VICTORIA UNIVERSITY OF WELLINGTON

MASTERS THESIS

Seismic Anisotropy at the Hikurangi Subduction Margin

Author:

Thomas Wilson

Supervisor:

Martha Savage

*A thesis submitted in partial fulfilment of the requirements
for the degree of Master of Science*

in the

School of Geophysics

School of Geography, Environment and Earth Sciences

August 2016

Declaration of Authorship

I, Thomas Daniel Wilson, declare that this thesis titled 'Seismic anisotropy at the Hikurangi Subduction Margin', and the work presented in it are my own. I confirm that:

- This work was done wholly or mainly while in candidature for a research degree at this University.
- Where any part of this thesis has previously been submitted for a degree or any other qualification at this University or any other institution, this has been clearly stated.
- Where I have consulted the published work of others, this is always clearly attributed.
- Where I have quoted from the work of others, the source is always given. With the exception of such quotations, this thesis is entirely my own work.
- I have acknowledged all main sources of help.
- Where the thesis is based on work done by myself jointly with others, I have made clear exactly what was done by others and what I have contributed myself.

Signed:

Date:

Abstract

To determine the stress state in the southern North Island of New Zealand, we use shear wave splitting analysis to measure seismic anisotropy and infer the orientation of the maximum horizontal stress directions (S_{hmax}) in the crust. We use data recorded by 44 temporary seismometers deployed as part of the Seismic Array Hikurangi Experiment, and from six permanent stations from the national GeoNet network. Using 425 local earthquake events recorded across the 50 stations we made 13,807 measurements of the two splitting parameters, ϕ (fast direction) and δt (delay time). These measurements are compared to S_{Hmax} directions obtained from previous focal mechanism studies (S_{Hmax}^{focal}), and stresses due to the weight of topography (S_{Hmax}^{grav}). Generally there is good agreement between the alignment of S_{Hmax}^{focal} , S_{Hmax}^{grav} , and the mean ϕ measured at each station. We also find a $\sim 90^\circ$ change in the trend of ϕ in the Wairarapa region for stations across the Wairarapa Fault trace. Based on the variation of ϕ , we divide the study region into three regions (West, Basin, and East), whose bounds approximately coincide with the Wairarapa and Dry Creek faults. We find the average ϕ of the West region average agrees with previous anisotropy studies, which were undertaken within the bounds of the West region on the Tararua array. Also, we use our delay time measurements to estimate a $3.7 \pm 1.2\%$ strength of anisotropy in the overriding Australian Plate, which agrees with the 4% crustal anisotropy measured previously. There is close alignment of the region average ϕ of the West and East regions, which also agrees with the deep splitting measurements previously obtained. There is no significant difference between the mean ϕ and S_{hmax}^{grav} for the West and Basin regions; however, we find a difference of $31 \pm 19.5^\circ$ for the East region. We argue that this difference is due to tectonic loading stresses being sufficiently large in the East region to cause the total stress field to deviate from the gravitational stress field.

Acknowledgements

Firstly I wish to acknowledge my project supervisor Martha Savage for her excellent support and patience with me through out this project. I have found Martha's dedication to her students and passion for her work very inspiring. A massive thank you to Resource Development Consultants Limited (RDCL) for financially supporting me through my Master's degree. Without their help I may have never returned to university to do this research. I am extraordinarily grateful for the opportunity that RDCL have provided me. Thank you to: Adrian Shelley for his help verifying seismometer orientations; Hamish Hirschberg for supplying me with a model of the gravitational stress field model for my study region; Simon Lamb for the thought provoking discussions and suggestions with the interpretation of my results; Euan Smith for help with statistical analysis of my data set; John Townend for providing focal mechanism stress inversion data; Holly Godfrey for her wonderful assistance with proof reading my thesis, and great advice on making pretty figures; Aleksandr Beliaev and Adrian Benson for IT and technical support, particularly in instances when I have somehow managed to get my computer in a position where I need to reinstall my operating system! I would also like to thank my office mates Gabe, Cam and Steve for providing great office banter and making the it a great environment to work in; and my awesome flatmates: Billy, Elena, Jenö and Mattithas for keeping me partially sane. Lastly I would like to thank my family: Robin, Ginge, Paddy; and my partner Lara, for their great support through out my masters degree and helping me push through the challenges I faced throughout the journey.

Contents

Declaration of Authorship	ii
Abstract	iii
Acknowledgements	iv
Contents	v
List of Figures	ix
List of Tables	xi
Abbreviations	xiii
Symbols	xv
1 Introduction	1
1.1 Motivation	1
1.2 Seismic waves and anisotropy	2
1.2.1 Seismic waves	2
1.2.2 Seismic anisotropy and shear wave splitting	3
1.2.3 Origin of seismic anisotropy	4
1.3 Tectonic setting	7
1.3.1 New Zealand tectonic setting	7
1.3.2 Southern North Island tectonic setting	8
1.3.3 Strain and locking	9
1.3.4 Stresses at the Hikurangi Margin	12
1.4 Isotropic crustal structure	13
1.4.1 Crustal heterogeneity	13
1.4.2 Seismic Array Hikurangi Experiment	15
1.5 Anisotropic structure at the southern Hikurangi Margin	18
2 Data and Methods	21
2.1 Earthquake data	21
2.2 P-wave arrivals	24

2.2.1	APK Phase picking algorithm	25
2.2.2	P wave pick quality	26
2.3	Picking S wave arrivals	27
2.3.1	Converted phases	28
2.3.2	Polarization filter	30
2.3.3	Autoregression method (AR-AIC)	31
2.3.4	S-wave quality	32
2.4	Multiple Filter Automatic Splitting Technique (MFAST)	33
2.4.1	Determining and applying the best filter	34
2.4.2	Splitting correction	35
2.4.3	Cluster analysis	35
2.4.4	Grading	37
2.4.5	Null measurements	37
2.4.6	Cycle skipping	37
2.4.7	Converted phases	38
2.4.8	Incidence angles	38
2.5	Delay time tomography	39
2.5.1	Regional gridding	39
2.5.2	Delay time inversion	40
2.5.3	Checker board test	42
2.5.4	Spatial averaging of fast direction	42
2.5.5	Amendments to the TESSA methodology	42
2.6	Gravitational stress modelling	44
3	Results and Analysis	47
3.1	Shear wave splitting results	47
3.2	Splitting by region	53
3.3	Splitting populations	54
3.4	Dependence of delay time on ray path and frequency	55
3.5	Splitting by depth	56
3.6	Gravitational stress due to topography	59
3.7	Region Average delay times	60
3.8	Topographic stresses versus mean fast directions	61
3.9	Delay time tomography	64
4	Discussion	71
4.1	Origin of anisotropy	71
4.1.1	Comparison to deep splitting measurements	73
4.1.2	Delay time tomography	75
4.2	Strength of anisotropy	77
4.3	Strain loading	78
4.4	Stress analysis	81
4.4.1	Stresses due to slab flexure and topography	82
4.4.2	Tectonic loading stress estimation	83

