	2004-10
		WHZ	-45.89404	167.94698	345	Guralp CMG-3ESP T3B60	greywacke	2004-10
		APZ	-46.83339	167.98878	601	Guralp CMG-3ESP T3B61	diorite	2004-10
		OPZ	-45.88596	170.59768	375	Guralp CMG-3ESP T3P66	basalt	2004-10
		PYZ	-46.16786	166.68066	277	Guralp CMG-3ESP T3P68	-	2004-10
		SYZ	-46.53848	169.13876	52	Guralp CMG-3ESP T3B59	sandstone	2004-10
		CASS	-43.01519	171.72678	770	Guralp CMG-3T T3E30	greywacke	2009-10
		FREW	-43.45200	171.90410	410	Guralp CMG-3T T35525	greywacke	2009-10
		KELY	-42.26340	172.22755	455	Guralp CMG-3T T3E63	schist	2009-10
		CROE	-42.33653	171.39842	342	Guralp CMG-3ESP T34974	sandstone	2009-10

†bore-hole station (azimuth=185°)

**Northern South Island

*Central South Island

§Southern South Island

†Temporary deployments in NSI and SSI by VUW in collaboration with Colorado State University

A.2 Event information of the A-grade splitting parameters

Event information of the A-grade best filter measurements are given in the table A.2. A-grade selection criteria and cluster analysis criteria are explained in section 3.2.3 and *Savage et al.* [2010].

Table A.2 Events yield A-grade best filter measurement**.

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
2165241	2004	001	-41.213	172.544	219.765	4.3
2166031	2004	003	-43.159	170.603	8.085	3.8
2166422	2004	004	-41.369	173.372	93.746	3.3
2166642	2004	005	-41.342	173.052	93.820	3.1
2166804	2004	005	-45.604	167.031	9.000	4.0
2167639	2004	007	-45.083	167.425	93.908	4.2
2168610	2004	009	-43.434	170.389	5.000	4.1
2168698	2004	010	-44.580	168.766	5.000	3.3
2169096	2004	011	-40.647	173.867	85.919	3.3
2169162	2004	011	-41.324	172.415	5.000	3.8
2169838	2004	013	-42.339	174.037	12.000	3.5
2169945	2004	013	-45.248	167.022	11.653	3.6
2171083	2004	016	-41.858	174.126	12.716	3.9
2171148	2004	016	-42.973	172.393	5.000	3.5
2172290	2004	019	-42.081	173.787	31.550	3.1
2172542	2004	020	-42.366	174.084	7.337	3.7
2172782	2004	021	-45.224	166.916	26.699	3.8
2173116	2004	021	-42.226	173.055	11.994	3.3
2174766	2004	024	-42.300	174.018	39.201	3.3
2175242	2004	025	-45.435	167.156	103.872	3.7
2175527	2004	026	-44.625	168.189	85.499	3.5
2176121	2004	027	-43.654	173.423	17.183	3.1
2176328	2004	028	-41.640	174.782	31.094	3.1
2176454	2004	028	-45.151	167.103	13.158	3.6
2176479	2004	028	-44.615	167.289	16.160	3.1
2177341	2004	029	-42.452	172.943	5.000	3.2
2177405	2004	029	-43.508	172.668	8.419	3.1
2177589	2004	029	-41.724	172.618	89.887	3.0
2179414	2004	031	-45.181	166.992	25.744	3.7
2179877	2004	032	-42.954	171.931	12.000	3.2
2180565	2004	034	-45.582	167.072	11.251	4.2
2180670	2004	034	-41.010	172.853	216.869	3.6
2180744	2004	034	-42.636	173.172	33.000	3.0
2181094	2004	035	-41.386	173.071	97.351	3.3
2183311	2004	041	-45.088	167.040	19.148	4.1
2183966	2004	041	-45.219	166.960	26.935	3.6
2184019	2004	042	-45.609	167.120	84.066	3.8
2184403	2004	042	-42.193	174.772	33.785	4.6
2184814	2004	043	-45.275	166.939	24.493	3.7
2185761	2004	045	-43.079	173.369	23.454	3.2
2185957	2004	046	-44.541	168.674	12.000	3.0
2187002	2004	048	-45.305	166.963	24.012	3.9
2187366	2004	049	-44.995	167.541	129.781	3.7
2187501	2004	049	-43.198	173.488	12.000	3.0
2187515	2004	049	-40.848	174.257	61.025	3.1
2187735	2004	050	-43.415	173.134	15.447	3.4
2188069	2004	050	-44.531	168.051	18.045	3.1
2190336	2004	055	-42.058	173.164	9.675	3.7
2191681	2004	058	-45.379	167.208	108.854	4.1
2191915	2004	058	-45.067	167.429	97.054	3.7
2192455	2004	059	-45.243	166.929	23.402	4.9
2192669	2004	060	-45.238	166.938	24.664	4.4
2192783	2004	060	-41.142	173.001	129.925	3.7
2192878	2004	060	-45.242	166.919	24.545	3.6
2193185	2004	061	-42.029	171.586	12.000	3.1

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
2197549	2004	070	-42.723	172.800	33.000	3.1
2197714	2004	071	-45.129	167.438	80.991	3.6
2198118	2004	071	-44.033	169.618	33.000	4.1
2198446	2004	072	-43.160	173.825	21.924	3.2
2198750	2004	073	-40.978	174.399	53.554	3.2
2198958	2004	073	-44.181	168.598	5.000	3.1
2199054	2004	074	-45.269	166.685	19.673	4.0
2199408	2004	074	-42.297	173.343	38.842	3.0
2200334	2004	076	-41.891	174.017	19.128	4.5
2200400	2004	076	-41.887	174.012	18.021	3.0
2200660	2004	077	-42.469	173.850	27.597	3.5
2201439	2004	079	-44.045	168.492	5.000	3.0
2201726	2004	079	-45.200	167.349	119.236	3.6
2201956	2004	080	-45.231	166.958	22.207	3.5
2202855	2004	082	-43.526	171.187	5.000	3.1
2205598	2004	088	-41.654	173.875	15.683	3.7
2207974	2004	092	-45.173	167.069	23.692	3.8
2208003	2004	092	-45.196	167.530	130.698	3.5
2208493	2004	009	-43.460	170.389	5.000	4.4
2208792	2004	093	-45.130	167.405	117.216	4.8
2210424	2004	096	-41.389	172.853	115.079	3.6
2210987	2004	097	-42.636	172.155	7.476	4.4
2213304	2004	102	-45.241	166.924	23.010	4.2
2216109	2004	109	-41.463	174.974	25.104	3.6
2217019	2004	111	-41.359	174.198	35.148	3.1
2217222	2004	111	-41.291	174.773	47.265	3.4
2219145	2004	115	-41.394	173.284	85.142	3.1
2219507	2004	116	-44.624	168.237	71.484	3.4
2219741	2004	117	-45.141	167.322	120.638	3.8
2220857	2004	117	-41.788	171.898	14.416	3.3
2221137	2004	118	-45.043	167.424	93.405	3.9
2221191	2004	118	-44.086	168.577	12.000	3.0
2222177	2004	119	-41.148	172.534	5.000	3.3
2222853	2004	121	-42.376	174.494	34.109	3.1
2226065	2004	127	-41.419	172.971	106.561	3.6
2226081	2004	127	-45.133	167.546	90.905	3.6
2226251	2004	127	-45.174	167.014	20.000	3.5
2229236	2004	133	-44.978	167.489	121.836	3.7
2230096	2004	135	-41.624	172.857	86.515	4.0
2230781	2004	136	-41.730	171.984	7.500	4.1
2232751	2004	139	-40.608	173.854	94.019	3.3
2233748	2004	141	-40.822	173.715	95.601	3.4
2234048	2004	141	-45.154	167.462	112.262	3.8
2235298	2004	144	-42.023	172.830	81.578	3.5
2235722	2004	144	-42.739	172.832	15.580	3.0
2236446	2004	146	-44.950	167.595	67.560	4.2
2237498	2004	148	-45.061	167.026	12.000	3.8
2237811	2004	148	-40.836	172.964	221.514	4.1
2238221	2004	149	-44.521	168.288	5.000	3.0
2238715	2004	150	-41.344	174.375	42.394	3.4
2238854	2004	150	-45.241	167.345	102.035	3.9
2240587	2004	153	-45.149	167.290	114.989	4.0
2241544	2004	155	-41.408	172.890	98.179	3.6
2241772	2004	155	-45.082	167.383	104.494	4.6
2242087	2004	156	-41.456	172.836	99.662	3.2
2243719	2004	159	-43.561	172.220	20.095	3.5
2248250	2004	167	-41.250	172.648	202.426	3.9
2252524	2004	175	-42.171	173.926	31.620	3.5
2252734	2004	175	-45.064	167.414	115.608	3.6
2255507	2004	181	-43.330	171.866	9.007	3.1
2256302	2004	182	-44.934	167.560	77.679	3.8
2256750	2004	183	-45.232	166.912	12.000	3.8
2257665	2004	185	-45.087	167.023	20.000	4.4
2257908	2004	186	-45.300	167.246	68.488	3.6
2258122	2004	186	-42.203	173.886	19.785	4.0
2263205	2004	194	-42.822	172.443	9.272	3.4
2265342	2004	197	-42.816	171.360	5.000	3.1
2265509	2004	198	-45.058	167.113	5.000	3.8
2266037	2004	199	-40.962	174.926	31.764	3.3
2266184	2004	200	-41.212	173.381	94.991	3.4

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
2267941	2004	203	-43.358	171.930	9.457	3.4
2268502	2004	204	-40.564	174.290	74.992	3.4
2268698	2004	204	-41.252	172.640	192.408	3.7
2269062	2004	205	-40.530	173.674	127.693	4.6
2269844	2004	207	-42.697	173.667	35.883	3.1
2271646	2004	210	-44.563	168.363	5.000	3.2
2272438	2004	212	-41.153	174.598	60.205	4.6
2272755	2004	213	-41.810	172.737	87.062	3.2
2272849	2004	213	-41.647	173.935	8.502	3.1
2302197	2004	276	-42.268	172.738	11.423	3.5
2302434	2004	276	-42.333	173.145	5.000	3.4
2303253	2004	278	-42.383	172.818	12.136	3.4
2303625	2004	279	-41.508	173.386	69.272	3.1
2304495	2004	281	-45.143	166.954	25.343	3.8
2306884	2004	286	-45.054	167.858	12.000	4.1
2307072	2004	287	-44.266	168.661	5.000	3.1
2307605	2004	288	-40.518	174.306	86.071	3.3
2307840	2004	288	-45.151	167.348	109.969	3.5
2308266	2004	289	-42.578	171.911	5.000	3.3
2308271	2004	289	-40.702	174.566	72.498	3.1
2308546	2004	290	-41.405	173.709	58.294	3.4
2310945	2004	295	-43.173	173.513	22.207	3.4
2311679	2004	297	-41.256	172.700	185.281	4.2
2311774	2004	297	-45.182	167.303	125.062	3.7
2312015	2004	298	-45.346	166.817	12.000	3.7
2312381	2004	298	-42.189	174.228	33.000	3.6
2312482	2004	299	-40.686	173.916	89.276	3.2
2313500	2004	301	-44.311	168.346	20.613	3.1
2313713	2004	302	-45.126	167.057	20.000	3.9
2314163	2004	303	-41.282	172.569	206.230	3.5
2315087	2004	305	-43.068	172.036	5.000	3.1
2315598	2004	306	-43.006	172.825	27.545	3.6
2316723	2004	308	-45.159	167.260	114.188	3.6
2316812	2004	308	-40.925	172.903	208.856	3.5
2318753	2004	312	-40.504	173.467	142.411	4.3
2318769	2004	312	-40.647	173.628	110.160	3.8
2320355	2004	316	-41.723	172.201	94.633	3.4
2320849	2004	317	-44.528	167.397	5.000	3.3
2320957	2004	317	-41.012	173.790	83.954	3.1
2321673	2004	318	-41.664	174.242	9.413	3.4
2321906	2004	319	-40.998	172.434	5.000	3.4
2321935	2004	319	-41.546	173.511	64.303	3.5
2322081	2004	319	-44.595	168.214	85.843	3.9
2322299	2004	320	-41.208	174.004	53.002	3.2
2323155	2004	321	-45.639	166.985	9.000	3.6
2324239	2004	324	-40.966	174.639	61.547	3.1
2324418	2004	324	-44.589	168.240	83.787	3.6
2325873	2004	327	-45.206	166.960	12.623	4.0
2325951	2004	327	-45.069	166.963	5.000	4.1
2326696	2004	023	-41.401	172.222	5.000	3.1
2328987	2004	333	-42.605	173.536	39.294	3.7
2329087	2004	333	-40.985	172.432	5.000	3.8
2329697	2004	334	-41.278	172.661	174.165	3.7
2329763	2004	334	-41.014	172.916	186.242	3.7
2329949	2004	335	-44.081	168.754	8.169	3.1
2331522	2004	337	-43.061	170.847	5.957	3.3
2336444	2004	346	-45.127	167.415	124.185	4.5
2337155	2004	347	-41.705	172.847	79.358	3.4
2337190	2004	347	-41.497	174.130	35.995	3.0
2337656	2004	348	-40.955	174.601	42.325	3.6
2338463	2004	349	-41.647	174.276	14.946	3.1
2340108	2004	352	-44.038	172.908	26.398	3.6
2340467	2004	353	-45.066	166.818	5.000	4.0
2341023	2004	355	-44.691	167.957	94.786	3.3
2341413	2004	356	-44.523	168.465	5.000	3.1
2341695	2004	356	-42.412	172.301	5.000	3.2
2341801	2004	357	-45.144	167.370	112.310	4.7
2342689	2004	359	-45.227	170.068	5.000	3.8
2342705	2004	359	-45.163	167.463	109.112	3.7
2342814	2004	359	-42.873	172.940	14.281	3.4

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
2343092	2004	360	-45.166	167.347	83.699	3.7
2343259	2004	360	-44.091	168.797	8.681	3.0
2343713	2004	362	-44.949	167.089	5.000	3.5
2344142	2004	363	-44.410	168.200	5.000	3.6
2344228	2004	363	-41.290	172.562	212.870	3.8
2344284	2004	363	-43.978	169.321	5.000	3.9
2344960	2004	364	-45.101	167.026	5.000	3.5
2345202	2004	365	-45.057	167.432	109.337	4.7
2345474	2004	366	-41.253	174.507	55.374	3.0
2345598	2004	366	-41.311	172.773	136.783	3.7
2348362	2005	007	-42.815	173.229	33.000	3.5
2348447	2005	007	-43.505	170.717	6.288	3.5
2352265	2005	016	-44.625	166.971	5.000	3.7
2352652	2005	017	-45.445	166.825	20.946	3.5
2352729	2005	017	-45.001	167.479	97.802	4.0
2353253	2005	018	-45.090	167.427	86.901	4.2
2355829	2005	024	-45.129	166.917	20.000	4.6
2355961	2005	024	-41.307	172.719	146.361	3.6
2360692	2005	035	-44.587	168.173	84.153	3.9
2361453	2005	037	-41.410	172.962	111.823	3.5
2361698	2005	038	-45.027	167.540	77.697	4.0
2363065	2005	041	-47.039	166.319	33.000	4.1
2364350	2005	044	-45.330	166.922	21.376	5.1
2366807	2005	049	-41.423	172.861	104.594	3.9
2366974	2005	050	-45.229	167.312	61.399	3.7
2367019	2005	050	-45.123	167.392	87.299	3.7
2367324	2005	051	-45.155	167.382	122.686	3.5
2368297	2005	053	-40.830	173.155	172.635	3.5
2369901	2005	057	-40.515	173.549	144.554	4.2
2371160	2004	109	-44.554	167.763	11.212	3.8
2372178	2005	062	-40.720	173.244	170.675	3.5
2373267	2005	065	-41.297	172.806	138.617	3.7
2373531	2005	065	-41.289	172.519	212.966	3.7
2374145	2004	114	-44.959	167.093	22.682	3.6
2375614	2005	070	-45.082	167.538	114.894	3.5
2376410	2005	072	-45.072	167.488	115.635	3.5
2376477	2005	072	-41.267	172.591	202.700	3.8
2378677	2005	078	-45.312	166.999	22.063	3.7
2379429	2005	079	-41.292	172.576	192.780	4.0
2382229	2005	085	-45.001	167.526	93.320	4.1
2388715	2004	127	-45.050	167.581	114.912	3.7
2388847	2005	101	-41.650	174.550	34.607	4.8
2391405	2005	106	-44.350	169.530	11.858	4.8
2393740	2005	111	-45.153	167.377	116.756	3.8
2395344	2005	115	-45.878	166.865	86.948	5.3
2397298	2005	119	-44.093	169.183	58.931	3.2
2397632	2005	120	-40.992	172.889	192.719	3.6
2398948	2005	123	-43.932	168.970	12.000	3.5
2402919	2005	132	-41.416	172.809	109.481	3.6
2403463	2005	133	-45.014	167.427	84.950	4.2
2403620	2005	133	-44.944	167.458	100.371	3.7
2404541	2005	135	-45.075	167.395	118.371	3.9
2407258	2005	141	-45.311	166.874	27.534	3.8
2408156	2005	143	-41.451	174.368	30.826	4.4
2411621	2004	186	-45.142	166.927	12.000	3.9
2412564	2005	153	-44.608	168.209	86.236	3.5
2418341	2005	166	-44.626	168.283	86.070	0.0
2419267	2005	168	-41.242	172.662	218.667	3.8
2421529	2005	173	-41.336	172.539	197.634	3.8
2422085	2005	174	-41.982	173.892	12.000	3.5
2422240	2005	175	-44.853	167.347	54.200	3.8
2422650	2005	175	-41.309	172.551	200.614	3.7
2425170	2004	201	-41.708	172.475	11.515	3.5
2426106	2005	182	-45.103	167.454	119.185	3.9
2432882	2005	196	-41.099	173.512	105.670	4.2
2432954	2005	196	-42.406	173.890	16.441	3.9
2433378	2005	197	-44.571	168.291	87.890	0.0
2433564	2005	198	-45.045	167.403	86.817	3.6
2438565	2005	208	-41.791	172.395	89.152	3.6
2439182	2005	210	-45.065	167.380	88.450	4.0