4.4.3	Expected vs. observed stress deviation	86
4.5	Variation in stress loading	87
4.5.1	Plate coupling	87
4.5.2	Shear stress concentration in the over-riding block	88
4.5.3	Release of stored elastic energy	89
4.5.4	Confidence in the deviation estimates	90
4.5.5	Additional stress model limitations	91
5	Conclusions	93
5.1	Summary	93
5.2	Future work and recommendations	95
A	Appendix A	97
A.1	Seismometer orientation	97
A.1.1	Misalignment calculation	98
A.1.2	Rotapy and Protate	103
A.1.3	Stachnik method	107
A.1.4	Ambient noise cross-correlation analysis	109
	Bibliography	111

List of Figures

1.1	P and S wave propagation	3
1.2	Anisotropy due to stress aligned micro cracks	7
1.3	New Zealand tectonic setting and study region features	9
1.4	Interplate locking	11
1.5	Stress contour map for the North Island of New Zealand	12
1.6	Tomographic grid	14
1.7	Cross sectional Vp structure	14
1.8	Cross sectional Vp/Vs structure	15
1.9	Structure of the subducting slab	17
1.10	Tararua Array station locations	19
2.1	GeoNet earthquake locations	22
2.2	SAHKE and GeoNet station locations within the study area	23
2.3	Seismogram showing APK and TauP picks	25
2.4	Seismogram showing APK and TauP picks	29
2.5	Rotation of the seismogram into the ray coordinate system	31
2.6	Histogram of MFAST bandpass filters	34
2.7	Example MFAST shear wave splitting measurement	36
3.1	Regional fault map	48
3.2	Study subregions map	49
3.3	Splitting map for earthquakes of all depths	53
3.4	Example: Station S022 splitting parameter heatmaps	54
3.5	Station S022 splitting parameter scatter plots	56
3.6	Map of earthquakes locations and depths in relation to the subducting slab	57
3.7	Map of splitting from earthquakes above and below 30km	58
3.8	Nearest neighbour S_{Hmax} due to topographic loading	59
3.9	Splitting and horizontal stress comparison	61
3.10	Comparison of ϕ and S_{Hmax}^{grav} by region	62
3.11	TESSA ray plot and spatially averaged ϕ	65
3.12	Input model for TESSA checker board test	66
3.13	Output model for TESSA checker board test	66
3.14	Delay time tomography resolution map	67
3.15	Delay time tomography map	68
3.16	Spatially averaged ϕ measurements on a grid	69

4.1	Strain across the southern North Island	80
4.2	Schematic figure of expected stress deviation	86
4.3	Schematic of the strain distribution in the overriding plate	89
A.1	Misalignment calculation schematic	98
A.2	Obtaining stations misalignment	100
A.3	Rotation of the seismogram into the ray coordinate system	104
A.4	Scatter plot of the correction angle vs the earthquake back azimuth for measurements used to orient the station BSWZ	106
A.5	Re-oriented seismometer comparison	107

List of Tables

2.1	Index table of seismometer numbering	23
2.2	SpickerC thresholds for determining the quality grade (class) of a S-pick made by SpickerC (default values [Castellazzi, 2014]. Column 1: S-pick classification; Column 2: Maximum error of the final S pick; Column 3: Maximum pick error for a epicentre to receiver distance less than 100 km; Column 4: Maximum pick error for a epicentre to receiver distance less than 100 km.	28
2.3	Input SpickerC parameters	29
2.4	S-wave checking table	33
3.1	West region splitting measurements	50
3.2	Basin region splitting measurements	51
3.3	East region splitting measurements	52
3.4	Orientation of the maximum horizontal compressive stress due to gravitational loading at each station	60
3.5	Average delay time for each region	61
3.6	Table of region averaged fast directions and S_{Hmax}^{grav}	63
3.7	Table of statistics for the difference of mean $\mu(\phi)$ and $\mu(g)$ for each region	63
4.1	S-phase periods from deep splitting studies [Karalliyadda, 2014] . .	74
A.1	Station locations and correction values for stations in the East region (figure 3.2)	101
A.2	Station locations and correction values for stations in the Basin region (figure 3.2)	102
A.3	Station locations and correction values for stations in the West region (figure 3.2)	102
A.4	Results from the rotapy program for three GeoNet test stations . .	106

Abbreviations

BAZ	Source to receiver B ack AZ imuth
B.P.	B efore P resent
GPE	G ravitational P ontential E nergy
MFAST	M ultiple F ilter A utomatic shear-wave S plitting T echnique
LTA	L ong T erm A verage
SAC	S eismic A nalysis C ode
SAHKE	S eismic A rray HiK urangi E xperiment
ScS	S hear-wave → c ore reflected → S hear-wave
SKS	S hear-wave → through mantle (K) → S hear-wave
SNR	S ignal to N oise R atio
STA	S hort T erm A verage
Tel. S	T eleseismic S hear-wave
TESSA	T omography E stimation of shear-wave S plitting and S patial A veraging

Symbols

V_p	Compressional wave (P wave) speed	$km s^{-1}$
V_s	Shear wave (S wave)	$km s^{-1}$
V_p/V_s	ratio of V_p to V_s	dimensionless
ϕ	Fast direction azimuth	degrees
δt	delay time	s
$spol$	Initial S wave polarization	degrees
k	strength of anisotropy	dimensionless
M_w	Earthquake moment magnitude	dimensionless
M_L	Earthquake local magnitude	dimensionless
S1	Maximum principle stress direction	degrees
S2	principle stress direction	degrees
S_{Hmax}	Maximum horizontal stress direction	degrees
S_{Hmax}^{focal}	S_{Hmax} computed from focal mechanism inversion	degrees
S_{Hmax}^{grav}	S_{Hmax} due to gravitational stress loading	degrees

Chapter 1

Introduction

1.1 Motivation

Beneath the Wellington region, the Pacific plate obliquely subducts at a low angle beneath the Australian plate. Geodetic studies indicate that beneath the southern end of the North Island of New Zealand there is strong locking on the interface between the subducting and overriding plates, causing a strain deficit and associated stress accumulation in the overriding crust [Wallace et al., 2004]. Because the material and the associated faults in the crust have a finite strength, once the stress accumulation exceeds the strength of the faults, rupture will occur and the accumulated strain will be relieved [Stein and Wyssession, 1991]. The sudden release of accumulated stress results in an earthquake where seismic waves propagate from the region of failure. At subduction zones the failure of large areas of locked plate interface results in the release of large amounts of seismic energy and can devastate nearby towns and cities in the proximity of the epicentre, and generate tsunamis (eg 2004 Sumatran earthquake [Lay et al., 2005]). The Earth's largest earthquakes are produced by subduction zones, e.g. the $M_w = 9.1$ 1964 Alaskan and $M_w = 9.5$ 1960 Chile earthquakes [Stein and Wyssession, 1991]; and the more recent the $M_w = 9.0$ 2011 Tohoku earthquake.

One of the primary controls on crustal deformation and faulting processes is the regional stress field (eg [Kusznir et al., 1991], [Zoback, 1992], [Townend et al., 2012]). Therefore, to understand the seismogenic potential in the southern North Island we need to understand the stress field. In the upper 15 km of the crust the maximum horizontal compressive stresses (S_{Hmax}) orientation can be approximated by the direction of maximum shear-wave speed through the crust (ϕ), which can be obtained using shear-wave splitting analysis [Crampin, 1994]. The aim of this project is to use measurements of seismic anisotropy to improve the understanding of the stress state at the southern end of the Hikurangi Margin.

1.2 Seismic waves and anisotropy

1.2.1 Seismic waves

There are two types of elastic body waves that propagate through a continuous medium. The first is a compressional wave, commonly known as the P-wave. The P-wave is the fastest travelling phase from an energy source, and has associated particle motion parallel to the propagation direction (figure 1.1). The second type of body wave is the shear-wave, commonly referred to as S-wave. S-waves are transverse waves, meaning that the particle motion associated with the wave is perpendicular to the propagation direction (figure 1.1).

FIGURE 1.1: P wave motion (red vector) is parallel to the propagation vector. The motion associated with an S-wave (blue) is perpendicular to the propagation direction in the SV (vertical), SH (Horizontal) plane.

1.2.2 Seismic anisotropy and shear wave splitting

Seismic anisotropy is the dependence of wave speed on the orientation that a wave travels through a medium [Babuska and Cara, 1991]. Because of their transverse nature, shear-waves can be decomposed into two orthogonal polarisations (SV and SH, see figure 1.1). In an isotropic medium, where the wave propagation speed is the same in all directions, the two orthogonal polarisations propagate at the same speed. However, when a polarised shear-wave enters an anisotropic medium, the orthogonal components (SV and SH) can propagate at different speeds, resulting in a time difference between the arrival of the phases on a seismograph. This phenomenon is known as seismic birefringence, or more commonly as shear wave splitting [Savage, 1999]. By measuring the shear-wave splitting exhibited by

earthquake waveforms, seismologists can measure the anisotropy sampled by an earthquake ray, and infer details of the structure and deformation in the earth.

A splitting measurement is characterised by three splitting parameters. ϕ , the azimuth shear waves travel through the medium fastest; dt , the delay time between the fast and slow incoming polarizations; and the initial polarization of the incoming wave, $spol$. The method in which measurements of these parameters are obtained will be discussed in chapter 3. The strength of anisotropy is expressed as a percentage k , which is proportional to the velocity difference between the fast and slow polarisations speeds [Babuska and Cara, 1991]:

$$k = \frac{v_{max} - v_{min}}{\bar{v}} \times 100 \quad (1.1)$$

where:

- v_{max} = Speed along the fast direction
- v_{min} = Speed along the slow direction
- $\bar{v}$ = mean of v_{max} and v_{min}

1.2.3 Origin of seismic anisotropy

There are two types of sources that cause the anisotropy in the crust:

1. Intrinsic rock properties
2. Extensive-dilatancy of fractures in the rock body resulting from regional stresses applied to the rock volume [Nur, 1971]

There are three main sources of intrinsic anisotropy. The first is the alignment of minerals along the direction of shear in fault zones, which results in anisotropy with the fast direction aligned parallel to the fault strike (eg [Zinke, 2000]). The

second is the dominant alignment of anisotropic minerals or grains in a parent rock (referred to herein as anisotropic domains), which causes the anisotropic fast direction of the parent material to align parallel to the anisotropic domains. Generally anisotropic domains are randomly oriented and the wavelength of the shear rays traversing the material is long enough to average out the anisotropic effects; however, if there is a dominant alignment of anisotropic domains in a material, it can exhibit significant amounts of anisotropy [Babuska and Cara, 1991]. In some metamorphic rocks, such as schist and gneiss, mineral alignment results in foliation planes parallel to the orientation of the strain applied during formation [Savage, 1999]. This phenomenon is commonly observed in the upper mantle where deformation causes a preferential alignment of olivine crystals in the direction of deformation (eg [Marson-Pidgeon and Savage, 1997], [Russo et al., 1996], [Margheriti, 1998]). By measuring the fast direction using SKS phases, which sample mantle anisotropy, the direction of maximum strain can be inferred [Savage, 1999]. The last potential source of intrinsic anisotropy is from horizontal layering. Propagation along the bedding planes allows the waves to refract around slow layers and travel in faster layers, whereas waves propagating perpendicular to the layering must travel through the fast and slow layers. Generally, the layering of bedding is horizontal, or near horizontal, and no significant splitting occurs for vertically incident waves [Savage, 1999].

Anisotropy in the crust is generally modelled by assuming the medium has a series of near vertical fluid filled micro-cracks [Savage, 1999]. Differential stresses in the crust preferentially close cracks that are aligned perpendicular to the maximum principle stress direction (S1) [Nur, 1971]. Shear-wave speed is decreased for waves traversing the open cracks, thus open cracks in the vertical plane result in anisotropy with the fast direction parallel to S1 [Crampin, 1994]. For S-waves propagating vertically, ϕ is parallel to the open cracks and to the maximum principle stress direction, and because of the assumed vertical crack orientation approximates the maximum horizontal stress direction S_{Hmax} (see figure 1.2). Thus by measuring the fast direction, ϕ , via shear wave splitting analysis, seismologists can infer the orientation of open cracks and measure the orientation of S_{Hmax} . Below

15 km, depth the lithostatic pressure, due to the weight of overlying rock, becomes sufficiently large to close many of the fluid filled cracks. Below this depth stress aligned micro cracks are no longer the dominant source of anisotropy [Nur, 1971]. At this depth, the main source of anisotropy is the dominant alignment of microscopic anisotropic crystals and grains within a material [Stein and Wysession, 1991].

It is generally considered that in the upper 15 km of the crust, the main source of anisotropy is due to the preferential alignment of fluid filled cracks [Crampin, 1994]. However, there may be significant contributions from intrinsic sources; the relative roles played by stress and structure in determining crustal anisotropy are not universally agreed upon (eg [Balfour et al., 2005]). It is therefore important to differentiate between the structural and stress induced anisotropy of a rock body before interpreting shear wave splitting measurements [Johnson et al., 2011].

FIGURE 1.2: Shear wave splitting due to fluid filled micro cracks, aligned to S_{Hmax} . The fast direction, ϕ , measured using shear wave splitting analysis aligns parallel to stress aligned fluid filled micro cracks (shown in blue); δt : delay time between the fast phase (S1, red) and slow phase (S2, blue); Grey lines are closed micro cracks anti-parallel to S_{Hmax} ; $spol$ is the polarisation of the incoming S-phase.

1.3 Tectonic setting

1.3.1 New Zealand tectonic setting

Deformation of the Earth's surface is driven by the relative motion of tectonic plates. At plate boundaries, interplate interactions cause high rates of surface deformation and seismic activity. New Zealand sits astride the boundary of two major tectonic plates, the Australian plate and the Pacific plate (see inset map figure 1.3). Convergence of the two plates results in high rates of surface deformation and seismic activity. Off of the eastern coast of the North Island the Hikurangi Trough marks where the Pacific Plate begins to subduct beneath the

Indo-Australian Plate, known as the Hikurangi subduction zone [Wallace et al., 2004]. In the North Island margin-normal interplate convergence is accommodated by subduction of the Pacific plate westward beneath the Australian plate, while margin-parallel motion is accommodated by a combination of strike-slip on the North Island Dextral Fault Belt (NIDFB) and clockwise rotation of the North island forearc [Wallace and Beavan, 2010].