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
2439371	2005	210	-41.004	173.832	68.969	4.1
2439596	2005	211	-47.108	166.217	33.000	3.5
2439652	2005	211	-41.457	174.194	35.214	4.0
2442391	2005	217	-41.568	173.068	77.608	3.7
2446969	2005	227	-44.481	168.199	87.498	3.2
2448668	2005	231	-41.673	172.818	79.869	3.8
2449310	2005	233	-41.550	174.763	29.484	3.7
2452869	2005	241	-41.257	172.626	209.444	3.8
2463808	2005	266	-44.594	168.177	85.846	3.2
2466872	2005	274	-43.265	170.785	5.000	3.7
2470940	2004	276	-42.336	173.137	5.000	3.8
2472594	2005	286	-44.733	167.364	28.880	0.0
2474138	2005	290	-41.292	173.992	48.391	3.9
2475493	2005	293	-45.139	167.513	123.548	3.5
2475730	2005	294	-45.943	166.973	100.194	3.5
2477338	2005	297	-45.880	166.897	84.383	4.7
2477589	2005	298	-41.341	172.893	125.677	3.6
2478135	2005	299	-44.976	167.503	141.369	4.2
2480719	2005	305	-42.193	173.468	33.000	3.7
2481459	2005	307	-45.213	167.364	74.090	3.7
2484883	2005	314	-40.992	172.947	185.232	3.5
2485606	2005	316	-41.240	172.853	150.868	3.5
2486372	2005	317	-46.092	166.841	108.171	4.0
2486404	2005	318	-45.300	167.289	68.224	3.5
2486477	2005	318	-41.330	172.562	188.748	3.9
2487262	2005	319	-41.549	174.316	27.022	3.9
2488793	2005	323	-44.863	167.684	77.032	3.5
2489083	2005	324	-42.680	172.906	33.000	0.0
2489325	2005	324	-42.682	172.875	33.000	0.0
2490144	2005	326	-44.735	167.891	99.940	0.0
2491398	2005	329	-44.632	168.206	86.610	0.0
2492944	2005	332	-45.450	166.768	22.023	3.8
2494342	2005	335	-45.139	167.444	113.861	3.7
2495389	2005	338	-41.788	173.836	37.268	3.6
2497818	2005	343	-41.431	173.311	79.862	4.2
2498619	2005	345	-41.016	172.846	204.702	4.0
2499328	2005	347	-41.096	174.862	29.746	4.6
2501023	2005	351	-41.244	172.627	215.232	3.6
2501759	2005	353	-45.108	167.453	118.745	3.9
2503622	2005	357	-45.251	166.899	23.898	4.5
2506620	2005	364	-41.184	173.267	106.339	4.2
2509731	2006	005	-44.738	167.808	76.154	3.2
2510085	2006	006	-41.103	174.670	32.864	4.0
2514232	2006	016	-40.480	173.428	154.684	4.6
2516214	2006	021	-41.049	173.942	63.417	3.7
2518655	2006	026	-45.987	166.864	99.089	3.5
2519069	2006	027	-41.015	174.087	59.417	3.8
2520364	2006	031	-45.469	167.052	68.482	3.5
2520592	2006	031	-41.102	174.682	32.430	3.6
2520889	2006	032	-41.056	173.528	118.189	3.9
2521211	2006	033	-41.120	172.960	155.757	3.5
2522760	2006	037	-41.279	172.529	213.133	3.6
2522859	2006	037	-45.771	167.051	105.195	4.4
2526910	2006	047	-41.421	174.656	53.036	4.0
2528587	2006	051	-45.108	167.394	66.537	4.2
2529755	2006	054	-41.430	173.972	43.275	3.7
2533155	2006	061	-41.715	174.650	28.259	3.9
2533834	2006	063	-45.026	167.422	112.797	4.7
2534340	2006	064	-44.878	167.603	73.675	3.6
2536786	2006	069	-41.230	172.589	201.429	3.6
2537149	2006	069	-45.289	167.183	85.405	4.6
2538232	2006	072	-40.681	174.466	79.016	3.5
2538892	2006	073	-41.612	173.586	58.367	3.9
2540415	2006	076	-44.727	167.901	97.290	3.2
2543160	2006	082	-45.110	167.415	113.802	3.5
2543194	2006	082	-40.898	174.468	47.491	4.1
2544672	2006	085	-44.567	168.137	82.278	3.6
2546014	2006	088	-42.552	173.714	24.512	3.2
2546560	2006	089	-41.389	174.654	53.828	3.7
2547781	2006	092	-41.207	174.777	52.669	4.4

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
2549731	2006	097	-41.117	173.032	156.406	4.0
2551363	2006	099	-41.270	172.781	142.044	3.7
2551460	2006	100	-41.245	172.790	148.244	3.6
2555839	2006	109	-45.394	166.874	24.911	4.3
2558232	2006	113	-45.126	167.358	84.280	4.3
2558585	2006	114	-45.280	167.194	110.134	3.9
2559763	2006	117	-43.299	170.508	5.000	4.2
2559903	2006	117	-43.298	170.506	5.000	4.1
2562489	2006	123	-45.108	167.494	116.478	3.7
2563468	2006	124	-45.158	167.391	118.291	3.7
2565311	2006	128	-45.117	167.397	125.476	4.3
2567466	2006	133	-44.953	167.538	97.879	3.6
2567689	2006	134	-41.113	174.566	40.056	3.7
2567873	2006	134	-40.943	174.454	59.406	5.0
2568379	2006	135	-41.306	172.815	140.379	3.8
2568890	2006	137	-40.605	174.353	67.052	3.8
2570686	2006	141	-41.367	173.122	95.110	3.7
2572359	2006	144	-45.072	167.474	111.482	3.5
2573232	2006	146	-45.063	167.439	114.176	3.6
2573533	2006	147	-45.085	167.420	90.624	3.9
2574820	2006	149	-45.077	167.458	84.538	3.9
2576189	2006	152	-44.843	167.702	72.853	3.5
2578100	2006	156	-44.606	168.180	81.094	4.1
2579538	2006	159	-40.293	173.522	174.344	3.9
2580443	2006	161	-40.892	174.603	46.415	4.7
2586781	2006	175	-41.771	173.089	95.596	3.6
2592468	2006	187	-45.167	167.292	114.247	4.3
2597464	2006	198	-45.131	167.430	65.116	3.6
2598293	2006	200	-44.666	167.937	95.734	3.6
2598839	2006	201	-44.392	170.237	9.498	3.4
2599831	2006	204	-40.600	173.286	172.748	5.5
2599966	2006	204	-41.277	172.773	142.145	3.7
2600075	2006	204	-45.127	167.395	132.827	4.0
2600789	2006	206	-41.273	172.810	128.647	3.5
2601210	2006	206	-45.102	167.417	110.845	4.0
2602165	2006	208	-42.662	173.195	33.000	3.5
2602486	2006	209	-41.202	174.615	57.818	3.6
2603539	2006	211	-40.397	173.244	213.894	3.9
2608311	2006	221	-45.122	167.322	94.803	4.3
2609143	2006	223	-41.463	174.209	21.458	3.7
2609766	2006	225	-41.764	172.652	94.128	5.9
2613953	2006	233	-44.471	168.286	5.000	3.8
2615935	2006	237	-43.597	170.621	5.000	3.6
2616404	2006	238	-45.062	167.458	129.042	3.6
2616532	2006	239	-46.079	168.019	18.577	3.5
2619948	2006	246	-41.015	172.872	214.607	3.6
2623315	2006	254	-41.271	172.709	183.045	5.2
2624333	2006	256	-41.187	173.465	94.649	4.2
2624977	2006	257	-42.011	174.089	21.373	3.9
2625245	2006	258	-40.965	174.478	48.504	5.1
2625655	2006	259	-45.149	167.359	92.730	4.0
2628986	2006	266	-40.964	174.820	32.156	3.9
2630049	2006	268	-41.215	172.836	170.210	3.9
2630581	2006	269	-41.360	172.565	189.341	4.2
2630758	2006	270	-45.374	167.287	108.065	3.5
2632675	2006	274	-41.245	172.732	143.735	3.5
2634680	2006	279	-41.240	172.750	167.250	3.7
2635139	2006	280	-45.095	167.435	112.653	3.6
2637385	2006	284	-44.718	167.870	96.248	3.1
2639387	2006	288	-44.847	166.920	33.000	4.0
2641098	2006	292	-41.184	172.791	178.461	3.7
2643586	2006	297	-45.341	167.161	96.678	4.2
2647980	2006	307	-43.655	169.695	5.000	3.9
2648753	2006	309	-47.034	166.211	33.000	3.5
2653841	2006	319	-45.101	167.340	84.302	4.3
2654099	2006	320	-41.218	172.695	202.560	3.6
2654530	2006	320	-41.037	174.173	57.658	4.6
2654849	2006	321	-41.062	174.334	71.594	5.5
2655095	2006	322	-41.364	172.789	131.216	3.6
2655189	2006	322	-43.318	172.523	35.182	3.2

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
2655520	2006	322	-42.988	172.102	5.000	4.2
2664838	2006	341	-44.660	167.621	26.118	3.3
2666076	2006	344	-45.031	167.512	72.623	3.5
2666824	2006	346	-45.153	167.383	124.553	4.1
2667016	2006	346	-44.713	168.043	83.294	3.3
2672673	2006	357	-44.825	167.643	82.576	3.7
2674538	2006	361	-45.184	167.321	114.743	3.5
2676137	2006	364	-43.394	172.548	22.401	3.0
2676169	2006	364	-44.602	168.137	81.162	3.6
2678505	2007	005	-41.256	172.733	156.048	4.4
2679334	2007	007	-44.599	168.028	92.554	3.3
2680518	2007	009	-44.581	168.181	87.087	3.0
2681421	2007	011	-45.074	167.442	116.071	4.1
2685645	2007	021	-45.416	167.205	102.262	3.7
2690111	2007	031	-45.072	167.137	29.157	3.6
2692934	2007	038	-41.016	174.477	51.571	4.7
2698711	2007	049	-45.446	167.045	75.588	3.7
2699618	2007	051	-41.128	174.145	49.820	4.4
2699849	2007	052	-45.323	167.251	118.705	4.7
2702307	2007	057	-40.921	173.820	108.029	4.5
2704689	2007	062	-45.549	166.971	68.565	4.3
2707519	2007	068	-46.003	166.818	100.045	3.7
2707727	2007	069	-41.196	173.030	132.392	3.9
2711310	2007	076	-41.303	172.566	190.406	3.7
2713754	2007	081	-40.383	173.470	177.212	4.0
2713982	2007	081	-42.007	172.714	78.945	4.7
2718373	2007	090	-44.623	168.175	79.191	3.3
2720328	2007	095	-41.251	172.603	192.920	3.6
2722537	2007	099	-44.738	167.718	58.479	3.3
2724187	2007	103	-45.095	167.436	122.909	4.2
2724360	2007	103	-44.703	167.222	29.669	3.4
2726034	2007	107	-45.162	167.314	94.971	3.9
2726331	2007	107	-44.817	167.326	21.166	4.4
2726729	2007	108	-41.232	172.521	231.010	3.6
2727588	2007	110	-41.246	172.550	212.435	3.6
2727967	2007	111	-45.135	167.237	52.040	4.3
2732430	2007	121	-40.695	173.213	176.194	4.0
2733458	2007	123	-45.370	167.130	83.072	3.7
2734475	2007	125	-41.381	172.930	116.220	3.6
2734605	2007	126	-40.565	173.267	193.383	3.5
2735669	2007	128	-44.588	168.171	86.900	3.5
2737094	2007	131	-42.350	173.641	28.166	4.4
2737975	2007	133	-45.371	167.150	100.502	3.5
2737984	2007	133	-41.217	173.549	89.750	5.4
2741903	2007	142	-41.017	172.877	206.570	3.5
2741903	2007	142	-41.017	172.877	206.820	3.5
2743833	2007	146	-45.021	167.565	88.321	3.8
2743899	2007	146	-41.242	172.610	197.143	3.6
2744487	2007	147	-41.034	172.868	202.298	3.5
2747840	2007	155	-45.000	167.513	59.995	3.5
2748555	2007	156	-44.626	168.200	81.052	4.2
2748555	2007	156	-44.626	168.200	88.683	4.2
2751226	2007	162	-41.417	174.690	56.064	3.8
2754371	2007	169	-45.076	167.449	114.475	3.6
2754669	2007	170	-43.186	172.267	28.144	3.3
2757147	2007	175	-41.598	174.035	38.907	4.0
2758600	2007	178	-41.999	174.376	31.184	3.5
2760737	2007	183	-45.053	167.393	49.226	3.7
2762590	2007	187	-46.029	167.118	21.397	3.8
2763016	2007	188	-45.214	167.321	82.588	4.5
2768313	2007	200	-45.181	167.337	73.925	3.9
2769800	2007	203	-41.219	172.586	217.863	3.7
2770871	2007	206	-41.298	172.569	188.682	3.5
2775327	2007	216	-44.458	168.361	5.000	3.7
2778812	2007	223	-41.319	173.652	72.117	3.9
2780655	2007	227	-42.430	173.649	29.924	3.6
2780734	2007	227	-40.937	173.357	130.358	5.1
2780735	2007	227	-40.943	173.442	136.420	4.8
2780892	2007	228	-41.052	174.194	57.323	4.2
2782916	2007	232	-44.402	167.755	20.508	3.4

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
2783015	2007	232	-42.688	172.626	42.533	3.1
2783557	2007	233	-41.174	173.073	152.386	3.7
2784297	2007	235	-45.516	166.986	110.364	3.8
2784513	2007	236	-42.582	172.847	5.745	4.0
2786597	2007	240	-41.639	172.793	94.768	4.1
2786597	2007	240	-41.651	172.814	85.143	4.1
2788623	2007	245	-44.663	167.925	94.201	3.5
2788623	2007	245	-44.731	167.930	93.617	3.6
2790417	2007	249	-45.322	166.609	20.453	4.6
2791597	2007	251	-40.646	173.016	219.333	3.5
2791978	2007	252	-41.404	173.193	85.918	3.7
2791978	2007	252	-41.405	173.205	87.662	3.6
2794086	2007	257	-45.177	167.293	134.334	3.6
2795163	2007	259	-40.951	172.797	225.224	3.7
2795420	2007	260	-45.067	167.450	85.148	3.5
2796284	2007	262	-44.592	168.109	82.181	4.5
2797472	2007	265	-43.420	172.942	32.885	3.1
2801221	2007	273	-45.032	166.837	19.037	3.9
2802758	2007	276	-42.258	173.066	92.390	4.0
2802763	2007	276	-42.143	172.825	72.516	3.6
2802792	2007	276	-42.151	172.870	61.711	3.7
2802991	2007	277	-45.176	167.378	120.162	3.7
2803171	2007	277	-42.100	172.850	58.189	4.0
2803171	2007	277	-42.154	172.842	59.814	4.0
2804091	2007	279	-42.201	172.933	70.369	4.8
2804091	2007	279	-42.202	172.904	67.840	4.7
2804454	2007	280	-44.882	167.675	78.063	3.6
2805690	2007	282	-41.258	172.605	195.854	3.7
2806153	2007	283	-42.151	172.838	58.841	3.7
2806153	2007	283	-42.160	172.818	51.075	3.7
2809105	2007	290	-44.996	167.557	86.156	3.5
2809382	2007	290	-41.404	172.848	129.651	3.7
2810009	2007	292	-44.727	167.445	19.387	3.2
2811409	2007	295	-40.694	173.804	107.215	4.8
2812268	2007	297	-44.880	167.285	17.337	4.5
2814738	2007	303	-44.750	167.454	19.281	3.2
2815104	2007	303	-44.797	167.382	17.578	3.7
2815873	2007	305	-44.778	167.416	20.827	4.4
2815887	2007	305	-45.280	167.253	80.174	3.7
2816258	2007	306	-45.115	167.369	118.740	3.5
2818146	2007	310	-42.100	172.893	57.038	5.4
2818174	2007	310	-42.123	172.875	58.925	4.5
2819042	2007	313	-45.549	166.965	70.506	3.6
2820110	2007	315	-44.498	167.854	15.364	3.5
2821549	2007	318	-44.486	166.930	20.833	3.3
2823057	2007	320	-44.406	168.380	23.384	3.1
2826859	2007	328	-41.880	172.844	93.681	3.7
2827926	2007	330	-45.041	167.511	83.621	4.1
2829153	2007	333	-45.064	167.257	117.077	5.9
2829912	2007	335	-45.333	167.261	58.636	3.7
2841741	2007	359	-40.969	173.188	155.055	3.5
2842767	2007	361	-40.926	172.878	220.807	4.3
2842976	2007	362	-45.149	167.023	28.565	4.2
2847841	2008	008	-44.740	167.466	52.052	3.4
2848597	2008	010	-42.219	172.807	65.233	4.3
2848762	2008	010	-42.281	172.998	65.980	3.7
2849290	2008	012	-40.317	173.335	176.305	4.3
2849572	2008	012	-45.450	167.076	104.781	3.5
2850317	2008	013	-41.229	172.542	217.440	3.6
2855227	2008	024	-45.075	167.392	119.892	3.6
2855529	2008	024	-44.961	167.463	101.868	3.9
2856129	2008	026	-41.086	174.248	54.669	3.7
2859104	2008	032	-45.063	167.392	107.260	3.7
2859712	2008	033	-44.915	167.393	118.057	4.5
2859889	2008	034	-45.220	167.337	128.290	3.7
2859915	2008	034	-41.271	172.625	213.545	3.7
2860839	2008	036	-41.569	173.528	65.788	4.3
2862831	2008	040	-40.511	173.183	239.539	3.8
2863084	2008	040	-45.205	167.345	118.213	3.5
2864471	2008	043	-40.676	174.560	59.173	4.0

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
2865339	2008	045	-45.075	166.463	21.367	3.7
2866423	2008	047	-41.250	172.945	194.620	3.7
2867818	2008	050	-45.123	167.696	122.296	3.9
2867952	2008	050	-44.978	167.423	120.998	3.7
2869115	2008	053	-40.948	172.880	220.531	3.7
2869234	2008	053	-45.098	167.450	58.094	3.7
2871428	2008	057	-45.523	167.067	66.863	4.0
2872889	2008	060	-40.673	174.052	92.549	3.5
2876720	2008	068	-45.191	167.349	117.612	3.6
2877311	2008	069	-40.895	172.852	216.311	4.2
2880560	2008	076	-40.699	173.786	97.901	4.6
2880972	2008	077	-45.141	166.752	22.439	3.5
2884471	2008	084	-41.187	172.735	174.637	3.5
2885017	2008	084	-43.050	172.629	19.317	3.3
2887310	2008	090	-44.674	167.921	99.741	3.3
2888755	2008	093	-45.012	167.560	89.280	3.9
2889129	2008	094	-45.011	167.462	94.652	4.4
2889513	2008	095	-41.266	173.815	64.932	3.5
2889830	2008	096	-40.822	174.895	38.275	3.5
2890301	2008	097	-43.844	173.328	25.204	3.2
2891738	2008	099	-45.063	167.477	92.569	3.6
2892435	2008	101	-44.597	168.126	91.573	3.0
2892935	2008	102	-44.905	167.603	72.297	4.0
2893405	2008	103	-45.154	167.360	145.167	3.3
2894271	2008	105	-40.550	174.394	59.177	3.9
2894504	2008	105	-45.486	167.129	108.723	3.9
2897192	2008	111	-41.177	173.596	93.843	3.9
2897985	2008	113	-41.233	172.564	213.022	4.0
2897985	2008	113	-41.233	172.564	215.676	4.0
2901496	2008	120	-44.860	167.512	26.972	2.1
2902528	2008	122	-43.225	171.019	5.000	4.6
2904522	2006	324	-40.970	172.850	192.007	3.5
2904865	2008	127	-44.530	168.053	81.539	3.0
2907267	2008	132	-45.256	167.229	56.734	4.5
2910138	2008	139	-44.614	168.032	73.617	3.2
2911911	2008	143	-44.627	168.032	92.807	3.1
2914692	2006	349	-41.409	172.952	115.340	3.6
2918151	2008	156	-40.576	174.583	91.385	3.6
2920375	2008	160	-41.246	172.610	214.473	3.6
2920666	2008	161	-45.472	167.049	64.402	3.5
2923267	2008	166	-41.175	174.417	39.559	3.7
2925442	2008	170	-42.605	172.713	54.512	3.5
2926002	2008	172	-41.281	172.570	227.521	3.6
2929377	2008	178	-45.461	166.907	51.086	3.6
2929831	2008	179	-40.553	174.292	64.570	3.9
2931647	2008	183	-40.646	173.290	162.902	4.6
2936807	2008	193	-42.717	172.889	33.000	3.0
2936916	2008	193	-45.382	167.066	67.627	3.7
2936936	2008	194	-45.791	166.804	102.662	6.0
2938906	2008	198	-40.938	174.653	68.084	4.2
2944056	2008	208	-41.750	174.333	55.651	3.5
2944471	2008	209	-43.506	169.944	10.561	3.2
2946142	2008	212	-41.222	173.857	63.434	3.7
2950527	2008	221	-45.002	167.516	80.035	4.4
2950769	2008	222	-40.533	173.176	240.017	3.5
2951523	2008	223	-42.891	172.881	28.597	3.1
2952161	2008	225	-45.066	167.469	81.384	3.8
2953023	2008	226	-44.616	168.154	85.148	3.3
2954005	2008	228	-45.081	167.346	109.954	4.1
2956873	2008	234	-41.204	172.518	213.742	4.4
2961674	2008	244	-41.219	172.757	147.890	3.8
2961889	2008	237	-41.202	172.904	156.906	3.7
2962981	2008	246	-41.398	172.841	109.057	3.7
2963425	2008	247	-41.479	173.517	62.070	4.5
2964617	2008	250	-40.585	173.284	164.971	4.5
2966534	2008	253	-42.090	173.003	81.566	3.9
2967823	2008	256	-44.989	167.472	54.182	3.5
2968431	2008	257	-45.007	167.560	70.056	4.5
2969446	2008	259	-41.212	172.605	206.442	3.6
2970989	2008	262	-41.832	172.632	83.879	3.6

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
2972380	2008	265	-42.957	172.673	33.000	3.2
2973670	2008	268	-42.025	173.922	21.118	3.6
2974157	2008	269	-42.284	173.887	20.249	4.6
2974665	2008	270	-42.309	173.923	23.894	5.2
2974715	2008	270	-42.286	173.889	21.674	3.8
2976419	2008	273	-44.968	167.408	81.125	4.1
2977473	2008	275	-41.289	172.656	179.403	3.9
2978678	2008	278	-40.236	173.441	164.403	3.5
2979524	2008	270	-42.279	173.887	20.545	3.7
2979619	2008	280	-41.737	173.143	69.120	3.7
2980909	2008	282	-42.803	172.719	30.422	3.1
2985839	2008	292	-41.251	172.810	135.250	3.7
2985966	2008	292	-42.035	172.620	76.907	4.0
2987302	2008	295	-40.653	174.201	67.895	3.5
2987304	2008	295	-45.123	166.864	26.745	4.1
2993491	2008	307	-41.024	174.571	39.025	3.7
2993608	2008	307	-45.029	167.455	73.528	3.6
2993623	2008	307	-45.491	167.010	69.198	4.0
2993712	2008	308	-45.109	167.404	92.821	3.9
2995312	2008	310	-41.093	174.581	60.404	3.6
2996979	2008	314	-45.161	167.306	122.081	5.0
2997750	2008	315	-41.741	174.391	55.840	4.0
2998010	2008	316	-44.963	167.499	133.586	3.7
2999479	2008	319	-41.815	172.701	87.947	4.3
2999492	2008	319	-45.083	167.424	91.125	3.8
2999830	2008	320	-41.953	174.343	24.764	3.5
3000748	2008	321	-40.838	172.832	210.909	4.8
3001721	2008	323	-44.371	169.835	5.000	3.9
3002448	2008	324	-45.221	167.123	28.573	4.0
3003193	2008	326	-43.575	171.461	20.444	3.1
3003350	2008	326	-41.146	172.722	200.846	3.7
3004305	2008	328	-40.804	174.473	52.653	3.8
3005451	2008	330	-45.322	167.064	30.274	4.2
3008283	2008	336	-45.013	167.492	77.751	3.9
3008539	2008	336	-44.349	170.116	13.324	4.1
3010647	2008	340	-42.695	171.278	5.000	3.6
3011535	2008	342	-44.955	167.429	70.787	3.7
3011793	2008	343	-44.663	168.120	85.215	3.6
3013837	2008	347	-44.099	169.282	33.000	3.1
3019466	2008	358	-45.365	167.250	123.754	4.2
3019527	2008	358	-45.344	167.185	103.036	4.1
3021098	2008	361	-40.554	174.867	52.144	5.1
3021243	2008	362	-41.933	174.408	33.000	3.6
3024412	2009	002	-45.409	167.026	55.084	4.0
3025017	2009	003	-40.561	174.858	49.503	3.8
3025488	2009	004	-44.737	167.803	62.004	3.7
3028730	2009	011	-41.582	173.238	99.906	4.6
3028783	2009	010	-41.302	172.594	208.350	4.1
3030153	2009	014	-44.870	167.555	116.042	3.7
3035179	2009	024	-45.081	167.429	90.692	4.2
3035714	2009	025	-45.146	167.398	109.224	3.6
3041931	2009	038	-41.233	172.566	213.946	3.6
3042565	2009	039	-45.537	167.071	76.165	3.9
3042860	2009	040	-41.668	173.420	54.237	3.6
3046321	2009	047	-42.148	172.848	54.321	2.2
3047203	2009	049	-42.838	172.948	30.006	2.6
3047204	2009	049	-41.708	173.094	90.129	2.5
3047503	2009	049	-45.096	167.389	117.456	3.9
3049306	2009	053	-42.736	172.847	5.000	2.6
3050923	2009	056	-43.504	171.544	12.000	3.5
3051399	2009	057	-41.221	172.725	201.834	3.4
3051685	2009	058	-43.063	173.362	26.056	3.8
3051705	2009	058	-43.084	173.428	33.000	2.7
3052037	2009	059	-42.086	172.753	58.838	2.5
3052463	2009	059	-43.208	171.474	5.000	2.8
3053423	2009	061	-41.742	173.968	31.649	2.3
3053698	2009	062	-42.735	173.341	33.000	2.6
3053791	2009	062	-41.594	172.722	100.908	2.5
3054051	2009	063	-41.856	173.107	55.935	2.5
3054325	2009	063	-43.214	172.156	5.000	4.1

Continued on next page. . .

Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3054326	2009	063	-43.211	172.167	5.000	3.3
3055648	2009	066	-43.609	171.143	5.000	2.4
3055781	2009	066	-41.400	172.991	104.129	2.8
3056338	2009	067	-41.169	172.869	141.322	3.1
3057524	2009	069	-42.968	171.446	5.000	3.5
3057877	2009	070	-41.688	174.477	54.052	4.3
3058398	2009	071	-42.147	172.811	49.076	2.7
3058975	2009	072	-45.095	167.398	117.789	3.8
3059515	2009	073	-40.542	173.522	158.489	4.1
3060955	2009	076	-42.632	172.026	21.210	2.1
3061005	2009	076	-40.434	173.543	158.239	3.6
3061192	2009	076	-41.232	172.593	215.319	3.5
3061664	2009	077	-41.405	173.401	64.561	2.6
3062588	2009	079	-41.830	173.260	50.894	2.6
3062982	2009	080	-40.968	172.762	186.497	3.3
3064185	2009	082	-41.302	172.804	131.611	2.4
3065595	2009	085	-45.061	167.422	97.395	4.7
3066315	2009	086	-44.256	169.993	14.145	4.8
3066448	2009	086	-42.705	172.346	11.069	2.5
3066652	2009	087	-44.264	170.012	14.891	3.2
3066767	2009	087	-42.893	173.163	33.000	2.1
3067191	2009	088	-42.426	172.797	52.821	2.5
3067364	2009	088	-41.739	172.521	86.280	2.6
3068099	2009	090	-41.499	172.888	123.965	4.6
3068852	2009	091	-45.025	167.393	125.892	3.6
3069262	2009	092	-43.132	171.461	5.000	2.3
3069643	2009	093	-43.007	172.992	33.000	2.9
3071401	2009	096	-41.660	172.901	91.317	4.5
3073195	2009	100	-40.979	172.875	200.923	3.1
3074691	2009	103	-41.469	172.876	121.837	3.9
3074989	2009	103	-41.072	172.880	181.175	2.7
3075089	2009	103	-43.519	171.640	5.000	2.5
3075740	2009	105	-43.486	171.607	5.000	2.3
3076083	2009	104	-41.959	172.196	84.238	2.5
3076229	2009	106	-42.928	173.063	33.000	3.3
3077046	2009	107	-42.540	171.933	5.000	2.1
3077049	2009	107	-41.171	173.220	109.017	3.1
3077937	2009	109	-40.402	173.590	154.805	3.7
3079510	2009	112	-40.741	174.077	96.377	3.0
3079670	2009	112	-41.231	172.704	146.663	2.6
3080021	2009	113	-41.377	172.966	116.455	3.7
3080313	2009	113	-41.314	172.707	216.569	2.3
3080807	2009	114	-41.284	172.678	208.877	3.8
3084351	2009	121	-41.187	174.805	30.875	4.1
3084790	2009	122	-44.259	170.003	14.189	3.6
3086566	2009	125	-41.357	173.115	92.435	2.9
3086578	2009	125	-42.503	172.979	12.000	3.2
3087732	2009	128	-43.152	170.960	5.000	4.0
3088222	2009	129	-43.233	171.603	5.000	2.2
3088333	2009	129	-41.245	173.869	55.237	3.0
3090764	2009	133	-42.643	171.961	17.417	2.7
3090908	2009	134	-42.153	173.842	33.000	2.5
3091002	2009	134	-43.113	171.023	5.000	3.0
3091380	2009	134	-41.374	174.118	39.063	4.5
3091508	2009	135	-43.395	171.445	5.000	2.5
3091656	2009	135	-41.134	173.919	67.400	3.7
3091866	2009	135	-41.815	172.450	84.029	2.6
3093057	2009	137	-41.283	172.799	129.064	3.0
3093437	2009	138	-43.179	171.559	5.000	2.0
3093885	2009	139	-41.795	173.216	87.277	2.7
3094669	2009	140	-41.974	173.359	30.572	2.5
3096079	2009	143	-45.410	167.196	136.906	3.6
3096312	2009	143	-44.179	171.186	12.000	2.3
3096871	2009	144	-45.158	167.356	120.805	3.9
3097684	2009	146	-41.534	173.302	86.897	2.6
3097924	2009	146	-41.077	174.582	38.910	3.5
3099415	2009	149	-42.956	172.413	12.000	2.4
3101119	2009	152	-40.324	173.314	172.081	4.2
3101277	2009	152	-41.335	172.735	103.093	2.3
3101867	2009	153	-41.252	172.589	195.736	3.4

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3101969	2009	153	-42.951	172.422	17.954	2.6
3102158	2009	154	-43.238	172.012	10.210	2.5
3102845	2009	155	-41.060	174.540	44.080	4.0
3103763	2009	157	-42.133	172.787	57.840	2.8
3104152	2009	158	-41.474	173.899	42.670	2.8
3104352	2009	158	-41.071	174.557	43.478	3.9
3105343	2009	160	-42.592	172.863	5.000	2.5
3106393	2009	162	-41.708	172.189	102.730	2.6
3106824	2009	163	-42.045	172.356	76.179	3.3
3107010	2009	163	-43.672	171.604	5.000	3.3
3107189	2009	163	-41.414	172.504	118.326	3.0
3107602	2009	164	-45.214	167.304	105.020	3.8
3108186	2009	165	-44.827	167.250	33.000	3.7
3108837	2009	166	-40.555	173.419	165.474	4.1
3109454	2009	167	-42.038	172.925	75.936	3.3
3110536	2009	169	-41.113	172.970	160.957	3.1
3111177	2009	171	-42.115	173.337	75.988	3.0
3112187	2009	172	-42.042	172.837	62.720	2.3
3112232	2009	155	-43.608	171.514	5.000	2.2
3112571	2009	173	-43.130	172.329	27.711	2.1
3112716	2009	173	-40.943	172.834	193.891	3.9
3113703	2009	175	-41.181	173.016	128.605	2.3
3114247	2009	176	-43.218	172.297	17.991	2.0
3114306	2009	176	-41.241	172.603	198.377	3.1
3114457	2009	176	-42.429	173.047	20.000	2.3
3114581	2009	177	-43.477	171.467	5.000	2.3
3115714	2009	178	-42.502	169.801	33.000	4.8
3115973	2009	179	-42.693	173.438	33.000	3.6
3118326	2009	183	-41.550	172.995	91.464	4.2
3118345	2009	183	-41.578	172.993	91.361	4.6
3119086	2009	185	-43.394	172.451	12.000	2.9
3119449	2009	185	-40.854	172.931	210.082	3.4
3119699	2009	186	-42.069	172.923	80.538	2.9
3120449	2009	187	-41.264	172.776	144.004	2.5
3120467	2009	188	-43.022	172.722	25.211	3.0
3120521	2009	188	-41.431	172.987	92.780	2.7
3120655	2009	188	-43.416	172.538	5.000	2.9
3120890	2009	188	-44.728	167.851	96.719	3.0
3121142	2009	189	-41.289	172.777	146.310	2.8
3121525	2009	180	-44.284	171.890	33.000	2.3
3121535	2009	190	-41.234	172.641	195.414	2.7
3122125	2009	191	-41.489	172.996	78.819	2.2
3122292	2009	191	-40.985	172.916	174.270	4.2
3123338	2009	193	-40.917	173.065	172.778	3.5
3123875	2009	194	-41.262	172.635	172.555	2.6
3123888	2009	194	-41.327	173.373	70.068	2.3
3124008	2009	194	-42.240	173.892	33.000	3.7
3124085	2009	194	-41.295	172.602	190.798	3.4
3124516	2009	195	-40.893	172.790	243.850	3.4
3124878	2009	196	-46.278	166.267	177.420	5.2
3125061	2009	196	-45.521	167.033	65.012	5.1
3125201	2009	197	-45.897	166.211	33.000	5.4
3125559	2009	198	-45.702	166.529	19.531	4.6
3125643	2009	198	-42.763	171.934	5.000	3.2
3125686	2009	198	-45.727	166.531	22.396	3.9
3125689	2009	198	-45.395	166.706	21.864	4.9
3125707	2009	198	-45.384	166.703	21.886	3.5
3125797	2009	198	-45.557	166.564	20.887	4.3
3125822	2009	198	-45.494	166.621	21.520	3.9
3125829	2009	198	-46.019	166.482	32.839	3.8
3125831	2009	198	-45.428	166.625	21.897	4.3
3125939	2009	198	-45.420	166.701	21.574	4.5
3125991	2009	198	-45.529	166.558	21.431	4.4
3126025	2009	199	-45.363	166.658	21.554	4.8
3126043	2009	199	-45.459	166.600	22.887	3.7
3126108	2009	199	-45.523	166.679	21.855	3.5
3126142	2009	199	-45.398	166.675	23.063	3.6
3126218	2009	199	-45.863	166.345	21.578	4.1
3126262	2009	199	-45.490	166.553	20.181	3.6
3126393	2009	199	-45.501	166.542	21.470	4.3

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3126549	2009	200	-43.101	171.003	5.000	2.4
3126568	2009	200	-45.682	166.528	21.560	4.1
3126670	2009	200	-45.714	166.493	20.943	3.5
3126729	2009	200	-45.545	166.654	22.410	4.4
3126734	2009	200	-43.056	171.245	5.000	2.9
3126801	2009	200	-45.673	166.529	18.602	3.7
3126843	2009	200	-45.437	166.663	21.870	3.5
3127061	2009	201	-45.464	166.746	21.983	3.6
3127203	2009	201	-45.470	166.631	21.900	3.8
3127714	2009	202	-45.881	166.438	26.520	4.0
3127867	2009	203	-43.039	173.246	20.779	3.1
3128341	2009	204	-42.668	172.595	33.000	3.3
3128712	2009	204	-45.409	166.730	22.087	3.8
3128785	2009	204	-45.376	166.704	24.653	3.7
3128864	2009	204	-45.460	166.673	23.413	4.3
3129136	2009	205	-45.393	166.722	23.897	4.6
3129198	2009	205	-44.615	168.159	80.070	3.2
3129333	2009	205	-45.390	166.591	19.728	3.6
3129804	2009	207	-41.290	172.857	144.939	3.2
3129834	2009	207	-41.420	172.992	79.919	2.3
3129894	2009	207	-41.292	172.704	158.588	3.6
3130125	2009	207	-45.889	166.258	26.656	4.8
3130282	2009	208	-45.056	166.940	33.000	3.7
3130541	2009	208	-45.499	166.588	22.036	4.6
3130656	2009	208	-45.454	166.648	23.256	3.8
3130676	2009	209	-45.471	166.657	21.433	3.6
3131488	2009	210	-45.548	166.632	20.584	5.1
3131501	2009	211	-45.539	166.646	20.917	4.0
3131736	2009	211	-45.618	166.469	19.518	3.7
3131896	2009	211	-45.646	166.448	21.977	4.7
3132013	2009	212	-45.535	166.611	20.605	3.8
3132372	2009	212	-45.446	166.710	24.893	4.3
3132601	2009	213	-44.971	167.535	127.785	3.6
3132839	2009	213	-41.766	173.048	90.793	2.5
3133043	2009	214	-45.464	166.759	22.554	4.6
3133200	2009	214	-46.105	166.305	28.945	3.9
3133354	2009	214	-45.681	166.642	22.519	3.8
3133815	2009	215	-46.161	166.277	33.000	4.2
3133914	2009	215	-46.248	166.279	33.000	3.8
3134050	2009	215	-41.460	172.926	111.391	3.2
3134115	2009	216	-41.811	174.345	29.308	4.2
3134378	2009	216	-45.498	166.660	22.454	3.8
3134387	2009	216	-44.593	168.208	81.415	3.0
3134616	2009	200	-45.413	166.714	20.706	4.0
3135253	2009	218	-45.442	166.676	21.755	3.5
3135313	2009	218	-45.476	166.746	21.993	3.8
3135487	2009	200	-46.106	166.206	32.296	3.6
3135552	2009	219	-45.858	166.425	39.614	3.7
3135754	2009	219	-45.551	166.658	20.658	4.5
3136366	2009	220	-41.314	173.396	106.713	3.0
3136653	2009	221	-41.067	174.027	60.141	3.5
3136713	2009	221	-45.542	166.627	23.347	4.8
3137032	2009	198	-45.491	166.569	21.123	3.6
3137460	2009	223	-45.963	166.434	30.215	3.7
3137567	2009	223	-41.497	172.590	105.322	2.3
3137861	2009	198	-45.380	166.703	21.864	3.7
3138040	2009	224	-42.249	173.018	63.885	2.3
3138587	2009	225	-45.339	166.767	26.962	3.7
3138705	2009	225	-41.370	172.838	122.640	3.0
3139106	2009	226	-43.256	170.790	3.000	4.1
3139431	2009	227	-41.293	172.778	139.405	2.6
3139540	2009	227	-41.497	174.177	61.478	3.1
3139605	2009	227	-43.207	170.921	5.000	2.6
3139869	2009	228	-41.362	172.767	136.649	3.1
3139883	2009	228	-45.668	166.651	21.164	3.5
3140195	2009	229	-45.552	166.656	22.020	4.1
3140201	2009	229	-42.800	172.885	33.000	2.9
3140802	2009	230	-43.260	170.919	12.000	2.9
3141083	2009	231	-44.642	168.156	81.365	3.5
3141134	2009	231	-41.739	172.297	96.294	2.6