In the northern part of the South Island a transition takes place from oblique subduction to oblique transpression. The plate motion through this zone is accommodated by the Marlborough fault system [Little et al., 1998]. Further to the south west the transpression is primarily accommodated along the Alpine Fault, a which extends 460 km along the western side of the south Island. Offshore to the southwest of the South Island the Australian plate subducts eastward beneath the Pacific plate, in the opposite sense to the Hikurangi subduction zone to the north, at the Puysegur subduction zone [Walcott, 1978].

1.3.2 Southern North Island tectonic setting

Off of the eastern coast of the southern North Island the old thickened oceanic crust of the Hikurangi Plateau is subducted beneath the southern North Island at an estimate rate of 42 mm/yr [Nicol and Beavan, 2003] (figure 1.3). Wallace et al. [2009] estimate that 80% of the margin normal component is accommodated by motion on the subduction thrust, with the remaining 20% accommodated by upper crustal reverse faults. The margin parallel component of plate motion is accommodated by strike-slip faulting (figure 1.3) and clockwise rotation of the eastern North Island [Wallace et al., 2009].

FIGURE 1.3: Map of the study region showing: Regional faults, labelled mountain ranges, and plate motion vector (Bold black arrow). Top left inset: New Zealand map showing the continental margin, and labelled features; Black vector shows subduction direction and magnitude at the southern Hikurangi margin; Black triangles along the continental margin point in the convergence direction.

1.3.3 Strain and locking

To understand the seismic hazard at plate margins, such as in New Zealand, it is necessary to understand the crustal deformation and stresses driving the deformation. If tectonic plates are unable to move past each other, due to friction on the plate interface, they become locked resulting in the accumulation of elastic stress and strain in the crust. Strain accumulation is expressed as a strain deficit, which is the rate of build-up of potential slip at the plate interface due to ongoing plate motion (eg [Beavan and Darby, 2004]).

The degree to which the interface between a subducting plate and an overlying plate is locked between slip events is described by the inter-seismic coupling coefficient, a dimensionless value between zero and one [Wallace et al., 2004]. A coupling

coefficient of zero is indicative of deformation that occurs with no associated seismic signal (aseismic) whereas a value of one indicates strong coupling where most of the deformation occurs as slip events that generate earthquakes (coseismic slip) [Wallace et al. \[2004\]](#). When plates are locked, the interface between the plates can only withstand a finite loading stress before it fails and slip occurs relieving the strain deficit. During the slip event, the elastic stress in the surrounding rock is released as seismic wave energy and heat. This either occurs as one large event over a short time frame, generating an earthquake or over a long period as a slow slip event (SSE) [\[Stein and Wysession, 1991\]](#).

Deformation at the Earth's surface can be observed by measuring relative offsets on positions of the Earth's surface through time. Modern techniques use networks of continuous GPS (cGPS) receivers to measure a surface velocity field. [Wallace et al. \[2012\]](#) simultaneously invert GPS velocities, earthquake slip vectors, and geological fault slip rates to model angular velocities of the North island, which is broken up into elastic crustal blocks, and the degree of coupling on faults separating the blocks. Using a model of the distribution of interseismic coupling on the subduction zone interface beneath the North Island, [Wallace et al. \[2012\]](#) infer a slip deficit in the over riding Australian plate (see figure 1.4). This model shows high rates of slip deficit at the southern end of the North Island due to strong coupling on the plate interface.

FIGURE 1.4: Subduction interface interseismic coupling resulting in strain (shown in terms of a slip deficit rate) for the Hikurangi subduction zone. Figure from: [Wallace et al., 2012]

The large strain accumulation on the locked plate interface at the southern Hikurangi Margin poses a large seismic hazard to the southern North Island; and rupture on the plate interface could result in a $M_w > 8$ earthquake [Wallace et al., 2004]. Although no subduction zone earthquakes have been occurred at the southern Hikurangi Margin in recorded history, geological evidence indicates that two subduction zone earthquakes have occurred of the past 1000 years; one 880-800 yr before present (B.P) which had synchronous upper plate activity, another 520-470 yr B.P with no identified upper plate fault earthquake associated with the event. The time interval between these events is ~ 350 years, which is shorter than the recurrence interval modelled for the National Seismic Hazard Model [Clark et al., 2015].

1.3.4 Stresses at the Hikurangi Margin

One of the primary factors controlling crustal deformation and faulting processes is the regional stress field [Townend et al., 2012]. To date there are few studies of the stress state in the overriding Australian Plate in the southern North Island that are associated with the subducting slab. To investigate the stress state throughout central New Zealand Townend et al. [2012] use P-wave first motion data for events with local magnitude greater than three to compile a set of 3424 focal mechanisms.

Using earthquakes in the upper 30 km Townend et al. [2012] infer that the maximum horizontal stress direction at the southern end of the Hikurangi margin to be approximately parallel to the strike of the subducted slab. However, the resolution of the model at the southern Hikurangi margin is relatively low, due to the large scale of observation.

FIGURE 1.5: Map showing the inferred stress regime in terms of the ratio $v = (S1 - S2)/(S1 - S3)$ across the North Island calculated from clusters of focal mechanisms. The cluster symbol denotes the stress regime (upright triangle - normal stress; square strike-slip; inverted triangle - reverse), symbol size denotes value of v , symbol colour represents depth. Black dashed lines - depth contours (figure credit: [Townend et al., 2012])

1.4 Isotropic crustal structure

1.4.1 Crustal heterogeneity

The structure and properties of the overriding wedge, subducting plate, and subduction interface can have significant affects on the tectonics in the overriding slab (eg [Eberhart-Phillips and Reyners, 2012]). Eberhart-Phillips and Reyners [2012] use a tomographic inversion of local earthquake phase arrivals to obtain three-dimensional models of V_p and V_p/V_s to image the Hikurangi subduction interface. Slices were taken though the three-dimensional volumes to make inferences about crustal structure and properties throughout the North Island. Figures 1.7 and 1.8 respectively show the V_p and V_p/V_s structures along the cross sectional slice $Y = -198$ shown by figure 1.6. This slice is in approximately the same location as an active source SAHKE seismic line (see section 1.4.2).

FIGURE 1.6: Map of the stations (plotted as triangles) used for the velocity inversion by [Eberhart-Phillips and Reyners \[2012\]](#). The rectangular grid shows the region that is inverted during the tomography. Black dashed line shows the position of the cross sectional slice, which is approximately same location as the SAHKE line

FIGURE 1.7: Cross sectional V_p structure taken as a slice at $Y=-198$ (figure 1.6) of the volume coloured velocity, obtained from the velocity inversion by [Eberhart-Phillips and Reyners \[2012\]](#)

FIGURE 1.8: Cross sectional V_p/V_s structure (coloured by V_p/V_s ratio) taken as a slice at $Y=-198$ (figure 1.6) of the volume obtained from the velocity inversion by [Eberhart-Phillips and Reyners \[2012\]](#)

Along the cross sectional slice of the tomography the margin interseismic coupling determined from cGPS studies by [Wallace et al. \[2009\]](#) is spatially correlated to the region with the highest V_p/V_s and sharpest velocity gradient [[Eberhart-Phillips and Reyners, 2012](#)]. This also correlates with a zone of high seismicity observed by [[Reyners and Eberhart-Phillips, 2009](#)]. These observations suggest that strong interplate coupling is related to fluids being unable to cross the plate interface. [Eberhart-Phillips and Reyners \[2012\]](#) propose that an impermeable metamorphosed layer traps fluids that are released as the subducting slab dehydrates.

1.4.2 Seismic Array Hikurangi Experiment

The Seismic Array Hikurangi Experiment (SAHKE) was a large onshore-offshore seismic survey, spanning from late 2009 until mid 2011, across the southern North Island of New Zealand [[Henry et al., 2013](#)]. The aim of the project was to image the plate interface, resolve physical properties of the locked interface, and to better understand the mechanics of slip that occurs on the subduction interface [[Henry et al., 2013](#)]. The project comprised two primary phases. The first phase, SAHKE-I, was an onshore/offshore wide-angle seismic reflection and refraction

survey, which also encompassed a passive instrument deployment from late 2009 to early 2010. The second phase, SAHKE-II, was an onshore active source experiment that was recorded on an array extending 350 km across the southern North Island, perpendicular to the strike of the subducting slab. The seismic network and survey duration of the SAHKE experiment was designed to collect crossover recordings of active sources, local earthquakes, and teleseismic events [Henry et al., 2013]. The entire data set provides good three-dimensional coverage of the southern North Island, with high spatial resolution of the subducting Pacific Plate. The data collected during the passive SAHKE-I deployment, complemented by data collected on the New Zealand GeoNet network during this time, comprises the data set used for this study.

Henry et al. [2013] use a first-arrival ray-tracing inversion technique to model the active source onshore-offshore data, producing a two-dimensional P-wave velocity model of the crustal structure beneath the study region, which was then used to depth migrate reflection phases and image the subducting slab. The P-wave velocity model and the reflection horizons were combined to build a cross-sectional image of the subducting slab along the 350 km transect, and structural boundaries and faults were inferred. On this cross section, Henry et al. [2013] project earthquake locations from Reyners and Eberhart-Phillips [2009] and compare the composite section to slip deficit measurements from Wallace et al. [2012] (figure 1.9).

FIGURE 1.9: Interpreted section of refraction and wide angle reflection imaging using data from SAHKE

Henry et al. [2013] make three key interpretations from these comparisons. Firstly, the authors propose that the convergence of the Pacific and Australian Plates underplates sheets of sedimentary material from the offshore Hikurangi Plateau into the footwall of the Wairarapa Fault, causing uplift of the Tararua ranges. Henry et al. [2013] also speculate that the underplated structures may inhibit fluid release from the subducting oceanic plate. This would cause an accumulation of fluid in the upper plate to the west of the underplated structure as seen by a high attenuation an V_p/V_s zone by Eberhart-Phillips et al. [2005]. Secondly, a distinct change in character and dip of the subducting slab is seen beneath the Tararua Ranges, which is spatially correlated with geodetic locking [Henry et al., 2013], suggesting that the slab dip defines the transition from stable to unstable slip regimes. Henry et al. [2013] lastly infer that abrupt slab-dip changes have implications for seismicity.

Many of the findings and work of Henry et al. [2013] were verified by the receiver function analysis of Hall [2013]. Hall [2013] used stacked P-wave receiver functions, computed from teleseismic earthquakes (recorded by seismometers deployed during SAHKE-I, complemented with data from three permanent short period stations deployed by GeoNet and three temporary stations from an earlier project), to obtain a model of the S-wave velocity structure of the crust along the SAHKE

transect. The receiver function stacks show a distinct, thick, westward dipping, low velocity layer (LVL), at mid-crustal depths of ~ 18 km depth in the east to ~ 30 km in the west. [Hall \[2013\]](#) interprets this layer as imbricated sediments associated with the subducting slab.

1.5 Anisotropic structure at the southern Hikurangi Margin

Crustal anisotropy studies have previously been carried out in the southern North Island by [Gledhill and Stuart \[1996\]](#), using local earthquakes recorded on the temporary Tararua array. The Tararua array consisted of nine broad-band seismometers located to the west of the Wairarapa Fault, positioned in an L shape with one length parallel and the other perpendicular to the Wairarapa Fault trace (figure [1.10](#)). To image crustal anisotropy in the overlying Australian Plate, earthquakes near the plate interface were used, and a fast direction of $51 \pm 18^\circ$ was obtained. The maximum delay times measured using slab interface earthquakes indicate a shear-wave anisotropy of 4% in the overriding plate. [Gledhill and Stuart \[1996\]](#) use deeper earthquakes to sample the subducting slab and the mantle beneath the slab, giving a fast direction of $41 \pm 15^\circ$. Both sets of measurements are approximately parallel to the strike of the subduction zone and other dominant geological features of the region. [Gledhill and Stuart \[1996\]](#) also observe $\phi = 28 \pm 11^\circ$ from SKS (S-wave phases that have travel through the mantle [[Stein and Wysession, 1991](#)]) splitting, which mainly samples anisotropy in the mantle beneath the subducting slab. This result is similar to that found in south Wellington by [Marson-Pidgeon and Savage \[1997\]](#) and confirms the presence of near surface anisotropy.

FIGURE 1.10: Tararua Array station locations (Purple triangles) used by [Gledhill and Stuart, 1996] to obtain measurements of seismic anisotropy.

Karalliyadda [2014] uses shear wave splitting analysis of SKS, ScS (S-wave phases which reflect of the Earth's core [Stein and Wyssession, 1991]), and teleseismic S-phases recorded by the SAHKE array to investigate the deep anisotropic structure at the southern end of the Hikurangi subduction zone, and to infer details of lithospheric deformation and mantle flow. Karalliyadda [2014] observe a general NE-SW trend of splitting fast directions across the southern North Island (in agreement with Gledhill and Stuart [1996]), which is interpreted to be caused by trench-parallel flow of the mantle. Karalliyadda [2014] also measures a change in teleseismic S, SKS, and ScS phases delay times of 1-1.5s (typical values for upper mantle anisotropy) to 2-2.5 s across the Wairarapa Fault. This equates to a change of approximately 0.5 s across the Wairarapa Fault and coincides with a slight change in ϕ . Furthermore, Karalliyadda [2014] finds that good quality ScS and teleseismic S-phases show a clear increase in δt with periods up to 11 s. The sudden change in delay times implies lateral variations of the anisotropic

structure, which was not expected in this region, given the shallow depth of the subducting slab and the absence of volcanic activity.

Because recording seismometers are on the Earth's surface, shear rays that are used to investigate anisotropy of mantle rocks also sample the crust. This means that contributions of the crustal anisotropy effectively contaminate shear wave splitting measurements used to interpret mantle anisotropy [[Savage, 1999](#)]. It was therefore proposed that lateral variations in the splitting parameters could be caused by crustal anisotropy from: crustal and upper-plate structures and properties in the shallow-wedge and subducting slab. It is hoped that crustal splitting measurements from this study can help explain the laterally varying structure observed by [Karalliyadda \[2014\]](#).