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3141338	2009	231	-40.530	173.610	118.346	3.8
3141375	2009	231	-41.726	172.188	100.391	3.3
3141471	2009	231	-42.992	172.660	12.000	2.4
3141478	2009	231	-46.032	166.272	33.000	3.6
3141703	2009	232	-44.594	168.173	80.351	3.3
3141850	2009	232	-41.714	172.304	91.897	2.4
3142187	2009	233	-42.078	172.786	56.183	2.2
3142436	2009	234	-42.713	171.693	5.000	2.7
3142644	2009	234	-41.161	172.859	145.078	3.5
3143358	2009	199	-45.475	166.638	22.118	4.1
3143553	2009	236	-41.621	172.818	86.083	2.7
3143670	2009	199	-45.747	166.529	22.234	3.5
3143705	2009	237	-45.150	167.280	121.351	3.8
3144863	2009	199	-45.496	166.634	19.625	3.7
3145196	2009	239	-41.434	174.702	33.979	4.3
3145383	2009	240	-45.499	166.789	17.847	4.0
3145405	2009	240	-43.399	170.791	5.000	2.8
3145684	2009	240	-45.510	166.481	19.859	4.0
3146145	2009	241	-40.910	172.832	214.090	3.5
3146444	2009	242	-45.539	166.552	20.120	4.9
3146610	2009	242	-45.429	166.770	23.174	3.9
3147147	2009	243	-41.603	173.409	76.593	2.9
3147878	2009	244	-41.844	172.722	94.546	2.7
3148257	2009	228	-41.314	172.584	183.094	3.4
3148278	2009	245	-45.438	166.582	22.305	4.0
3148588	2009	245	-40.926	172.922	187.429	4.0
3148730	2009	228	-41.431	173.680	50.641	2.5
3148981	2009	246	-41.681	172.714	77.311	2.0
3149453	2009	247	-46.261	166.292	33.000	4.9
3149488	2009	247	-42.136	172.758	66.106	2.6
3150135	2009	249	-40.915	172.912	220.487	3.4
3150506	2009	249	-41.255	172.764	111.313	2.1
3150827	2009	250	-41.309	172.827	139.828	3.8
3150963	2009	250	-42.060	172.849	74.711	3.0
3151539	2009	252	-45.535	166.695	23.847	4.0
3151678	2009	252	-41.292	172.719	156.583	3.3
3151852	2009	252	-45.546	166.690	20.138	4.9
3152083	2009	253	-40.725	174.347	72.184	4.1
3152416	2009	254	-41.781	172.626	80.212	2.7
3152666	2009	254	-43.066	173.530	12.000	2.5
3152685	2009	254	-40.813	174.620	44.762	3.7
3153108	2009	255	-41.311	174.546	41.211	3.6
3153511	2009	256	-43.053	172.816	29.129	3.4
3153542	2009	256	-41.280	172.554	148.831	2.7
3154121	2009	247	-42.717	171.684	5.000	2.6
3155299	2009	260	-45.094	167.368	123.759	4.2
3155307	2009	260	-41.679	172.516	84.173	2.1
3156245	2009	262	-41.262	172.576	193.485	3.3
3156489	2009	262	-42.761	171.933	5.000	3.1
3157895	2009	265	-41.724	173.195	73.163	3.0
3158405	2009	266	-41.014	173.225	129.210	3.3
3158502	2009	267	-41.748	172.473	80.029	3.2
3158781	2009	267	-42.740	171.980	12.000	3.2
3159511	2009	269	-41.727	172.215	100.475	3.1
3159733	2009	270	-45.709	166.541	16.656	3.8
3159932	2009	270	-41.424	174.243	63.695	3.3
3160158	2009	270	-41.331	173.291	78.377	3.5
3160218	2009	271	-43.024	172.569	26.633	2.2
3160832	2009	272	-41.447	173.271	82.509	2.6
3160872	2009	272	-43.245	171.020	5.000	2.8
3161447	2009	273	-45.305	166.844	27.355	3.7
3161988	2009	274	-45.392	166.700	21.507	3.6
3162137	2009	275	-44.592	168.173	81.789	3.1
3162238	2009	275	-42.984	171.469	5.000	3.5
3162524	2009	275	-41.448	172.821	110.288	2.4
3163469	2009	278	-40.966	173.431	89.585	3.1
3163716	2009	278	-41.718	172.277	98.095	2.6
3163741	2009	278	-45.440	166.627	21.401	3.5
3164576	2009	280	-42.114	172.815	55.253	2.9
3164688	2009	280	-41.218	172.628	212.621	3.1

Continued on next page. . .

Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3165024	2009	281	-45.561	166.723	21.242	4.8
3165121	2009	281	-41.458	172.827	89.094	2.4
3165297	2009	281	-41.525	172.760	97.597	2.8
3166123	2009	283	-43.191	170.807	5.000	2.6
3166147	2009	283	-41.274	172.647	177.074	3.0
3166368	2009	262	-43.040	172.670	5.000	2.1
3166997	2009	285	-45.062	167.386	92.545	3.6
3167025	2009	285	-41.714	172.265	100.618	2.4
3167546	2009	286	-41.269	173.251	111.029	2.9
3167757	2009	287	-41.256	172.772	137.411	2.9
3168364	2009	288	-43.174	171.586	5.000	2.3
3168423	2009	288	-45.604	166.766	18.208	3.5
3168594	2009	288	-41.258	172.742	154.633	2.8
3168595	2009	280	-41.247	172.618	154.227	2.4
3170003	2009	290	-42.735	172.420	5.000	2.6
3170073	2009	290	-45.083	167.454	94.385	4.1
3170768	2009	292	-43.311	171.030	5.000	2.6
3170794	2009	292	-41.404	174.285	61.820	3.0
3171001	2009	292	-45.148	167.089	33.000	4.3
3171353	2009	293	-41.460	173.537	70.743	2.7
3171506	2009	293	-41.706	172.205	102.110	2.5
3171550	2009	293	-45.665	166.876	62.468	3.6
3171667	2009	294	-40.715	174.451	67.502	3.1
3171768	2009	294	-42.879	171.606	5.000	2.6
3171833	2009	294	-41.064	172.958	155.883	3.6
3172135	2009	274	-41.201	172.595	211.171	3.0
3172220	2009	295	-41.285	172.933	117.325	3.7
3172297	2009	295	-42.414	171.514	12.000	2.5
3172529	2009	295	-41.096	173.597	92.472	3.5
3172864	2009	296	-41.865	172.449	79.415	2.6
3173133	2009	297	-41.283	172.720	147.248	3.0
3173202	2009	297	-41.749	173.399	44.254	2.1
3173371	2009	297	-41.133	173.188	113.744	3.0
3174442	2009	299	-45.134	167.331	114.108	3.7
3176325	2009	303	-40.537	173.355	179.809	3.4
3176331	2009	303	-43.103	172.098	5.000	2.1
3177164	2009	305	-43.198	172.225	29.387	2.1
3177784	2009	306	-41.141	173.411	95.189	2.7
3178098	2009	307	-40.969	173.820	79.762	3.1
3178429	2009	308	-42.048	172.761	89.719	4.5
3178945	2009	309	-41.694	172.060	23.009	2.5
3181530	2009	314	-42.476	172.891	42.801	3.5
3182026	2009	315	-40.537	173.544	153.829	3.4
3182458	2009	316	-45.463	166.746	21.672	4.0
3183648	2009	318	-41.814	174.435	20.547	3.7
3185157	2009	321	-41.215	172.635	129.567	2.0
3185640	2009	322	-41.389	172.348	159.390	2.7
3186228	2009	323	-43.218	172.871	33.000	4.2
3187010	2009	325	-45.021	167.431	89.211	3.5
3187561	2009	326	-42.530	171.929	5.000	2.6
3188308	2009	328	-44.603	168.197	88.074	3.0
3188343	2009	328	-43.531	173.028	32.755	2.9
3188409	2009	197	-45.510	166.549	16.951	4.2
3190570	2009	332	-43.339	171.961	8.759	3.5
3190843	2009	333	-45.258	166.970	33.000	3.8
3202873	2009	336	-44.941	167.103	20.512	3.7
3202903	2009	336	-42.964	172.437	5.000	3.2
3204025	2009	339	-40.805	172.910	209.131	4.0
3204627	2009	340	-41.359	173.801	53.861	3.3
3205351	2009	342	-46.011	166.840	81.918	4.1
3205715	2009	342	-40.562	174.853	48.911	5.1
3206052	2009	343	-45.117	167.373	95.279	3.8
3206940	2009	345	-41.083	174.318	67.175	3.7
3207677	2009	346	-43.139	172.290	30.905	2.3
3211310	2009	353	-45.000	167.494	97.385	3.6
3213039	2009	357	-41.838	172.885	85.668	2.3
3213667	2009	359	-44.533	167.708	8.983	3.0
3213731	2009	359	-41.760	172.748	87.916	2.5
3215575	2009	363	-43.136	170.989	5.000	2.8
3219434	2009	361	-41.950	173.029	48.971	2.2

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3219727	2010	006	-45.066	167.404	88.520	3.8
3228164	2010	008	-45.133	167.244	84.824	3.8
3229793	2010	011	-45.101	167.342	116.846	3.7
3229901	2010	012	-44.167	168.445	5.000	3.7
3239754	2010	013	-43.084	171.430	5.000	3.4
3241870	2010	014	-44.133	168.387	5.000	4.6
3241943	2010	014	-42.998	171.442	9.424	2.9
3242948	2010	016	-44.306	168.080	5.000	3.4
3243091	2010	016	-43.200	173.495	33.000	3.8
3244016	2010	019	-45.251	167.310	155.558	3.7
3245798	2010	022	-44.186	168.443	2.000	3.1
3245861	2010	022	-45.479	166.694	12.000	4.6
3245867	2010	022	-45.465	166.761	12.000	4.1
3245913	2010	022	-44.164	168.423	5.000	4.1
3246039	2010	023	-41.203	172.747	169.818	3.5
3246628	2010	024	-46.067	166.815	81.818	3.8
3253233	2010	037	-44.570	168.146	87.847	3.6
3253712	2010	038	-41.258	172.602	211.170	3.6
3254077	2010	039	-45.364	166.700	12.000	3.7
3254472	2010	040	-41.952	172.605	78.788	4.4
3255116	2010	041	-42.525	172.660	5.000	3.2
3259829	2010	042	-45.149	167.237	77.023	4.0
3260765	2010	044	-45.368	167.022	63.231	3.6
3260895	2010	044	-46.092	166.213	12.000	4.7
3261607	2010	046	-45.430	166.678	12.000	3.7
3261808	2010	046	-40.915	172.911	220.784	3.6
3267369	2010	057	-44.843	167.314	5.000	3.6
3267529	2010	058	-44.839	167.320	5.000	3.8
3267555	2010	058	-44.838	167.298	5.000	3.8
3267698	2010	058	-46.020	166.884	5.000	4.4
3268042	2010	059	-45.684	166.965	89.415	5.0
3269429	2010	062	-45.772	166.555	5.000	3.6
3270055	2010	063	-44.298	167.032	5.000	3.7
3270861	2010	065	-45.492	166.730	9.120	3.7
3273161	2010	069	-44.815	167.423	5.000	3.5
3274567	2010	072	-44.287	166.955	2.000	3.5
3274724	2010	072	-45.529	166.659	12.000	4.9
3275824	2010	075	-45.430	166.957	59.410	4.0
3275857	2010	075	-43.062	173.006	33.000	3.6
3275957	2010	075	-42.120	173.836	12.000	4.0
3276884	2010	077	-45.646	167.213	5.000	3.5
3277280	2010	077	-41.243	172.766	144.901	3.9
3277943	2010	079	-45.074	167.436	63.685	4.3
3278236	2010	079	-44.257	167.739	2.000	3.1
3281249	2010	085	-45.109	167.480	114.562	3.7
3283355	2010	089	-44.362	167.067	12.000	3.2
3284490	2010	091	-40.817	172.956	215.673	3.6
3285425	2010	093	-43.240	172.746	33.000	3.8
3287387	2010	097	-44.869	170.940	12.000	3.6
3287792	2010	098	-41.370	172.856	115.590	4.0
3288443	2010	099	-44.609	168.120	91.969	3.2
3291715	2010	105	-40.653	173.334	144.549	3.5
3293147	2010	108	-45.109	167.462	70.767	4.5
3295053	2010	112	-40.575	174.168	80.485	3.5
3296413	2010	115	-45.092	167.437	102.686	3.5
3296708	2010	115	-42.123	173.804	12.000	3.9
3299011	2010	119	-44.173	168.389	2.000	3.1
3299270	2010	120	-41.262	172.831	117.499	4.2
3299486	2010	120	-40.765	173.915	110.232	4.4
3301720	2010	125	-40.690	174.121	85.442	3.7
3301978	2010	125	-45.230	166.946	12.000	3.5
3302061	2010	125	-45.157	166.801	12.000	4.5
3302132	2010	125	-44.346	168.120	5.000	3.2
3302261	2010	126	-45.073	167.610	121.619	3.7
3302929	2010	127	-44.593	168.119	88.338	3.2
3303093	2010	127	-44.874	167.477	5.000	3.6
3306416	2010	133	-42.787	172.431	23.211	3.1
3306930	2010	134	-44.583	168.114	91.003	3.1
3309345	2010	139	-45.111	167.226	39.702	3.7
3310436	2010	141	-41.253	172.834	141.856	3.6

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Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3312931	2010	145	-44.611	168.125	88.426	3.5
3313020	2010	146	-43.380	172.533	12.000	3.1
3315096	2010	150	-42.575	172.916	12.000	3.9
3315329	2010	150	-43.598	170.629	5.000	3.0
3316380	2010	152	-44.670	168.120	89.393	3.0
3316474	2010	152	-42.476	173.329	53.212	4.3
3316477	2010	152	-42.550	173.352	50.625	3.4
3317333	2010	154	-44.188	168.348	5.000	3.2
3317665	2010	155	-41.155	173.446	110.390	3.6
3322168	2010	163	-44.400	168.288	5.000	3.3
3323536	2010	166	-41.286	172.596	189.853	3.6
3326488	2010	171	-45.088	167.332	126.430	3.8
3326729	2010	171	-45.139	167.311	85.557	3.6
3327324	2010	173	-45.120	167.328	88.927	3.7
3327568	2010	173	-44.154	168.585	2.000	3.3
3327688	2010	173	-46.838	166.324	33.000	3.8
3327812	2010	173	-45.135	167.368	85.541	3.8
3327852	2010	174	-45.429	166.753	12.000	3.9
3328970	2010	176	-45.685	166.634	5.000	3.5
3329809	2010	177	-42.707	171.705	5.000	3.5
3332362	2010	182	-44.233	171.569	12.000	3.1
3336735	2010	190	-45.724	166.534	12.000	3.5
3338289	2010	193	-40.869	172.860	218.186	4.2
3338766	2010	194	-40.512	173.562	142.791	4.9
3339606	2010	125	-45.153	166.759	12.000	3.9
3341085	2010	198	-43.675	168.284	5.000	3.5
3341372	2010	199	-41.239	172.799	154.449	3.6
3341682	2010	200	-45.213	167.239	76.964	4.4
3344300	2010	205	-42.819	172.298	12.000	3.0
3346364	2010	208	-45.075	167.317	101.166	4.6
3346371	2010	208	-45.036	167.412	116.906	3.7
3346616	2010	209	-45.164	167.340	118.585	4.0
3348741	2010	213	-46.020	166.288	20.000	3.5
3349104	2010	214	-41.264	172.950	114.212	4.3
3350372	2010	216	-45.139	167.390	87.999	4.4
3352922	2010	221	-45.039	167.476	94.879	4.5
3353879	2010	223	-45.106	167.147	33.000	3.4
3358415	2010	232	-40.617	173.204	162.130	3.7
3358526	2010	232	-45.200	166.868	33.000	3.7
3358599	2010	232	-42.825	171.810	5.000	3.7
3362156	2010	239	-43.210	171.991	5.000	3.3
3363606	2010	241	-41.166	173.969	61.378	4.0
3364383	2010	243	-41.273	172.551	209.676	3.6
3364515	2010	243	-44.171	168.609	5.000	3.3
3365555	2010	245	-44.304	168.091	5.000	3.4
3365975	2010	246	-44.489	168.143	12.000	3.7
3366206	2010	246	-43.603	172.385	5.000	3.9
3366305	2010	246	-43.604	172.346	5.000	3.9
3366329	2010	246	-43.546	172.374	5.000	3.9
3366506	2010	247	-43.548	172.422	12.000	3.8
3366542	2010	247	-43.748	172.310	5.000	3.9
3366601	2010	247	-41.190	172.573	215.865	3.8
3366669	2010	247	-43.507	172.201	5.947	3.2
3366754	2010	247	-43.595	172.190	12.646	4.1
3366806	2010	247	-43.579	171.963	5.000	3.7
3366920	2010	248	-43.563	171.928	3.117	3.8
3367084	2010	248	-43.552	171.916	10.244	3.3
3367309	2010	248	-43.596	171.852	5.000	3.3
3367331	2010	248	-43.609	172.219	5.000	3.2
3367453	2010	248	-43.468	172.166	7.223	3.5
3367923	2010	249	-43.566	171.880	7.858	3.5
3367932	2010	249	-41.260	172.622	213.314	3.7
3368008	2010	249	-43.526	171.879	10.020	3.9
3368099	2010	250	-41.346	174.529	40.330	3.8
3368110	2010	250	-41.353	174.535	39.662	3.7
3368872	2010	251	-45.147	167.349	91.204	3.8
3369037	2010	251	-43.634	171.975	5.000	3.0
3369555	2010	252	-43.484	172.122	5.000	3.0
3369677	2010	252	-43.587	172.025	9.251	3.5
3370794	2010	254	-43.540	171.876	10.464	4.1

Continued on next page...