Chapter 2

Data and Methods

2.1 Earthquake data

Initially during the SAHKE-I 57 short-period and 10 broad band three component seismometers were laid out in the Wellington and Wairarapa regions (figure 2.2) from November 2009 until March 2010 (See [Seward et al., 2010] appendix 1 for further equipment details). The primary aim of the deployment was to record air gun source signals from the offshore PEGASUS survey; however, after the array was left in place to record passive seismicity after completion of the survey. During the deployment 425 earthquakes were identified on the GeoNet network (www.quakesearch.geonet.org.nz, last accessed 04/02/2016) (figure 2.1) Using the event origin times reported by GeoNet events the earthquake waveforms were found in the SAHKE stations and extracted from the continuous data.

To complement the temporary SAHKE array for shear wave splitting analysis, we use data from the permanent National seismograph Network (NSN) operated by GeoNet. After initial data checking and processing, 52 SAHKE stations and 7 Geonet stations were used for this analysis. Initial checks of the waveform data for SAHKE stations: S034, S035, S037, S038, S041, S042, were either: Too noisy, discontinuous, or contained periodic spikes (probably due to electrical interference). We consider that these stations will not yield robust splitting measurements and

were discarded from further processing. We also note that during the SAHKE deployment some of the stations were not correctly aligned in the horizontal plane, and station misalignments were obtained on retrieval of the seismometers [Seward et al., 2010]. We attempted to verify the misalignment values (appendix A); however, we were unable to obtain robust misalignment measurements and used the correction values of Seward et al. [2010] without further verification.

FIGURE 2.1: GeoNet earthquake locations of for the events used in this study (pale yellow circles)

FIGURE 2.2: SAHKE-I stations (red) and GeoNet stations (blue) used in this study. Stations are numbered with the corresponding name in table 2.1 below

1	S001	20	S020	41	S041
2	S002	21	S021	42	S042
3	S003	22	S022	43	S043
3A	S003A	23	S023	44	S044
4	S004	25	S025	45	LTW1
5	S005	26	S026	46	LTW2
6	S006	27	S027	47	LTW3
7	S007	28	S028	48	LTN6
8	S008	29	S029	49	LE2
9	S009	30	S030	50	LE3
10	S010	31	S031	51	LE4
12	S012	32	S032	52	WAIK
13	S013	33	S033	53	HOWZ
14	S014	34	S034	54	MSWZ
15	S015	35	S035	55	OGWZ
16	S016	37	S037	56	PAWZ
17	S017	38	S038	57	PLWZ
18	S018	39	S039	58	TMWZ
19	S019	40	S040	59	TRWZ

TABLE 2.1: Table of the station names and assigned numbers of the deployed SAHKE and permanent Geonet seismometers

2.2 P-wave arrivals

The first arriving energy from an earthquake rupture source is the compression body wave, known as the P- wave. Although the P-wave is not directly used in shear wave splitting analysis, it is required for determining earthquake locations. Knowledge of P-wave arrival times, also known as P-wave picks, are also useful for picking the arrival of later arriving S-waves, and are a required input for many S-wave picking algorithms, for example the [Diehl et al. \[2009\]](#) algorithm which is used by the SpickerC automatic S picker used in this project.

In this project P-wave picks are made using multiple processing steps. First we make automatic P-wave picks for all recorded earthquakes at a given station using the Automatic PicKing function (APK, [[Allen, 1978](#)], see section [2.2.1](#)) through the Seismic Analysis Code (SAC, [[Goldstein et al., 2003](#)], [[Goldstein, P., Snoke, 2005](#)]). Secondly TauP [[Crotwell et al., 1999](#)] was used to predict both the P- and S-wave arrival times, which are calculated based on the estimate earthquake origin time provided by GeoNet and the regional velocity model of [Henry et al. \[2013\]](#). These picks were used for guiding manual picking and the checking of the P- and S-phases.

FIGURE 2.3: Seismogram showing APK and TauP picks. Picks are marked by vertical bars through the seismogram traces. APK P-wave picks are made on the vertical trace only and are marked by a four character code to the left of the pick, in this case 'IPU0': impulsive; Phase: P-wave, polarisation: up; Quality: 0 (indicating best quality). TauP P- and S-wave picks indicated by a 'P' to the right, and 'S' to the left of the picks respectively.

2.2.1 APK Phase picking algorithm

The APK function makes P-wave picks by applying the short term average long term average ratio (STA/LTA) algorithm of [Allen \[1978\]](#) to the vertical component of a seismogram. [Allen \[1978\]](#) regards a seismic trace as a time series function $f(t)$ with a time derivative function $f'(t)$, which they called the first difference function. Using these functions and a weighting constant, which varies the relative weight assigned to an amplitude spike depending on the seismogram sampling rate, define a characteristic function $E(t)$:

$$E(t) = f(t)^2 + f'(t)^2 + C_2 \quad (2.1)$$

The characteristic function, is calculated over a 2 second time window; and is the LTA ($\beta(i)$). $\beta(i)$ is multiplied by a threshold constant to produce a reference noise level $\gamma(i)$.

The characteristic function $\alpha(i)$, is then calculated over a short time window of 0.01 s, and is the STA. For each each 0.01 s STA window is iterated over the 2 s LTA window . If $\alpha(i) > \gamma(i)$ an event is declared, and the absolute time, trace amplitude, first difference, and $\beta(i)$ are recorded. The program then applies further testing algorithms to search for the peak amplitude and the closest prior time, within the STA window, where the seismogram trace amplitude is zero (zero crossing). At each zero crossing, the $\alpha(i)$ undergoes further checks to test if it is sufficiently above the background noise to be considered as a pick. The LTA and STA windows are then iterated over the seismogram until a the first arriving P-wave arrival is identified.

2.2.2 P wave pick quality

Where relevant, manual picks made during the checking process were used preferentially to the automatic picks in further analysis steps. Overall, 21% of the ~ 29000 events P-picks made were visually inspected. This included of all of the events $M_w > 3$, 12% of the M_w 2.0-3.0 events, and 30% of the M_w 1.0-2.0 events. Of the 21% checked picks, 35% had no automatic pick because the data was too noisy; 5.3% of the APK P-wave picks were incorrect. Based on this sample, we estimate at least 90% of the automatic P-wave picks are reliable. Manual P-picks were on 13% of the checked events that had no automatic P pick. This indicates that the APK function likely picks less P-wave arrivals than a manual picker; however, given the size of the data set, using the automatic algorithm over manual picking exclusively is the most efficient use of the available resources of this project. Furthermore, automatic picking is objective and uses a consistent method to pick arrival times across the whole data set, whereas manual picking is subjective.

2.3 Picking S wave arrivals

Accurate determination of S-wave arrival times is vital to achieve accurate shear wave splitting measurements. Manually picking S-waves can be time consuming for large data sets and is subjective and can introduce both systematic and random bias. In order to avoid these shortfalls we elect to use the automatic S-wave picker program SpickerC [Castellazzi et al., 2015]. SpickerC is based on the Spicker method of Diehl et al. [2009], which combine three different phase picking methods:

1. The STA/LTA algorithm of Allen [1978]
2. Polarisation filters
3. Auto-regression modelling

A weighting scheme is also applied to the results of the three picking methods to determine a final pick with an error estimation [Diehl et al., 2009]. Castellazzi et al. [2015] make modifications to the Diehl et al. [2009] method, including requiring that the STA/LTA algorithm and polarisation filters use the same search window (which improves picking results), and the addition of a function to differentiate between converted S to P phases originating near the Earth's surface, which were being picked as S wave arrivals, and S waves.

Overall, SpickerC works by searching for an S-wave arrival within a course search window, which is constructed based on the estimated origin time from the given earthquake location and velocity model, the P-wave arrival time, and a minimum and maximum V_p/V_s ratio (kpmin and kpmax). Two sets of search window limits (kpmin and kpmax) are defined by the user, one set for epicentral distances less than 100 km and the other for epicentral distances greater than 100 km. Within this search window, SpickerC applies the STA/LTA method (section 2.2.1), and polarization method (section 2.3.2. The auto-regression method uses a different window to the first two methods; however, we will not discuss the window selection for this method here (see [Castellazzi, 2014] for more details).

For each event, the S-picks from the three different methods are used to derive a final S-wave pick and an associated pick classification. From each of the three picking methods, minimum and maximum arrival times are derived based on a weighting scheme, which depends on the event epicentral distance. The difference between the upper and lower pick times defines the pick error interval, and the mean of these two values becomes the final S-wave arrival time. The final pick is graded by quality of: class0, class1, or class2, which is based on the noise levels around the pick compared to user-defined noise levels, and the pick error interval (see table 2.2).

Class	Max error (s)	SNR_{min} dist ≤ 100 km	SNR_{min} dist > 100 km
0	0.2	3	3
1	0.3	3	2
2	0.4	3	2

TABLE 2.2: SpickerC thresholds for determining the quality grade (class) of a S-pick made by SpickerC (default values [Castellazzi, 2014]). Column 1: S-pick classification; Column 2: Maximum error of the final S pick; Column 3: Maximum pick error for a epicentre to receiver distance less than 100 km; Column 4: Maximum pick error for a epicentre to receiver distance less than 100 km.

2.3.1 Converted phases

Castellazzi et al. [2015] find that for some data sets S- to P- converted phases were arriving within the S picking search window and were being picked by the STA/LTA giving false S-wave detections, and reducing the reliability of the final pick time. In the Darfield region (Christchurch, New Zealand) they observed converted phases could have a ratio of maximum absolute value of the vertical component could be twice as large as the maximum absolute horizontal amplitude. This ratio was used as a threshold to differentiate between the two phases. Castellazzi et al. [2015] modified the methods to check if to the ratio of vertical to horizontal absolute maxima within the STA pick window is greater than two; If so then the pick is flagged as a converted phase and disregarded.

FIGURE 2.4: Seismogram showing APK and TauP picks. Picks are marked by vertical bars through the seismogram traces. APK P-wave picks are made on the vertical trace only and are marked by a four character code to the left of the pick, in this case 'IPU0': impulsive; Phase: P-wave, polarisation: up; Quality: 0 (indicating best quality). TauP P- and S-wave picks indicated by a 'P' to the right, and 'S' to the left of the picks respectively. SpickerC S-picks indicated by S 0 (0 indicates best quality S pick)

We use the default SpickerC parameters [Castellazzi, 2014], accept for those listed in 2.3. These changes have been made to enhance the number of class0 picks picked at the GeoNet station TMWZ.

Parameter	Value
STA W.L (s)	0.05
LTA W.L (s)	3
tMHA (s)	4
kpmin ($d > 100km$)	1.4
kpmax ($d > 100km$)	2.1
kpmin ($d < 100km$)	1.6
kpmax ($d < 100km$)	1.825

TABLE 2.3: Table of values for the input SpickerC parameters that were changed from the default values. The parameters in 2.3 represent the following: STA W.L: short term average window length, LTA W.L: long term average window length, tMHA: limitation of the start of the search window with respect to the time of maximum horizontal amplitude, kpmin: V_p/V_s minimum, kpmax: V_p/V_s maximum, d : epicentral distance

2.3.2 Polarization filter

In order to distinguish between P- and S-wave arrivals, [Diehl et al. \[2009\]](#) construct a time varying characteristic function, $CF(t)$ (equation 2.2), which is 0 when the particle motion is parallel to the incoming S wave propagation direction and 1 when the particle motion is perpendicular to the incoming propagation direction. To achieve this [\[Diehl et al., 2009\]](#) define a vector (L) that is parallel to the propagation direction of the incoming P-wave (See appendix A for further details). [\[Diehl et al., 2009\]](#) assumes that the incident angles of the P- and S-waves are equal, thus L is parallel to the S-wave propagation direction. The three component seismogram is then rotated from the original coordinate system (Z, E, N) into the new orthogonal ray system (L, Q, T). In this coordinate system Q is in the plane defined by Z and L , and T is normal to the ZL plane (See figure A.3).

The program then computes the covariance matrix over time steps within a centred time window for each sample of the rotated components, and computes the $D(t)$, $P(t)$, $H(t)$, and $W(t)$ elements of equation 2.2.

$$CF(t) = D^2(t) \times P^2(t) \times H^2(t) \times W(t) \quad (2.2)$$

Where:

- $D(t)$ is directivity. This is the normalized angle between L and the vector corresponding to the maximum eigenvalue of the covariance matrix at the considered time.
- $P(t)$ is rectilinearity. This value is 1 when particle movement is parallel too L, Q or T vector, and reduces as the misalignment of particle motion relative to these vectors decreases.
- $H(t)$ is the ratio of energy on the tranverse component (which is proportional to the square of the amplitude of the particle motion: $Q^2 + T^2$) component to the and total energy ($Q^2 + T^2 + L^2$). For a P-wave $H(t) = 0$ (no transverse particle motion), whereas for an S wave $H(t) = 1$ ($L^2 = 0$).

- $W(t)$ is a weighting factor used to reduce between the P and S-wave picks.

Where the CF is maximised, over the picking window (see section 2.3), a S-phase arrival time is picked.

FIGURE 2.5: Rotation of the seismogram from the seismometer coordinate system (Z, E, N; Solid black arrows) into the ray coordinate system (L, Q, T; Dashed black arrows). Blue dashed line earthquake ray path along BAZ; Red object: distance earthquake source; Double headed arrows indicate orientation of particle motion; ψ rotation of the (Z, E, N) coordinate system around Z axis (so the projection of L down onto the NE plane aligns with N); ξ rotation of the (Z, E, N) coordinate system around E axis (rotates Z down onto L).