Table A.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3370904	2010	255	-43.453	172.121	6.311	3.3
3371267	2010	255	-43.599	172.107	5.000	3.2
3371307	2010	255	-42.450	173.696	10.157	3.6
3372347	2010	258	-42.450	173.706	9.925	3.7
3372458	2010	258	-42.441	173.696	13.599	3.7
3375021	2010	246	-43.652	172.413	5.000	4.5
3375052	2010	248	-43.493	172.170	6.004	3.1
3375623	2010	248	-43.285	171.942	7.918	3.8
3375633	2010	248	-43.296	171.935	8.206	3.3
3375648	2010	249	-43.430	172.139	5.000	3.2
3376108	2010	264	-40.873	172.909	215.607	3.6
3376255	2010	250	-43.549	171.936	5.000	3.1
3376734	2010	248	-43.590	172.178	13.465	3.7
3376937	2010	266	-45.342	166.857	12.000	3.6
3376944	2010	248	-43.616	172.607	5.000	3.4
3377329	2010	248	-43.563	172.387	5.000	3.1
3378555	2010	248	-43.625	172.387	5.000	3.1
3379185	2010	270	-42.479	173.035	9.697	4.3
3379423	2010	248	-43.623	172.240	5.000	3.1
3381332	2010	274	-41.630	174.092	12.000	3.6
3384822	2010	281	-41.336	173.723	73.492	4.0
3386185	2010	248	-43.586	172.375	5.000	3.2
3387174	2010	285	-45.043	167.401	111.853	4.7
3389489	2010	248	-43.558	171.891	5.000	3.4
3390981	2010	248	-43.572	172.385	5.000	3.3
3392078	2010	248	-43.583	172.186	5.000	3.0
3392092	2010	248	-43.543	171.887	8.899	3.7
3392139	2010	248	-43.621	172.418	5.000	3.4
3392641	2010	248	-43.606	172.068	5.000	3.0
3392673	2010	293	-41.237	173.319	102.931	4.3
3393152	2010	247	-43.659	172.440	5.000	3.0
3393206	2010	294	-45.536	166.645	12.000	3.5
3394547	2010	297	-44.954	167.473	64.402	3.5
3400618	2010	305	-42.480	172.962	12.000	3.6
3401807	2010	307	-40.952	172.813	241.865	3.6
3405049	2010	314	-45.164	167.378	118.524	3.7
3407572	2010	319	-45.698	166.461	5.000	3.9
3408796	2010	321	-40.903	172.897	220.197	3.6
3412101	2010	328	-45.390	166.669	26.097	3.7
3413579	2010	331	-45.714	166.573	5.000	3.7
3414693	2010	333	-45.231	167.334	70.062	3.8
3415424	2010	335	-45.176	167.353	135.442	4.1
3416605	2010	337	-41.436	173.335	82.453	3.7
3417589	2010	339	-41.175	174.639	33.000	3.7
3420141	2010	345	-46.285	166.178	12.000	4.0
3420248	2010	345	-45.451	167.107	53.942	4.0
3420825	2010	346	-45.113	167.384	97.743	4.3
3421891	2010	349	-41.753	172.210	100.618	3.5
3422970	2010	349	-41.266	172.563	225.037	3.5
3437269	2010	360	-40.624	173.301	174.468	3.5
3438304	2010	362	-40.541	174.377	78.451	3.7
3438312	2010	362	-45.334	166.761	12.000	4.3

**measurements within the incidence angle $\leq 35^\circ$ are given. Note that events with magnitude < 3 were used in the schist anisotropy study. All measurements and event information can be obtained on request.

A.3 Supplementary figures of chapter 4

Fig. A.2 Equal area stereographs display backazimuthal variations in splitting parameters (the best filter measurements) at some stations. The radial position of these plots indicates the incidence angle, which ranges from $0 - 40^\circ$. Split bars are plotted in the direction of BAZ and at the incidence angle. Note that splitting parameters from near by stations show no systematic variations with either BAZ or incidence angle.

Fig. A.3 displays the percent velocity anisotropy estimation from shallow events (depths ≤ 70 km) in southernmost part of the Hikurangi subduction zone using the δt measured at LTZ (yellow) and KHZ (blue).

Fig. A.4 shows some examples of stations that show scatter in δt with depth in northern and southern South Island.

Figure A.2 Examples of the equal area stereographs display variations in station splitting parameters (best filter splitting measurements) with BAZ and incidence angle. The radial position of these plots indicates the incidence angle, which ranges from $0 - 40^\circ$. Split bars are plotted in the direction of BAZ and at the incidence angle. The direction of each bar denotes the ϕ and the size corresponds to the δt . The upper right circular histogram of each plot shows all the individual measurements together and the number indicates the number of measurements that satisfy the best filter criterion (see section 4.2).

Figure A.3 Same as figure caption in Fig. 4.7, but velocity anisotropy is calculated from shallow slab events in the northern South Island. Blue and yellow color dots denote δ measurements at KHZ and LTZ respectively.

Figure A.4 Scatter in δt with depth observed at some stations around subduction zones in northern and southern South Island.

Appendix B

Offshore investigation

B.1 Station information

Table B.1 Station information of the OBS deployments.

station	sensor	st.lat [°]	st.lon. [°]	bathymetry (m)	sensor orie.([°]) [§]
NZ01	S116	-44.498	165.002	4682	-**
NZ02	S128	-43.997	166.508	3816	36.3
NZ03	S132	-43.002	165.999	4225	292.5
NZ04	S142	-41.999	165.508	4470	205.7
NZ05	S125	-40.644	167.250	1188	264.3
NZ06	S113	-44.986	168.001	859	197.5
NZ07	S115	-40.250	169.000	852	89.0
NZ08	S098	-40.998	168.501	995	355.3
NZ09	S112	-41.651	167.900	1335	315.3
NZ10	S082	-43.201	168.000	1193	40.6
NZ11	S124	-43.305	168.850	911	174.5
NZ12	S127	-43.002	169.598	625	116.4
NZ13	S134	-42.499	168.999	1165	93.0
NZ14	S114	-42.651	169.950	633	346.5
NZ15	S119	-42.151	170.400	561	249.5
NZ16	S096	-41.749	169.249	1063	54.8
NZ17	S118	-41.500	170.500	595	-**
NZ18	S097	-41.000	170.000	805	321.7
NZ19	S120	-40.000	170.900	549	243.2
NZ20	S122	-40.200	170.750	714	154.2
NZ21	S080	-39.501	170.000	676	237.7
NZ22	S135	-44.501	173.998	776	233.5
NZ23	S126	-45.149	174.852	1091	13.6
NZ24	S131	-45.499	176.003	1507	283.0
NZ25	S117	-45.998	174.750	1727	36.6
NZ26	S121	-46.998	174.498	1316	108.8
NZ27	S123	-46.501	173.249	1544	216.8
NZ28	S133	-45.964	171.751	1328	330.1
NZ29	S130	-45.247	172.254	1383	43.9
NZ30	S129	-45.498	173.501	1416	57.4

[§]sensor orientations (clockwise from north) as determined by *Stachnik et al.* [2012] from rayleigh-wave particle motion analysis

**missing sensors

B.2 Measurements selection criteria

AB-grade splitting parameters were used for the analysis (AB-grade and best filter criteria are given in section 3.2.3). Splitting parameters are graded as good quality if the $\lambda_{max} > 5$. Here, λ_{max} denotes the number of scaled energy contours around the best resolved splitting parameters (section 3.2.3). Measurements with $\delta t \leq 0.7$ s were selected to avoid ambiguous splitting parameters.

B.3 Events with AB-grade measurements

Table B.2 Events with AB quality high-grade measurement[‡].

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3041931	2009	038	-41.233	172.566	213.946	3.6
3042565	2009	039	-45.537	167.071	76.165	3.9
3043284	2009	041	-44.798	167.367	5.000	3.7
3045760	2009	046	-45.589	166.945	12.000	3.7
3047503	2009	049	-45.096	167.389	117.456	3.9
3049002	2009	052	-46.305	165.643	12.000	4.1
3049393	2009	053	-45.128	166.627	12.000	4.0
3049778	2009	054	-41.915	173.728	40.573	3.6
3050923	2009	056	-43.504	171.544	12.000	3.5
3051685	2009	058	-43.063	173.362	26.056	3.8
3052092	2009	059	-41.007	174.766	32.583	3.3
3054325	2009	063	-43.214	172.156	5.000	4.1
3055693	2009	066	-45.466	167.424	11.641	3.7
3057072	2009	069	-44.970	167.536	57.478	3.3
3057524	2009	069	-42.968	171.446	5.000	3.5
3057877	2009	070	-41.688	174.477	54.052	4.3
3058975	2009	072	-45.095	167.398	117.789	3.8
3059063	2009	072	-41.958	174.110	15.880	3.3
3059454	2009	073	-44.735	167.889	96.807	3.1
3059515	2009	073	-40.542	173.522	158.489	4.1
3059728	2009	074	-44.347	170.115	12.755	3.1
3060059	2009	074	-45.106	167.483	114.984	3.1
3060760	2009	076	-39.533	174.311	235.339	4.2
3061005	2009	076	-40.434	173.543	158.239	3.6
3062725	2009	079	-44.950	167.370	33.000	3.1
3063032	2009	080	-40.309	174.092	117.688	3.7
3063549	2009	081	-40.431	173.491	155.490	4.9
3064359	2009	083	-45.150	167.110	12.000	4.5
3064368	2009	083	-41.205	173.755	70.267	3.3
3064812	2009	083	-41.788	173.841	15.392	3.3
3065595	2009	085	-45.061	167.422	97.395	4.7
3066319	2009	086	-44.254	170.018	12.000	3.1
3066543	2009	087	-45.211	167.277	100.059	3.5
3066555	2009	087	-45.069	167.453	110.902	3.0
3067162	2009	088	-44.310	168.326	5.000	3.7
3067199	2009	088	-44.269	170.010	14.379	3.2
3068099	2009	090	-41.499	172.888	123.965	4.6
3068852	2009	091	-45.025	167.393	125.892	3.6
3069768	2009	093	-44.824	167.658	85.710	3.2
3070857	2009	095	-42.456	172.979	8.880	3.2
3071236	2009	096	-42.794	173.136	5.000	3.2
3071401	2009	096	-41.660	172.901	91.317	4.5
3072643	2009	098	-44.210	168.412	5.000	3.7

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Table B.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3072794	2009	099	-44.563	173.752	33.000	3.8
3073334	2009	100	-45.029	167.537	127.050	3.5
3074226	2009	102	-44.401	169.595	12.000	3.0
3074364	2009	102	-40.172	173.769	161.285	3.8
3074557	2009	102	-45.035	167.465	98.712	3.1
3074691	2009	103	-41.469	172.876	121.837	3.9
3074697	2009	103	-43.490	171.600	5.000	4.3
3075621	2009	104	-46.143	167.789	12.000	4.5
3077937	2009	109	-40.402	173.590	154.805	3.7
3080021	2009	113	-41.377	172.966	116.455	3.7
3080807	2009	114	-41.284	172.678	208.877	3.8
3083936	2009	120	-40.371	173.383	192.020	3.6
3084790	2009	122	-44.259	170.003	14.189	3.6
3087732	2009	128	-43.152	170.960	5.000	4.0
3088615	2009	129	-42.479	174.703	33.000	3.1
3088974	2009	130	-40.075	173.943	162.557	3.9
3090296	2009	133	-45.859	167.204	12.000	4.3
3091656	2009	135	-41.134	173.919	67.400	3.7
3092792	2009	137	-42.260	171.335	5.000	3.6
3095999	2009	143	-40.555	175.458	34.813	3.9
3096079	2009	143	-45.410	167.196	136.906	3.6
3096871	2009	144	-45.158	167.356	120.805	3.9
3097975	2009	146	-44.504	169.860	5.000	4.3
3099203	2009	149	-41.944	174.027	19.783	3.5
3102845	2009	155	-41.060	174.540	44.080	4.0
3104352	2009	158	-41.071	174.557	43.478	3.9
3106816	2009	163	-40.375	174.950	5.000	3.7
3107010	2009	163	-43.672	171.604	5.000	3.3
3107602	2009	164	-45.214	167.304	105.020	3.8
3107978	2009	165	-43.734	171.607	5.000	3.7
3108186	2009	165	-44.827	167.250	33.000	3.7
3108837	2009	166	-40.555	173.419	165.474	4.1
3109454	2009	167	-42.038	172.925	75.936	3.3
3112716	2009	173	-40.943	172.834	193.891	3.9
3115714	2009	178	-42.502	169.801	33.000	4.8
3118326	2009	183	-41.550	172.995	91.464	4.2
3118345	2009	183	-41.578	172.993	91.361	4.6
3119748	2009	186	-45.036	167.393	110.561	3.5
3120467	2009	188	-43.022	172.722	25.211	3.0
3122085	2009	181	-42.026	173.927	19.720	3.4
3122205	2009	191	-43.239	170.909	5.000	3.5
3122292	2009	191	-40.985	172.916	174.270	4.2
3123338	2009	193	-40.917	173.065	172.778	3.5
3124008	2009	194	-42.240	173.892	33.000	3.7
3124791	2009	196	-40.179	173.556	167.322	5.4
3124795	2009	196	-46.035	165.883	12.000	5.3
3124798	2009	196	-45.835	166.435	5.000	5.6
3124807	2009	196	-46.196	165.844	33.000	5.0
3124856	2009	196	-46.114	165.803	12.000	3.9
3124997	2009	196	-45.450	166.641	5.000	4.8
3125012	2009	196	-46.036	166.087	12.000	3.9
3125061	2009	196	-45.521	167.033	65.012	5.1
3125075	2009	197	-45.640	166.535	12.000	4.3
3125083	2009	197	-46.288	165.867	5.000	5.5
3125111	2009	197	-45.770	166.544	5.000	4.9
3125112	2009	197	-45.400	167.000	12.000	5.1
3125133	2009	197	-46.370	165.613	12.000	5.0
3125179	2009	197	-45.704	166.590	12.000	4.7
3125184	2009	197	-46.246	166.014	15.852	4.8
3125200	2009	197	-45.702	166.551	5.000	5.6
3125201	2009	197	-45.897	166.211	33.000	5.4
3125217	2009	197	-46.154	165.936	12.000	4.6
3125237	2009	197	-45.544	166.888	12.000	3.8
3125242	2009	197	-45.587	166.106	12.000	3.9
3125250	2009	197	-46.110	165.876	19.889	3.5
3125262	2009	197	-45.685	166.550	4.515	4.4
3125394	2009	197	-46.167	166.100	30.000	5.3
3125459	2009	197	-46.184	165.925	5.000	4.4
3125523	2009	197	-46.186	166.121	33.000	5.6
3125534	2009	197	-46.067	165.971	12.000	4.3

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Table B.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3125543	2009	197	-46.099	165.846	12.000	3.6
3125554	2009	197	-45.556	166.577	22.278	3.9
3125563	2009	198	-46.121	166.060	26.758	3.8
3125571	2009	198	-46.108	165.931	12.000	4.8
3125584	2009	198	-45.464	166.614	19.540	4.8
3125593	2009	198	-46.217	165.795	33.000	4.0
3125608	2009	198	-46.139	166.065	31.528	3.5
3125633	2009	198	-46.207	166.013	30.580	4.7
3125673	2009	198	-46.310	165.873	33.000	3.5
3125686	2009	198	-45.727	166.531	22.396	3.9
3125689	2009	198	-45.395	166.706	21.864	4.9
3125691	2009	198	-45.331	166.502	12.000	5.6
3125693	2009	198	-45.374	166.602	21.326	3.6
3125699	2009	198	-45.394	166.573	20.240	3.8
3125707	2009	198	-45.384	166.703	21.886	3.5
3125714	2009	198	-45.367	166.626	21.089	4.4
3125782	2009	198	-45.646	166.967	8.729	3.6
3125784	2009	198	-45.503	165.727	33.000	3.8
3125829	2009	198	-46.019	166.482	32.839	3.8
3125831	2009	198	-45.428	166.625	21.897	4.3
3125892	2009	198	-46.180	165.717	33.000	3.9
3125939	2009	198	-45.420	166.701	21.574	4.5
3125976	2009	198	-41.574	173.590	63.677	3.2
3126009	2009	199	-45.812	166.344	19.473	3.9
3126016	2009	199	-45.370	166.637	21.158	4.8
3126021	2009	199	-46.084	166.098	27.012	4.0
3126025	2009	199	-45.363	166.658	21.554	4.8
3126068	2009	199	-46.033	166.262	28.432	3.6
3126070	2009	199	-46.070	166.294	31.831	3.7
3126085	2009	199	-46.307	165.755	33.000	3.7
3126095	2009	199	-46.149	166.008	33.000	3.8
3126108	2009	199	-45.523	166.679	21.855	3.5
3126113	2009	199	-45.496	166.677	21.684	3.8
3126121	2009	199	-46.242	165.973	29.645	3.6
3126128	2009	199	-45.465	166.694	20.259	3.6
3126147	2009	199	-45.478	166.285	12.000	3.8
3126167	2009	199	-46.017	166.481	32.855	4.5
3126201	2009	199	-45.652	166.500	20.362	4.5
3126239	2009	199	-46.169	166.184	32.848	4.0
3126249	2009	199	-46.154	166.049	30.066	4.3
3126312	2009	199	-45.548	166.645	21.462	4.5
3126393	2009	199	-45.501	166.542	21.470	4.3
3126413	2009	199	-46.150	165.816	12.000	3.5
3126418	2009	199	-45.481	166.611	21.308	3.7
3126438	2009	200	-46.155	165.883	12.000	3.7
3126520	2009	200	-45.379	166.097	12.000	5.2
3126535	2009	200	-46.136	166.094	29.870	3.7
3126568	2009	200	-45.682	166.528	21.560	4.1
3126570	2009	200	-44.522	175.686	33.000	3.9
3126592	2009	200	-45.337	166.164	12.000	5.2
3126670	2009	200	-45.714	166.493	20.943	3.5
3126686	2009	200	-45.396	166.722	17.842	3.6
3126710	2009	200	-45.602	165.763	33.000	3.5
3126729	2009	200	-45.545	166.654	22.410	4.4
3126801	2009	200	-45.673	166.529	18.602	3.7
3126817	2009	200	-46.121	165.975	31.502	3.7
3126909	2009	201	-46.192	165.861	12.000	3.6
3127061	2009	201	-45.464	166.746	21.983	3.6
3127091	2009	201	-44.684	175.528	33.000	3.4
3127103	2009	201	-45.422	166.585	22.206	3.8
3127139	2009	201	-46.169	166.143	27.544	3.6
3127319	2009	202	-46.042	166.170	27.031	3.7
3127345	2009	202	-45.864	165.851	33.000	4.5
3127489	2009	202	-46.086	166.071	29.436	4.0
3127579	2009	202	-47.313	165.269	33.000	4.1
3127607	2009	202	-46.015	166.148	28.158	4.1
3127631	2009	202	-46.192	165.820	12.000	4.5
3127703	2009	202	-45.432	165.852	33.000	4.1
3127714	2009	202	-45.881	166.438	26.520	4.0
3127740	2009	202	-46.173	166.047	33.000	4.0

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Table B.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3127929	2009	203	-45.425	166.627	12.000	3.8
3128006	2009	203	-46.353	165.878	33.000	4.5
3128140	2009	203	-45.473	166.753	22.952	3.6
3128237	2009	203	-46.216	165.875	33.000	3.8
3128341	2009	204	-42.668	172.595	33.000	3.3
3128439	2009	204	-44.569	175.549	33.000	3.9
3128653	2009	204	-45.739	165.664	12.000	3.9
3128712	2009	204	-45.409	166.730	22.087	3.8
3128728	2009	204	-46.099	165.821	12.000	3.7
3128836	2009	204	-45.410	166.728	25.015	3.9
3128844	2009	204	-45.807	165.821	33.000	4.0
3128890	2009	205	-46.136	166.246	33.000	3.5
3128927	2009	205	-46.156	165.805	33.000	4.3
3128996	2009	205	-45.656	166.484	16.840	3.5
3129028	2009	205	-45.805	165.747	33.000	3.8
3129136	2009	205	-45.393	166.722	23.897	4.6
3129173	2009	205	-46.156	165.882	33.000	3.7
3129198	2009	205	-44.615	168.159	80.070	3.2
3129235	2009	205	-44.564	175.549	33.000	3.3
3129381	2009	206	-45.399	166.612	19.687	3.6
3129388	2009	206	-45.385	166.591	19.743	3.8
3129805	2009	207	-45.868	166.278	21.556	3.7
3129848	2009	207	-46.130	165.798	12.000	4.6
3129894	2009	207	-41.292	172.704	158.588	3.6
3130140	2009	207	-45.356	166.722	22.340	3.5
3130239	2009	208	-44.575	175.493	33.000	3.4
3130282	2009	208	-45.056	166.940	33.000	3.7
3130415	2009	208	-46.047	166.050	30.909	3.5
3130486	2009	208	-45.734	166.409	20.095	3.6
3130487	2009	208	-44.374	168.030	5.000	4.1
3130648	2009	208	-45.754	165.818	12.000	3.8
3130676	2009	209	-45.471	166.657	21.433	3.6
3130851	2009	209	-42.158	173.721	40.950	3.0
3131288	2009	210	-45.117	167.204	33.000	3.2
3131425	2009	210	-46.048	166.161	33.000	5.2
3131438	2009	210	-46.169	165.899	33.000	5.0
3131449	2009	210	-46.168	165.884	33.000	3.9
3131501	2009	211	-45.539	166.646	20.917	4.0
3131547	2009	211	-45.623	166.417	22.697	5.0
3131612	2009	211	-45.657	166.461	19.171	4.8
3131648	2009	211	-46.046	165.904	27.275	3.6
3131736	2009	211	-45.618	166.469	19.518	3.7
3131885	2009	211	-46.169	165.950	29.097	4.3
3131896	2009	211	-45.646	166.448	21.977	4.7
3132026	2009	212	-39.862	175.192	66.499	3.7
3132105	2009	212	-45.520	166.691	23.874	3.8
3132302	2009	212	-45.430	166.756	24.235	4.7
3132372	2009	212	-45.446	166.710	24.893	4.3
3132552	2009	213	-46.063	165.971	28.807	3.6
3132601	2009	213	-44.971	167.535	127.785	3.6
3132894	2009	213	-45.776	165.865	33.000	3.5
3133200	2009	214	-46.105	166.305	28.945	3.9
3133476	2009	214	-44.584	175.504	33.000	3.5
3133651	2009	198	-46.145	166.126	33.787	4.1
3133655	2009	198	-46.147	166.110	32.452	4.4
3133914	2009	215	-46.248	166.279	33.000	3.8
3133965	2009	215	-45.134	166.458	12.000	5.0
3134000	2009	215	-45.479	166.620	20.410	3.8
3134001	2009	215	-45.799	165.965	32.123	3.6
3134115	2009	216	-41.811	174.345	29.308	4.2
3134119	2009	200	-45.453	165.762	33.000	4.1
3134155	2009	216	-45.382	166.696	22.819	4.3
3134378	2009	216	-45.498	166.660	22.454	3.8
3134481	2009	216	-44.437	168.008	5.000	3.5
3134616	2009	200	-45.413	166.714	20.706	4.0
3134710	2009	217	-45.508	166.328	12.000	5.2
3134768	2009	217	-46.146	166.117	27.712	4.8
3134829	2009	217	-45.564	166.156	12.000	4.8
3135253	2009	218	-45.442	166.676	21.755	3.5
3135313	2009	218	-45.476	166.746	21.993	3.8

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Table B.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3135487	2009	200	-46.106	166.206	32.296	3.6
3135552	2009	219	-45.858	166.425	39.614	3.7
3135583	2009	202	-46.091	166.006	29.150	4.4
3135671	2009	219	-46.080	165.917	25.205	3.6
3135754	2009	219	-45.551	166.658	20.658	4.5
3135791	2009	219	-45.551	166.667	18.191	4.8
3135990	2009	219	-46.028	166.051	26.633	3.9
3136205	2009	220	-46.142	165.870	33.000	4.3
3136426	2009	198	-46.168	165.997	28.758	3.5
3136489	2009	221	-44.571	175.546	33.000	4.0
3136656	2009	221	-45.236	166.664	22.950	4.7
3136713	2009	221	-45.542	166.627	23.347	4.8
3136769	2009	221	-46.229	166.073	33.000	3.9
3136860	2009	221	-45.754	165.807	12.000	4.2
3137460	2009	223	-45.963	166.434	30.215	3.7
3137692	2009	223	-45.391	166.702	21.414	3.5
3137861	2009	198	-45.380	166.703	21.864	3.7
3138383	2009	198	-46.299	165.797	33.000	3.9
3138587	2009	225	-45.339	166.767	26.962	3.7
3139054	2009	226	-45.061	167.400	112.849	3.3
3139106	2009	226	-43.256	170.790	3.000	4.1
3139357	2009	209	-44.545	175.224	12.000	3.7
3139393	2009	198	-46.101	165.916	26.398	3.6
3139869	2009	228	-41.362	172.767	136.649	3.1
3140195	2009	229	-45.552	166.656	22.020	4.1
3140272	2009	229	-46.015	166.108	28.419	3.9
3140382	2009	229	-42.722	173.910	12.000	3.7
3140483	2009	229	-45.923	165.779	29.158	5.1
3140527	2009	229	-45.915	165.840	12.000	4.4
3141039	2009	230	-44.423	168.327	5.000	4.0
3141091	2009	231	-44.534	175.380	33.000	3.3
3141178	2009	231	-46.200	165.846	33.000	3.7
3141301	2009	231	-44.586	175.530	33.000	3.1
3141338	2009	231	-40.530	173.610	118.346	3.8
3141375	2009	231	-41.726	172.188	100.391	3.3
3141436	2009	231	-45.682	166.458	20.648	4.8
3141439	2009	231	-45.685	166.487	19.718	3.6
3141478	2009	231	-46.032	166.272	33.000	3.6
3141540	2009	232	-40.338	174.038	127.364	3.5
3141684	2009	232	-45.980	165.880	33.000	4.9
3141695	2009	232	-46.067	166.166	30.888	4.5
3141703	2009	232	-44.594	168.173	80.351	3.3
3141789	2009	232	-44.595	175.581	33.000	3.6
3142644	2009	234	-41.161	172.859	145.078	3.5
3142845	2009	235	-46.151	165.778	33.000	3.5
3142936	2009	235	-44.983	167.458	87.478	3.5
3143207	2009	235	-45.946	165.858	29.908	3.7
3143313	2009	217	-44.580	175.597	33.000	3.3
3143345	2009	236	-46.279	165.848	33.000	4.4
3143358	2009	199	-45.475	166.638	22.118	4.1
3143705	2009	237	-45.150	167.280	121.351	3.8
3143806	2009	237	-46.115	166.047	26.287	5.5
3143822	2009	237	-46.119	166.114	30.687	4.6
3144142	2009	237	-46.117	166.091	29.162	4.3
3145159	2009	239	-41.442	174.709	33.910	5.2
3145245	2009	239	-45.100	167.368	112.461	3.1
3145383	2009	240	-45.499	166.789	17.847	4.0
3145684	2009	240	-45.510	166.481	19.859	4.0
3146008	2009	241	-45.461	166.346	19.526	4.2
3146016	2009	241	-44.933	167.482	74.659	3.0
3146610	2009	242	-45.429	166.770	23.174	3.9
3147149	2009	243	-46.020	166.186	29.116	3.5
3147189	2009	243	-45.681	166.495	21.568	4.3
3147386	2009	243	-41.847	173.309	84.966	3.1
3147530	2009	244	-44.483	166.926	5.000	4.0
3148588	2009	245	-40.926	172.922	187.429	4.0
3150084	2009	249	-44.169	168.623	5.000	3.9
3150087	2009	249	-46.150	165.812	30.569	3.5
3150137	2009	249	-44.991	167.590	69.385	3.2
3150190	2009	249	-44.240	167.708	12.000	3.6

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Table B.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3150361	2009	249	-45.985	166.143	29.931	3.8
3150770	2009	250	-45.628	167.294	11.769	3.3
3150827	2009	250	-41.309	172.827	139.828	3.8
3151327	2009	251	-46.070	165.976	29.168	4.5
3151404	2009	251	-41.227	172.680	139.723	3.0
3151539	2009	252	-45.535	166.695	23.847	4.0
3151852	2009	252	-45.546	166.690	20.138	4.9
3152083	2009	253	-40.725	174.347	72.184	4.1
3152626	2009	254	-46.123	165.837	33.000	4.1
3152685	2009	254	-40.813	174.620	44.762	3.7
3153108	2009	255	-41.311	174.546	41.211	3.6
3153511	2009	256	-43.053	172.816	29.129	3.4
3153651	2009	243	-46.343	166.004	33.000	3.7
3153729	2009	257	-45.481	166.733	21.285	3.5
3154556	2009	258	-45.302	166.797	12.000	4.4
3154683	2009	259	-45.985	166.632	19.576	3.6
3154988	2009	259	-46.201	165.958	33.000	4.1
3155299	2009	260	-45.094	167.368	123.759	4.2
3156081	2009	262	-44.843	166.935	12.000	3.9
3156088	2009	262	-45.118	167.554	5.000	3.8
3157492	2009	265	-40.651	174.874	41.600	3.6
3157795	2009	265	-40.117	174.882	17.602	3.7
3157830	2009	265	-45.058	167.476	117.902	3.5
3158054	2009	266	-43.651	168.212	33.000	3.1
3158128	2009	266	-44.277	167.929	12.000	4.4
3158502	2009	267	-41.748	172.473	80.029	3.2
3159659	2009	269	-44.466	168.171	12.000	4.5
3159733	2009	270	-45.709	166.541	16.656	3.8
3160421	2009	271	-46.125	166.093	27.798	4.2
3161106	2009	272	-44.572	175.481	33.000	3.7
3161447	2009	273	-45.305	166.844	27.355	3.7
3161601	2009	274	-46.089	166.004	25.565	3.6
3161674	2009	274	-46.088	166.032	24.827	4.2
3161700	2009	274	-46.144	165.923	12.000	4.7
3161868	2009	274	-46.108	165.941	26.416	4.1
3161988	2009	274	-45.392	166.700	21.507	3.6
3162137	2009	275	-44.592	168.173	81.789	3.1
3162293	2009	275	-46.065	165.923	29.578	3.8
3163125	2009	277	-44.577	175.344	33.000	3.2
3163469	2009	278	-40.966	173.431	89.585	3.1
3164709	2009	280	-46.088	166.244	26.738	3.7
3164782	2009	280	-42.325	173.660	58.120	3.1
3165019	2009	281	-45.606	166.775	16.462	4.7
3165850	2009	283	-46.035	165.916	28.392	3.8
3165869	2009	283	-41.162	174.602	41.898	4.8
3166610	2009	284	-45.545	166.745	20.234	3.6
3166997	2009	285	-45.062	167.386	92.545	3.6
3168258	2009	288	-44.761	167.299	5.000	3.6
3168423	2009	288	-45.604	166.766	18.208	3.5
3170073	2009	290	-45.083	167.454	94.385	4.1
3171001	2009	292	-45.148	167.089	33.000	4.3
3171550	2009	293	-45.665	166.876	62.468	3.6
3171833	2009	294	-41.064	172.958	155.883	3.6
3172220	2009	295	-41.285	172.933	117.325	3.7
3172529	2009	295	-41.096	173.597	92.472	3.5
3173062	2009	297	-46.116	165.796	12.000	4.0
3174178	2009	299	-45.889	166.541	18.847	3.6
3174442	2009	299	-45.134	167.331	114.108	3.7
3176786	2009	304	-44.001	169.068	12.000	4.4
3178306	2009	307	-42.187	172.632	2.000	4.2
3178429	2009	308	-42.048	172.761	89.719	4.5
3178435	2009	308	-45.552	166.825	19.317	3.5
3179193	2009	309	-45.457	166.706	22.504	3.7
3181530	2009	314	-42.476	172.891	42.801	3.5
3182457	2009	316	-45.486	166.755	21.047	3.7
3182458	2009	316	-45.463	166.746	21.672	4.0
3183512	2009	318	-46.087	166.049	27.984	3.8
3185298	2009	321	-41.261	173.149	122.563	3.2
3185741	2009	322	-44.680	167.492	5.000	3.7
3186184	2009	323	-45.161	168.875	5.000	3.7

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Table B.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3186228	2009	323	-43.218	172.871	33.000	4.2
3186604	2009	324	-47.230	165.755	33.000	4.2
3187010	2009	325	-45.021	167.431	89.211	3.5
3188308	2009	328	-44.603	168.197	88.074	3.0
3188423	2009	197	-45.861	165.870	12.000	3.7
3188874	2009	197	-45.512	166.612	19.765	4.1
3188888	2009	329	-46.120	166.051	25.584	4.3
3190183	2009	331	-46.059	165.976	23.551	3.6
3190570	2009	332	-43.339	171.961	8.759	3.5
3190843	2009	333	-45.258	166.970	33.000	3.8
3191284	2009	334	-45.741	166.717	12.000	3.6
3202873	2009	336	-44.941	167.103	20.512	3.7
3203616	2009	338	-42.700	175.697	33.000	3.1
3204025	2009	339	-40.805	172.910	209.131	4.0
3204445	2009	340	-45.082	167.322	117.553	3.5
3204538	2009	340	-42.061	174.443	28.002	3.0
3205351	2009	342	-46.011	166.840	81.918	4.1
3205412	2009	342	-41.633	172.342	106.367	3.3
3205715	2009	342	-40.562	174.853	48.911	5.1
3205722	2009	342	-45.783	165.893	33.000	4.3
3206052	2009	343	-45.117	167.373	95.279	3.8
3206940	2009	345	-41.083	174.318	67.175	3.7
3207634	2009	346	-40.555	174.844	48.993	4.2
3207977	2009	347	-46.073	165.961	12.000	4.4
3209307	2009	349	-44.383	168.188	5.000	4.6
3210836	2009	353	-44.385	168.190	2.000	3.2
3211310	2009	353	-45.000	167.494	97.385	3.6
3211830	2009	354	-45.733	166.221	2.000	4.0
3212649	2009	356	-40.997	174.683	43.612	4.3
3212678	2009	356	-44.344	167.020	12.000	3.9
3212744	2009	357	-44.285	166.995	12.000	3.7
3212774	2009	357	-44.418	167.058	12.000	4.1
3213027	2009	357	-45.464	166.649	5.000	3.7
3215170	2009	362	-45.693	166.242	5.000	3.8
3215421	2009	362	-44.432	167.017	12.000	3.9
3215677	2009	363	-44.404	167.038	5.000	4.6
3215699	2009	363	-44.427	167.040	5.000	4.1
3215741	2009	363	-44.400	167.025	12.000	3.5
3215786	2009	363	-44.249	166.901	12.000	3.9
3215798	2009	363	-44.400	167.040	5.000	4.1
3215901	2009	363	-44.289	167.009	5.000	3.7
3215902	2009	363	-44.370	167.004	12.000	3.7
3218624	2010	004	-44.286	166.981	12.000	3.7
3218632	2010	004	-44.326	166.986	12.000	4.1
3218873	2010	004	-44.350	167.035	12.000	3.6
3218875	2010	004	-44.288	166.942	5.000	3.5
3218911	2010	004	-44.415	167.036	5.000	3.7
3219107	2010	005	-44.336	167.049	12.000	3.5
3219127	2010	005	-44.330	167.007	12.000	3.8
3219727	2010	006	-45.066	167.404	88.520	3.8
3226288	2010	007	-43.589	168.092	5.000	3.0
3227595	2010	007	-40.379	174.225	90.911	4.4
3228164	2010	008	-45.133	167.244	84.824	3.8
3228266	2010	008	-41.112	174.739	32.995	3.7
3229512	2010	011	-45.771	166.593	5.000	3.6
3229793	2010	011	-45.101	167.342	116.846	3.7
3229901	2010	012	-44.167	168.445	5.000	3.7
3239748	2010	013	-44.351	167.030	5.000	3.6
3239807	2010	013	-44.292	167.057	5.000	4.0
3241870	2010	014	-44.133	168.387	5.000	4.6
3243091	2010	016	-43.200	173.495	33.000	3.8
3243510	2010	017	-45.138	167.411	109.896	3.1
3244402	2010	019	-44.877	167.654	78.381	3.5
3244959	2010	021	-46.050	169.862	14.461	3.2
3245861	2010	022	-45.479	166.694	12.000	4.6
3245913	2010	022	-44.164	168.423	5.000	4.1
3246039	2010	023	-41.203	172.747	169.818	3.5
3248116	2010	027	-44.370	168.003	5.000	3.9
3248616	2010	028	-41.242	172.757	146.179	3.2
3249826	2010	031	-41.984	173.206	55.878	3.1

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Table B.2 – Continued

cuspid	year	jul.day	lat.°	lon.°	depth	mag.
3250965	2010	033	-46.273	166.185	23.050	3.8
3251871	2010	035	-44.938	167.685	72.093	3.1
3252436	2010	036	-44.977	167.512	63.735	3.2
3252455	2010	036	-44.726	167.410	5.000	5.0
3254472	2010	040	-41.952	172.605	78.788	4.4
3260100	2010	043	-40.983	173.886	70.729	3.4
3314094	2009	196	-46.032	166.085	12.000	5.2
3316575	2009	196	-45.485	166.643	12.000	5.2
3321339	2009	196	-46.067	166.077	20.000	4.7
3324631	2009	196	-45.932	166.149	12.000	4.3
3327447	2009	196	-46.024	166.055	12.000	4.1
3328516	2009	196	-45.546	166.634	12.000	4.8
3330099	2009	196	-45.680	166.740	22.207	3.6
3333207	2009	196	-46.043	166.085	20.000	3.7
3340769	2009	196	-45.627	166.716	5.000	4.4
3341255	2009	196	-46.035	166.093	12.000	4.9
3348155	2009	196	-46.046	165.989	12.000	3.9
3348162	2009	196	-46.033	166.168	12.000	3.5
3348710	2009	196	-45.924	166.223	12.000	3.9
3350237	2009	196	-46.007	166.108	12.000	4.9
3350700	2009	196	-45.424	166.649	12.000	4.3
3350709	2009	196	-46.016	166.182	20.000	4.7

B.4 Combined analysis of splitting tomography and spatial averaging

The splitting data set used for this analysis consists of all AB-grade best filter measurements with $\delta t \leq 0.7$ s (see the electronic supplement provided with this thesis for measurement information).

Table B.3 Measurement information

combined analysis	
stations	56
no. events	2984
event-station pairs	7035

[‡]high grade measurements are defined as AB grade measurements with $\lambda_{max} > 5$. Measurements with $\delta t > 0.7$ s have been eliminated.

Appendix C

Southern NI investigation

C.1 Station information

Table C.1 Station information of the SAHKE transect in southern NI

station	st.lat. [°]	st.lon. [°]	elevation (m)	logger type	logger no.	sensor type
LE4	-41.3579	175.6919	197	Reftek	9298	40T
LTN6	-41.1033	175.3238	103	Reftek	9926	3ESP
T004	-41.3403	175.6688	47	Reftek	AC13	3ESP
T007	-41.3041	175.6513	48	-	AUS	40T
T010	-41.2520	175.5825	111	-	AUS	40T
T014	-41.2075	175.5063	145	-	AUS	40T
T016	-41.1893	175.4737	39	Reftek	AC40	40T
T018	-41.1715	175.4257	56	-	AUS	40T
T020	-41.1461	175.3850	23	Reftek	9F96	40T
T022	-41.1251	175.3497	31	-	AUS	40T

C.2 Event information of the acceptable quality splitting measurements

Table C.2 Event information of the good/fair[†] quality splitting measurements.

station	phase	st.lon.°	st.lat.°	event lon.°	event lat.°	day	origin time (s)	depth (km)	mag.	BAZ°
T016	SKS	175.474	-41.189	-171.836	52.050	2011/06/24	03:09:39.000	52.000	7.3	7.812
T016	SKS	175.474	-41.189	159.843	52.498	2010/07/30	03:56:13.000	23.000	6.3	350.492
T016	SKS	175.474	-41.189	-97.782	16.396	2010/06/30	07:22:27.000	20.000	6.3	75.519
T016	SKS	175.474	-41.189	-175.870	51.451	2010/09/03	11:16:06.000	23.000	6.5	5.410
T016	SKS	175.474	-41.189	-178.649	51.423	2010/08/04	12:58:24.000	27.000	6.4	3.679
T016	SKS	175.474	-41.189	93.471	11.132	2010/05/31	19:51:45.000	112.000	6.5	283.557
T022	SKS	175.350	-41.125	-178.649	51.423	2010/08/04	12:58:24.000	27.000	6.4	3.756
T022	SKS	175.350	-41.125	92.831	13.667	2010/03/30	16:54:46.000	34.000	6.7	285.114
T020	SKS	175.385	-41.146	-171.836	52.050	2011/06/24	03:09:39.000	52.000	7.3	7.865
T020	SKS	175.385	-41.146	-175.361	51.374	2010/10/08	03:26:13.000	19.000	6.4	5.790
T020	SKS	175.385	-41.146	-97.782	16.396	2010/06/30	07:22:27.000	20.000	6.3	75.569
T020	SKS	175.385	-41.146	-178.649	51.423	2010/08/04	12:58:24.000	27.000	6.4	3.734
T020	SKS	175.385	-41.146	92.831	13.667	2010/03/30	16:54:46.000	34.000	6.7	285.090
T004	SKS	175.669	-41.340	-169.848	52.876	2010/07/18	05:56:44.000	14.000	6.6	8.748
T004	SKS	175.669	-41.340	-97.782	16.396	2010/06/30	07:22:27.000	20.000	6.3	75.417
T004	SKS	175.669	-41.340	-175.870	51.451	2010/09/03	11:16:06.000	23.000	6.5	5.289
T004	SKS	175.669	-41.340	63.951	28.777	2011/01/18	20:23:23.000	68.000	7.2	280.209
T004	SKS	175.669	-41.340	96.548	33.165	2010/04/13	23:49:38.000	17.000	6.9	301.973
LE4	SKS	175.692	-41.358	-169.848	52.876	2010/07/18	05:56:44.000	14.000	6.6	8.735
WEL	SKS	174.768	-41.286	-171.836	52.050	2011/06/24	03:09:39.470	52.000	7.3	8.243
WEL	SKS	174.768	-41.286	-175.361	51.374	2010/10/08	03:26:13.710	19.000	6.4	6.175
WEL	SKS	174.768	-41.286	-169.848	52.876	2010/07/18	05:56:44.930	14.000	6.6	9.285
WEL	SKS	174.768	-41.286	-97.782	16.396	2010/06/30	07:22:27.690	20.000	6.3	75.977
WEL	SKS	174.768	-41.286	-171.708	52.171	2011/09/02	10:55:53.590	32.000	6.9	8.299
WEL	SKS	174.768	-41.286	-175.870	51.451	2010/09/03	11:16:06.600	23.000	6.5	5.849
WEL	SKS	174.768	-41.286	-178.649	51.423	2010/08/04	12:58:24.200	27.000	6.4	4.120
WEL	SKS	174.768	-41.286	-63.029	-28.398	2011/09/02	13:47:09.620	578.000	6.7	131.742
WEL	SKS	174.768	-41.286	99.822	20.687	2011/03/24	13:55:12.010	8.000	6.9	295.150
WEL	SKS	174.768	-41.286	171.161	53.127	2010/12/23	14:00:32.330	18.000	6.3	357.822
WEL	SKS	174.768	-41.286	92.831	13.667	2010/03/30	16:54:46.730	34.000	6.7	285.481
WEL	SKS	174.768	-41.286	93.471	11.132	2010/05/31	19:51:45.870	112.000	6.5	284.016
WEL	SKS	174.768	-41.286	-161.290	54.787	2011/07/16	19:59:12.890	36.000	6.1	13.726
WEL	SKS	174.768	-41.286	63.951	28.777	2011/01/18	20:23:23.480	68.000	7.2	280.755
WEL	SKS	174.768	-41.286	-72.571	18.443	2010/01/12	21:53:10.060	13.000	7.0	90.224
T007	SKS	175.651	-41.304	-68.328	-22.227	2010/03/04	22:39:26.000	114.000	6.4	123.477
T010	SKS	175.583	-41.252	-175.870	51.451	2010/09/03	11:16:06.000	23.000	6.5	5.342
T010	SKS	175.583	-41.252	93.471	11.132	2010/05/31	19:51:45.000	112.000	6.5	283.485
T014	SKS	175.506	-41.208	-97.782	16.396	2010/06/30	07:22:27.000	20.000	6.3	75.501
T014	SKS	175.506	-41.208	-175.870	51.451	2010/09/03	11:16:06.000	23.000	6.5	5.389
T014	SKS	175.506	-41.208	-109.159	24.690	2010/10/21	17:53:13.000	10.000	6.7	62.170
LE4	TelS	175.692	-41.358	123.519	6.218	2010/07/24	05:35:01.000	553.000	6.6	301.560
LE4	TelS	175.692	-41.358	122.974	4.195	2011/02/10	14:39:27.000	523.000	6.5	299.686
LE4	TelS	175.692	-41.358	123.039	4.077	2011/02/10	14:41:58.000	525.000	6.6	299.655
LE4	TelS	175.692	-41.358	116.720	-6.873	2011/03/10	17:08:36.000	510.000	6.5	286.213
LE4	TelS	175.692	-41.358	123.467	6.486	2010/07/23	22:51:12.000	585.000	7.6	301.703
T004	TelS	175.669	-41.340	123.519	6.218	2010/07/24	05:35:01.000	553.000	6.6	301.573
T004	TelS	175.669	-41.340	122.974	4.195	2011/02/10	14:39:27.000	523.000	6.5	299.698
T004	TelS	175.669	-41.340	116.720	-6.873	2011/03/10	17:08:36.000	510.000	6.5	286.222

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Table C.2 – Continued

station	phase	st.lon.°	st.lat.°	event lon.°	event lat.°	day	origin time (s)	depth (km)	mag.	BAZ°
T004	Tel.S	175.669	-41.340	139.869	26.978	2011/01/12	21:32:55.000	527.000	6.4	327.394
T016	Tel.S	175.474	-41.189	139.154	28.360	2010/11/30	03:24:41.000	486.000	6.8	327.599
T016	Tel.S	175.474	-41.189	123.519	6.218	2010/07/24	05:35:01.000	553.000	6.6	301.679
T016	Tel.S	175.474	-41.189	123.247	6.528	2010/07/29	07:31:56.000	627.000	6.6	301.671
T016	Tel.S	175.474	-41.189	122.974	4.195	2011/02/10	14:39:27.000	523.000	6.5	299.800
T016	Tel.S	175.474	-41.189	123.039	4.077	2011/02/10	14:41:58.000	525.000	6.6	299.770
T016	Tel.S	175.474	-41.189	116.720	-6.873	2011/03/10	17:08:36.000	510.000	6.5	286.296
T016	Tel.S	175.474	-41.189	139.869	26.978	2011/01/12	21:32:55.000	527.000	6.4	327.533
T016	Tel.S	175.474	-41.189	123.409	6.718	2010/07/23	22:08:11.000	607.000	7.3	301.936
T016	Tel.S	175.474	-41.189	123.467	6.486	2010/07/23	22:51:12.000	585.000	7.6	301.823
T020	Tel.S	175.385	-41.146	123.519	6.218	2010/07/24	05:35:01.000	553.000	6.6	301.737
T020	Tel.S	175.385	-41.146	123.247	6.528	2010/07/29	07:31:56.000	627.000	6.6	301.729
T020	Tel.S	175.385	-41.146	122.974	4.195	2011/02/10	14:39:27.000	523.000	6.5	299.857
T020	Tel.S	175.385	-41.146	123.039	4.077	2011/02/10	14:41:58.000	525.000	6.6	299.826
T020	Tel.S	175.385	-41.146	116.720	-6.873	2011/03/10	17:08:36.000	510.000	6.5	286.342
T020	Tel.S	175.385	-41.146	123.467	6.486	2010/07/23	22:51:12.000	585.000	7.6	301.880
T022	Tel.S	175.350	-41.125	123.519	6.218	2010/07/24	05:35:01.000	553.000	6.6	301.758
T022	Tel.S	175.350	-41.125	123.247	6.528	2010/07/29	07:31:56.000	627.000	6.6	301.750
T022	Tel.S	175.350	-41.125	123.467	6.486	2010/07/23	22:51:12.000	585.000	7.6	301.902
WEL	Tel.S	174.768	-41.286	139.154	28.360	2010/11/30	03:24:41.000	486.000	6.8	328.180
WEL	Tel.S	174.768	-41.286	123.519	6.218	2010/07/24	05:35:01.060	553.000	6.6	302.999
WEL	Tel.S	174.768	-41.286	123.039	4.077	2011/02/10	14:41:58.820	525.000	6.6	300.393
WEL	Tel.S	174.768	-41.286	116.720	-6.873	2011/03/10	17:08:36.860	510.000	6.5	286.890
WEL	Tel.S	174.768	-41.286	139.869	26.978	2011/01/12	21:32:55.210	527.000	6.4	328.128
WEL	Tel.S	174.768	-41.286	123.467	6.486	2010/07/23	22:51:12.450	585.000	7.6	302.441
LE4	ScS	175.692	-41.358	179.116	-23.307	2010/06/30	04:31:02.000	581.000	6.4	10.111
LE4	ScS	175.692	-41.358	179.760	-23.784	2011/07/29	07:42:22.000	523.000	6.7	12.239
LE4	ScS	175.692	-41.358	-178.425	-17.794	2009/11/22	07:48:20.000	522.000	6.4	13.880
LE4	ScS	175.692	-41.358	178.394	-26.142	2011/02/21	10:57:52.000	558.000	6.5	9.217
LE4	ScS	175.692	-41.358	-178.826	-20.799	2010/08/16	19:35:49.000	603.000	6.2	14.432
T004	ScS	175.669	-41.340	178.394	-26.142	2011/02/21	10:57:52.000	558.000	6.5	9.305
T004	ScS	175.669	-41.340	-178.826	-20.799	2010/08/16	19:35:49.000	603.000	6.2	14.503
T014	ScS	175.506	-41.208	179.116	-23.307	2010/06/30	04:31:02.000	581.000	6.4	10.735
T016	ScS	175.474	-41.189	179.116	-23.307	2010/06/30	04:31:02.000	581.000	6.4	10.841
T016	ScS	175.474	-41.189	178.394	-26.142	2011/02/21	10:57:52.000	558.000	6.5	10.057
T016	ScS	175.474	-41.189	-178.826	-20.799	2010/08/16	19:35:49.000	603.000	6.2	15.103
T020	ScS	175.385	-41.146	179.116	-23.307	2010/06/30	04:31:02.000	581.000	6.4	11.125
T020	ScS	175.385	-41.146	178.394	-26.142	2011/02/21	10:57:52.000	558.000	6.5	10.387
T020	ScS	175.385	-41.146	-178.826	-20.799	2010/08/16	19:35:49.000	603.000	6.2	15.360
T022	ScS	175.350	-41.125	123.519	-23.307	2010/06/30	04:31:02.160	581.000	6.4	12.829
WEL	ScS	174.768	-41.286	179.116	-23.307	2010/06/30	04:31:02.160	581.000	6.4	12.829
WEL	ScS	174.768	-41.286	178.394	-26.142	2011/02/21	10:57:52.410	558.000	6.5	12.361
WEL	ScS	174.768	-41.286	-178.585	-17.642	2011/04/03	14:07:09.330	551.000	6.4	15.591
WEL	ScS	174.768	-41.286	-21.611	-21.611	2011/09/15	19:31:04.080	644.000	7.3	15.536
WEL	ScS	174.768	-41.286	-178.826	-20.799	2010/08/16	19:35:49.000	603.000	6.2	16.829
WEL	ScS	174.768	-41.286	179.467	-31.569	2009/11/22	22:47:27.480	435.000	6.2	22.774

C.3 Null-measurement information

Table C.3 Information of the null-splitting[†] measurements.

station	event lat.°	event lon.°	phase	δt (s)	error δt (s)	ϕ°	error ϕ°	ϕ_p^*
LE4	-171.836	52.050	SKS	5.888	1.425	12.000	16.250	7.680
LE4	-63.029	-28.398	SKS	6.000	1.500	-51.000	45.250	131.119
LE4	99.822	20.687	SKS	5.700	0.591	34.000	3.250	294.555
LE4	-68.228	-21.701	SKS	5.438	1.378	-46.000	14.750	123.165
T004	-68.216	-22.146	SKS	5.662	1.453	-53.000	24.750	123.489
T004	-171.836	52.050	SKS	6.000	0.356	13.000	1.000	7.694
T004	-77.306	-1.266	SKS	5.925	1.500	15.000	45.250	102.341
T004	99.822	20.687	SKS	5.287	0.384	33.000	1.250	294.570
T004	-68.228	-21.701	SKS	6.000	1.500	-57.000	45.250	123.180
T007	92.831	13.667	SKS	3.487	1.031	18.000	1.000	284.909
T014	-169.848	52.876	SKS	6.400	0.200	4.000	22.500	8.844
T014	-178.649	51.423	SKS	6.400	0.370	-88.000	22.500	3.659
T014	92.831	13.667	SKS	3.337	0.947	23.000	23.250	285.008
T016	-175.361	51.374	SKS	3.300	1.481	-90.000	22.000	5.735
T016	-169.848	52.876	SKS	4.013	0.506	-90.000	1.750	8.863
T016	-77.306	-1.266	SKS	5.625	0.656	-70.000	2.250	102.444
T016	99.822	20.687	SKS	5.550	1.500	21.000	45.250	294.703
T016	-68.228	-21.701	SKS	3.413	1.500	30.000	45.250	123.305
T016	92.831	13.667	SKS	2.700	0.956	23.000	4.250	285.030
T020	-77.306	-1.266	SKS	2.288	1.500	-72.000	45.250	102.497
T020	-68.228	-21.701	SKS	2.850	0.366	25.000	1.500	123.363
T020	96.548	33.165	SKS	3.000	1.500	-55.000	45.250	302.172
T022	-175.361	51.374	SKS	4.950	1.500	5.000	45.250	5.812
T022	-169.848	52.876	SKS	6.000	0.684	-88.000	24.750	8.936
T022	-175.870	51.451	SKS	3.562	1.378	85.000	24.000	5.486
T022	-77.306	-1.266	SKS	6.000	0.525	90.000	24.750	102.517
T022	96.548	33.165	SKS	5.438	1.031	-69.000	24.750	302.196
WEL	-70.547	-18.058	SKS	6.000	0.938	21.000	24.500	119.715
WEL	-63.136	-26.803	SKS	3.487	0.947	33.000	2.500	130.652
WEL	-94.338	17.208	SKS	3.750	0.966	65.000	5.250	77.505
WEL	-71.558	-8.087	SKS	6.000	1.022	12.000	4.750	112.015
WEL	-68.228	-21.701	SKS	3.150	1.087	25.000	4.750	123.796
WEL	-65.089	-15.364	SKS	2.513	1.181	26.000	5.000	121.780
WEL	96.548	33.165	SKS	2.250	1.462	-68.000	24.750	302.489
T004	139.154	28.360	Tel.	1.975	0.863	-73.000	18.500	32.058
T004	139.869	26.978	Tel.	2.725	0.562	37.000	4.750	204.416
T020	139.869	26.978	Tel.	0.900	0.594	-75.000	16.750	30.494
T020	123.409	6.718	Tel.	2.900	1.000	45.000	45.250	139.682
WEL	122.974	4.195	Tel.	3.225	0.875	40.000	16.500	142.929
WEL	123.409	6.718	Tel.	4.000	1.000	34.000	45.250	125.664
LE4	-179.528	-21.611	ScS	2.475	0.531	-40.000	29.250	4.660
T004	-178.585	-17.642	ScS	1.925	0.356	14.000	22.750	111.951

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Table C.3 – Continued

station	event	event	phase	δt (s)	error	ϕ°	error ϕ°	ϕ_p^*
	lat. $^\circ$	lon. $^\circ$			δt (s)			
T016	179.760	-23.784	ScS	2.800	0.219	-63.000	22.000	-49.668
T020	-178.585	-17.642	ScS	2.925	0.125	90.000	2.750	258.106
WEL	179.760	-23.784	ScS	1.375	0.950	16.000	16.250	96.065
WEL	-178.425	-17.794	ScS	1.575	0.281	43.000	4.250	143.270

*initial polarization direction. In the case of SKS, ϕ_p =BAZ

†measurements are considered to be null if the $|\phi - \phi_p| \leq 12^\circ$ or $|\phi + 90^\circ - \phi_p| \geq 80^\circ$

C.4 Multilayer analysis

We have detected lateral variations in splitting parameters (Fig. 6.8) along the SAHKE profile and delay times from S phases also exhibit frequency dependency throughout the southern Hikurangi subduction zone (Fig. 6.5 and *Marson-Pidgeon and Savage* [1997]). These observations suggest that lateral and/or depth varying anisotropic structure is plausible in the region. Since the frequency dependent splitting is obvious throughout the region, it is clear that this is not caused by a localized source of anisotropy. Anisotropy might, therefore, be characterized by a multilayer structure. To explain the observed lateral variations in anisotropy, this multilayer structure must contain layers with laterally varying anisotropy. To explain the frequency dependent splitting, this multilayer structure should accommodate anisotropy with varying length scales. It is possible these different length-scales of anisotropy are either localized in a single layer or distributed in two or more layers, mixing with different length scales of anisotropy.

Initial polarization (ϕ_p) versus splitting parameter analysis is used to determine the best-fitting 2-layer and 3-layer (section 3.5.1) models for the observed anisotropy in the study region. We use 2-layer and N-layer analysis codes provided by *Wustefeld et al.* [2008] and *Walker and Wookey* [2012] for this analysis. Both these codes use the *Silver and Savage* [1994] method to determine multilayer anisotropy. Our aim is to find whether the observed frequency dependent splitting and/or lateral variations are due to depth varying multilayer structure. Because of the lateral variations in splitting parameters, this modeling approach is applied to three distinct regions (regions E, C and S; Fig. C.1) that have been subdivided based on the lateral variations in splitting parameters and the spatial distribution of stations. Each model starts with a set of initially defined multi layer structures. We used the assumed anisotropic contributions in the region (e.g. absolute, and relative plate and upper-plate fault orientations; Fig. 6.2) to define anisotropic layers in our initial

Figure C.1 Map showing the spatial division of stations that two and three layer analysis. Stations and models in green, brown and yellow zones are named as region **E**, region **C** and **S** in the text respectively.

structures. In each layer, δt and ϕ are set to vary within a range of values (table C.4) and a best-fitting model is determined for that particular structure. In two layer case the best fitting model is found by minimizing the total normalized sum-squared residuals of ϕ and δt . The total sum-squared residuals of ϕ and δt are normalized by the maximum total sum-squared of ϕ and δt , respectively. Then the best fit 2-layer model is selected by finding the model with the minimum total (ϕ , δt) normalized sum-squared residuals (table C.5). Next we visually compared the best-fitting models obtained from different initially predicted structures to find the most appropriate model to explain the observed splitting parameters. Because of the short (5 s; ScS) and long period (11 s; SKS and Tel. S) S-phases Fig. 6.5) used in this study, some of the models are tested for different frequencies. However, due to the limited observed initial polarization directions for the short period phases (ScS), it is hard to explain whether relatively smaller delays and frequency dependent splitting from ScS phases are caused by a spatial or a depth effect. At least three possible

Table C.4 δt and ϕ search ranges in each layer of the different 2-layer models.

2-layer search ranges				
model	δt -top(s)	δt -bot.(s)	ϕ -top($^{\circ}$)	ϕ -bot.($^{\circ}$)
C1	1-2.5	0.8-1.2	70-90	-10-(-80)
C2	0.8-1.2	1-2.5	-10-(-80)	70-90
C3	0.7-1	1-1.5	10-40	70-85
E1	1-2.5	0.8-1.2	50-90	-10-(-80)
E2	0.7-1	1-1.5	10-40	70-85
S1	1-2.5	0.8-1.2	70-90	-10-(-80)
S3	0.7-1	1-1.5	10-40	70-85

sources could control the extent and alignment of anisotropic structure in the region (i.e. faulting in the upper plate, relative and absolute plate motions). Therefore, we have extended our two-layer analysis to a three-layer case. Best-fitting two-layer models are modified by adding another layer and by changing ϕ and δt in a small range around the best-fit values of the resolved 2-layer model. In the following sections, we discuss the best resolved two and selected 3-layer models in each of the regions separately.

C.4.1 Layered anisotropic structure in region C

This region lies in the Western end of the SAHKE transect (region C-brown zone in Fig. C.1). Splitting parameters from Tel. S and SKS (long period) phases in this region show a higher variability ($\sim 1-3.4$ s) and relatively large δt (3.4 ± 0.43 s) (Fig. 6.8) compared to eastern side ($\sim 2.2 \pm 0.3$ s) (region E-brown; green zone in Fig. C.1). Figure C.4 shows the best fitting models resolved from three different multi layer structures (note the initial structural configuration of the 2-layers displays at the top of the each model).

The model C_1 appears to fit reasonably well with the observed data, suggesting the possibility of 2-layer anisotropic structure that is characterized by lower APM

controlled anisotropy and upper RPM controlled anisotropy. The top layer of C_1 has high δt (2.5 s). This implies either higher strength of anisotropy or more unmodeled sub-layers of anisotropy in the upper layer, so that cumulative anisotropy can be as high as 2.5 s. The mantle-wedge contribution must be negligible as the subduction is very shallow in this region. But, subducting slab-lithosphere (~ 60 km thick),

Figure C.2 Best fitting 2-layer models obtained for three different structures in the central part of southern NI (region **C**). The best-fitting structural configuration with 2-layers displays top of the each model. Red and magenta symbols indicate the measured SKS and Tel. Splitting parameters respectively. The best fitting 2-layer model is marked in dashed-dotted line.

Figure C.3 Selected 3-layer models obtained for the extended versions of the best fitting two-layer structures in the central part of southern NI (region **C**). The resolved structural configuration with 3-layers is shown top of the each model. Red, magenta and blue symbols indicate the measured SKS, Tel. S and ScS splitting parameters respectively. The best resolved 3-layer model is marked by the dotted line. Correct mark denotes the selected model that agree with the majority of the data points.

Table C.5 Misfits[†] of the best-fit 2-layer and selected 3-layer models.

region	2-layer normalized misfit			3-layer misfit of δt & ϕ		
	1	2	3	x	y	z
C	0.21	0.21	1.04	4.77	5.65	4.73
				246.35	265.32	242.36
E	0.11	0.69	-	7.41	9.31	4.77
				227.60	268.35	164.59
S	0.10	-	0.44	3.23	4.78	2.88
				83.58	146.08	143.48

[†] minimum misfits was estimated from the normalized sum-squared residuals of ϕ and δt in 2-layer case. In 3-layer case, the misfits of ϕ and δt are determined separately by calculating the square root of sum-squared residuals of ϕ and δt . In 3-layer case we did not determine the best-fit, thus ϕ and δt are not normalized.

including its sub-slab region and upper-plate can act as two separate anisotropic layers (sub-layers) within the upper-layer of the well resolved 2-layer model. 2.5 s δt can be produced by these two sub-layers, if the anisotropy in the layers are coupled. However, small-scale variations in the alignment of anisotropy in each layer are possible. Based on this possibility, we have tested another 2-layer case that is characterized by upper moderately high anisotropic zone (maximum up to 1 s) and lower high anisotropic zone (maximum up to 1.5 s). Anisotropic orientations are allowed to vary around RPM direction (table C.4) in the lower layer and fault parallel direction in the upper layer (e.g. best fitting model C_3 in Fig. C.2). Model C_3 shows the best-fit found for the above approximation and this also fit fairly well with the observations. Thus, we extended out 2-layer case to 3-layer case combining the best fit models of C_1 and C_3 .

The initial 3-layer models are defined so that the top two layers are comparable with the resolved anisotropic layers by the best-fitting two layer models (Fig. C.2), but a third layer ($\phi = -20^\circ$, $\delta t = 0.8$ s) is added to represent potential APM-derived anisotropy (Fig. C.3). Figure C.3 (bottom) shows the 3-layer models (C_x , C_y , and C_z) that are obtained from extended two-layer structures. We choose C_z as the best-resolved 3-layer model that agrees well with our splitting data set. C_x also

shows a good fit to data, but we think that the upper δt of 0.9 s is too high to occur within the upper-most layer. The selected 3-layer model (C_z) (Table C.5), consists of layers that are approximately aligned (from deep to shallow) with the APM, RPM and general trend of the upper-plate faulting from bottom to top respectively. The δt value of 2.2 s is constrained in the middle layer. Anisotropy responsible for such high δt could be caused by a the thick anisotropic layer and/or a source with high strength of anisotropy. This model has a top most layer anisotropy of 0.7 s, which is considerably higher than the typical crustal anisotropy (~ 0.2 s). However, high delays in the top two layers can be explained by anisotropic structure associated with the high strength of anisotropy (e.g. fault structure with intense foliations and a broad shear-zone). This structure could be vertically localized in the upper most two layers, coupled with the background layer anisotropy (i.e. typical anisotropy of the layer). Therefore, high delays in the topmost two layers can be explained by this argument. Since this region underlies the large-scale dextral fault (Wairarapa fault), it is likely characterized by intense foliations and a broad shear-zone. We think that the uppermost of the three layers represents crust with high structural anisotropy (e.g. upper-plate crust with fault structures). The middle layer could have anisotropic contributions from the lower crustal, upper-mantle (sub-slab) and subducting-slab lithosphere, but this anisotropy could be enhanced by the vertically extending fault shear zone in order to produce 2.2 s δt . Depth extent of this fault zone is hard to constrain accurately, but when compared with lower δt (0.8 s) and APM parallel ϕ from the bottom most layer, we argue that the shear zone below the fault does not extend beyond the sub-slab region. It is possible that this fault shear zone penetrates into the subducting-slab lithosphere, and possibly down to sub-slab region. About 2.8 - 3 s large δt observed from few ray paths close to the Wairarapa fault may indicate the cumulative affect of anisotropy of top-most two layers, suggesting anisotropy due to foliations and shear zone and background anisotropy of the layer are vertically coupled. If we test the same layered structures for region E (Fig. C.1); we will find a good fit to data from a 3 layer structure, but

with lower split δt for the topmost two layers. This may be due to lack of large-scale faulting in the eastern part (see section C.4.2).

C.4.2 Layered anisotropic structure in region E

Region E (green zone in Fig. C.1) includes the eastern part of the SAHKE profile. This region does not contain large-scale fault traces on the surface, but a few splay-faults of the Wairarapa fault are predicted to be exposed in this region [Henry et al., 2013]. We used the same approach we applied to region C to find the multilayer structure that could explain the observed splitting parameters. We have found two best fitting two layer models (Table C.5) with different layer approximations and they match fairly well with the observed splitting parameters. However, we prefer model E_2 as it fits well with the observed δt (Figs. C.4). Since there are two different best fitting models (Table C.5) that explain two different layered structures, we searched for the 3-layer case by introducing a third layer to both best-fitting two layer models and considering the possible depth distribution of anisotropic layers in the region. We have selected E_z as the well resolved 3-layer model that explains the observations. This model also agrees with the 3-layer model resolved for the region C_z . It has the uppermost layer δt anisotropy of 0.5 s that is aligned about 12° NE, middle layer with 1.6 s with RPM sub-parallel (71°) anisotropy, and the lower layer of 0.8 s delay with approximately APM parallel anisotropy (-20° NW).

When comparing the best constrained 3-layer models for regions C and E along the SAHKE profile, there is a clear discrepancy between δt measured in the top two layers of the C_z (top=0.7 s, middle=2.2 s) and E_z (top=0.5 s, middle=1.6 s). Both areas have the lowermost layer anisotropy of 0.8 s roughly parallel to APM direction (20°). Although the general trend of ϕ in these two layers are NE-SW aligned, ϕ in C_z is more NE-SW aligned (45°) compared to NNE-SSW (12°) alignment found for region E. Higher δt in the top two layers in C_z compared to that in E_z is evidence

Figure C.4 Best fitting 2-layer models obtained for two different structures in eastern part of southern NI (region **E**). The best-fitting structural configuration with 2-layers displays top of the each model. Red and magenta symbols indicate the measured SKS and Tel. Splitting parameters respectively. The best fitting two layer model is marked in dash-dot line.

Figure C.5 Well resolved 3-layer models obtained for the extended versions of the best fitting two-layer structures in the eastern part of southern NI (region **C**). The resolved structural configuration with 3-layers displays top of the each model. Red, magenta and blue symbols indicate the measured SKS, Tel. S and ScS splitting parameters respectively. The best fitting two layer model is marked by the dotted line. Correct mark denotes the selected model that agree with the majority of the data points.

for a predicted vertically extending high anisotropic structure, cross-cutting at least the topmost two layers. We think that this structure likely represents the Wairarapa fault and its associated shear-zone. Both crustal and sub-crustal layer anisotropy in region C appears to be enhanced by this structure. The lateral variation in δt along the profile (Fig. 6.9) implies that the spatial extent of the shear zone at depth may be asymmetric and distributed in the Wairarapa basin area, but may not extend as

far as region E from the fault.

C.4.3 Layered anisotropic structure in region S

We used splitting measurements obtained from WEL station, which is located in region S, (Fig. C.1) to determine a multilayer structure in the southwest of the SHAKE profile. This region is also located around a large dextral strike-slip fault zone (Wellington fault). Anisotropy in this region was previously approximated with a 2-layer structure by using a different set of stations [Gledhill and Gubbins, 1996, Marson-Pidgeon and Savage, 2004a]. Dipping-anisotropy layer study by Marson-Pidgeon and Savage [2004a] suggested a complex anisotropic structure in the region. Here, we analyze a new splitting data set with a better backazimuthal coverage that is measured on the GeoNet permanent station (WEL).

Figure C.6 3-layer and 2-layer models determined for the southwest region of the SHAKE profile (region S). The figure symbols are same as in figures C.2 and C.3.

Both best fitting 2-layer (S_1) and expanded 3-layer structure (S_x) that are determined from the splitting measurements at WEL are shown in figure left (S_1) and right (S_x) respectively. Both models do not fit-well with the observations (Table C.5), suggesting a more complex anisotropic structure that could not be explained by a simple 2- or 3- layer model.

C.4.4 Overall structure

The anisotropic structure of the southern North Island is likely stratified, but in complex manner. It is hard to distinguish between the best-fit models from different structures that are initially proposed for each region. Multilayer models suggest that the anisotropy of the individual layers can be enhanced by localized anisotropic zones that are possibly at shallow depths around faults, creating a more complex anisotropic structure and producing the observed lateral variation. Thus, this laterally varying anisotropy should be investigated using a different approach. These multilayer modeling results provide a first-order approximation on the possible layered anisotropic structure, however they do not fully explain the observations.

C.5 Combined analysis of short and long period frequency-dependent splitting

Figure C.7 shows the frequency-dependent estimations in the Hikurangi subduction zone using both local and teleseismic δt . The extrapolated least-squares fits of both local and teleseismic measurements fit well with the observed measurements (Fig. C.7), suggesting S-phase with different periods will be sensitive to different anisotropic length scales and, thus, give a range of δt .

Figure C.7 An integration of frequency-dependent δt from local splitting (blue circles) and teleseismic (ScS and Tel. S) splitting (black circles) in the Hikurangi subduction zone. Blue circles denotes the frequency-dependent local δt that are determined at the northern SI stations (NSI in Fig. 4.9). Black circles represents frequency-dependent splitting observed in the southern SI (Fig. 6.5). Black and blue dashed-lines indicate the least-squares fit of the teleseismic and local data, respectively. Two pink squares denote the predicted local δt measurements at periods 1 s and 0.5 s based on the teleseismic least-squares fit.

Predicted teleseismic δt from the least-squares fit of local δt versus period plots in figure Fig. 4.9 have been compared with the in-situ teleseismic δt in SI. The average period of 6 s is assumed in the predictions. For large periods, local models seem to over-predict teleseismic δt (Fig. 4.9).

According to *Rumpker and Silver* [1998], the measurements that lie in the region

Figure C.8 Same as figure caption in Fig. C.7, but the frequency dependence is compared with the modeling approximations (red-bold and red-dashed lines) by *Rumpker and Silver* [1998]. The red-bold line represents: $\delta t = 0.2T$, the red-dashed line indicates: $\delta t = T$ and the magenta-dashed line denotes: $\delta t = 0.1T$.

$T/\delta t \lesssim 1$ (i.e. region above the red-dashed line) are biased towards the upper-most layer anisotropy. Only high frequency measurements (blue circles) agree with this model. The measurements with $T/\delta t \gtrsim 5$ in figure C.8 (i.e. region below the red-bold line) likely represent the apparent splitting parameters from an underlying multi-layered anisotropy. The majority of estimated δt from both high-frequency and long-period S-waves splitting are less than 1/10 of a corresponding S-wave period (i.e. region above the magenta-dashed line).

Table C.6 Comparison between predicted SKS δt_{tel} at 6 s period from the local least-squares fits in figure 4.9 and measured δt_{tel} values from previous SKS studies in approximately the same locations in SI [*Savage et al.*, 2007a, *Duclos et al.*, 2005].

station	predicted δt_{tel} (s)	in-situ δt_{tel} (s)
Northern South Island		
LTZ	1.392 ± 1.328	1.76 ± 0.72
KELY	2.461 ± 1.527	1.4 ± 0.20
NNZ	2.115 ± 1.530	-
Average NSI ^a	1.445 ± 0.088	-
Southern South Island		
APZ	4.188 ± 2.026	-
WHZ	2.937 ± 0.859	2.95 ± 0.55
MSZ	3.702 ± 1.421	-
Average SSI ^b	2.834 ± 0.540	-
Average South Island	2.1395 ± 0.314	1.6 ± 0.2

^aNorthern South Island

^bSouthern South Island

Appendix D

Data-processing and archiving information

- Data from the temporary deployments were archived using PASSCAL-PASSOFT and Antelope softwares. The basic steps involved with this process are documented and located in the archived folder[‡] on the Geophysics network, and these have also been provided to the Geophysics data unit. This information and data can be obtained on request.
- Sac-file records of the OBS and SAHKE data are placed in the archived folder[‡].
- An integrated method was implemented to merge hour-long miniseed files, to generate continuous sac records from the Earth-Data, and to extract events depending on the origin-time and phase arrivals. This method used PASSCAL(mseed2sac, mseedmod), SEISAN and SAC softwares. Processing steps and scripts related to this raw-data conversion can be found in the archived folder[‡] on the Geophysics network and these documents can be obtained on request.

“Experience is the universal mother of sciences.”

-Miguel de Cervantes

[‡]path: /Volumes/GeoPhysics_07/users-data/karallsa

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