2.3.3 Autoregression method (AR-AIC)

Autoregression refers to the pattern where time series data of a period exhibits a causal relationship with the prior-period data. Autoregressive (AR) models are used to forecast future values of a time series based on the past values of the

function. [Küperkoch et al. \[2012\]](#) present an automated AR algorithm for the prediction of S-phase arrival times. The method compares the predicted waveforms of the horizontal components within prediction window of a seismogram (using a character function obtained using AR on the waveforms) to the observed seismograms over the same prediction window. By applying the Akaike Information Criteria (AIC) to the characteristic function, the difference between the AR model and the observed seismogram can be estimated. Where the difference between the predicted and observed amplitudes is larger than the prediction error, a phase arrival is detected. To ensure the S-phase is picked [Diehl et al. \[2009\]](#) set a picking window around the predicted S-wave arrival time, based on the observed P-arrival time and the epicentral distance. The method is applied to the horizontal components of the seismogram in both the seismometer frame and the rotated coordinate frame used by the polarisation filter, and also to a combined stack of the horizontal components, to provide multiple pick times for error assessment of the final pick time.

2.3.4 S-wave quality

To test the quality of the S-picks across our dataset, we selected 10 stations and checked all of the class0 S-wave picks, which are used for making shear wave splitting measurements. Across these stations, approximately 10% of the class0 S-picks were unreliable and were either corrected or discarded. Based on this sample we estimate that 90% of the SpickerC class0 picks are reliable. Approximately 10 events were checked for all other stations (not presented in table [2.4](#)), which again returned reliability rate of approximately 90%.

Station	Number of class0 events	Number changed	Percent changed
LE2	100	2	2
OGWZ	115	3	3
PAWZ	50	0	0
S002	90	9	10
S004	83	8	10
S013	71	10	14
S020	35	1	3
S032	59	1	2
WAIK	73	13	18

TABLE 2.4: Table of sample stations used to check SpickerC S-wave pick reliability. The number of events changes is indicative of the number of unreliable points in the sample. The average percentage of unreliable events across the whole sample is *sim*10 %.

2.4 Multiple Filter Automatic Splitting Technique (MFAST)

We use the Multiple Filter Automated Splitting Technique (MFAST) developed by [Savage et al. \[2010\]](#) for the shear-wave splitting analysis in this project. MFAST uses three distinct processing steps to produce shear-wave splitting measurements:

1. Applies a set of band-pass to the seismogram in order to find the filter that produces the maximum product of the signal-to-noise (SNR) and bandwidth
2. Applies the shear wave splitting algorithm of [\[Silver and Chan, 1991\]](#) to perform a grid search of over ϕ , δt parameter space
3. Repeats the previous step to find $(\phi, \delta t)$ coordinates for a series of different measurement windows, then applies the cluster analysis technique of [Teanby et al. \[2004\]](#) to the $(\phi, \delta t)$ coordinates to determine the most stable solution.

2.4.1 Determining and applying the best filter

In order to make a high quality splitting measurement, we require a sufficiently high signal to noise ratio (SNR) over the analysis window [Savage et al., 2010]. This generally requires the application of a band-pass filter to remove frequencies outside the dominant bandwidth the incoming S-wave. Rather than applying one band-pass filter to all of the data, MFAST applies a set of 14 predefined bandpass filters to each event. The product of the SNR of the waveform and filter bandwidth (SNR bandwidth product) is calculated for the East and North components of each filtered seismogram by dividing the RMS amplitude of a signal window by a noise window in the P wave coda, and averaging the two RMS values. The three filtered waveforms that give the highest SNR bandwidth product are used for further analysis, if $\text{SNR} > 3$ (see figure for most commonly used bandpass filters for this data set).

FIGURE 2.6: Histogram of bandpass filters used by MFAST. This plot only shows the filters which MFAST used to make over 100 measurements.

2.4.2 Splitting correction

In order to determine the best splitting operator for a given waveform MFAST uses the splitting method of [Silver and Chan \[1991\]](#). This method iterates over all possible values of ϕ (in 1° increments) and delay times (in 0.0 - 1.0 s in 0.01 s increments) to estimate all possible $(\phi, \delta t)$ solutions. [Silver and Chan \[1991\]](#) construct a covariance matrix, which describes the covariance of the cross components of a rotated split wave and its parent non-split wave. Each possible solution is used to effectively un-split the waveform by rotating split waveform by ϕ and time shifting it by δt . The co-variance matrix is computed for each set of $(\phi, \delta t)$ coordinates; the coordinates that minimises the second eigenvalue of the covariance matrix are chosen as the best splitting parameters. This is equivalent to maximizing the particle linearity of the corrected split wave (see figure [2.7](#)).

2.4.3 Cluster analysis

Reliable and robust measurements are assumed to be stable over many different windows. Therefore good measurements should be contained within clusters of similar measurements [[Savage, 1999](#)]. MFAST applies the [Silver and Chan \[1991\]](#) method over multiple analysis windows and uses the unsupervised cluster analysis technique of [Teanby et al. \[2004\]](#) to identify parameters clusters. The best cluster is selected based on the number of measurements within the cluster and variance of the individual splitting measurements from the cluster mean (see figure [2.7](#)).

FIGURE 2.7: Example of a shear wave splitting measurement for station SAHKE station S002 using MFAST. (a) shows the 2-6 Hz bandpass filtered seismogram components, with an S-pick shown as solid red line. Dashed lines (1) and (4) mark the minimum start and maximum end times respectively. The grey shaded region is window of the final measurement. (b) Waveforms rotated to the determined incoming polarisation direction (p) and its perpendicular value ($p_{\perp}$). Times 2 and 3 are maximum start and minimum end times, 1 and 4 have same meaning as (a). (c) ϕ and δt measurement for each window vs. window number. Final measurement is shown by a blue cross. (d) Clustering of best measurements in $(\phi, \delta t)$ space. The cross indicates the final measurement from cluster analysis. (e) The top boxes show the split and the rotated and delayed waveform match, the bottom boxes correspond to the particle motion of the waveform above. (f) Same as (d) but also shows the contours of the smallest eigenvalue of the covariance matrix for the final measurement, cross is final measurement. The bottom label shows the grade assigned to the measurement and final ϕ and δt values with their associated uncertainty. Figure credit: [Godfrey et al. \[2014\]](#) (modified figure for our measurement)

2.4.4 Grading

To evaluate measurement quality MFAST automatically grades measurements of δt and ϕ based on the SNR , uncertainty, and distinctiveness of the clusters [Savage et al., 2010]. For this study only events which are graded B or better were used in the analyse of splitting results. That is measurements that fit the following criteria: Non-null (see the polarisation criteria in section 2.4.5 above), $SNR > 3$, δt less than the expected maximum anisotropy $\delta t_{max} = 0.8s$, less than 25° error in ϕ .

2.4.5 Null measurements

Null measurements result when material is isotropic or the shear waves travel through the medium in such a way that no shear wave splitting occurs. If a shear ray is incident on an anisotropic medium with an initial polarisation that is parallel to the anisotropic fast direction, then the particle motion associated with the shear wave is linear and entirely along the fast direction plane. This results in no shear wave splitting. Null measurements can also occur if the ray is incident with an initial polarisation perpendicular to the fast direction, as the associated linear particle motion is entirely in the slow plane. If the difference between the initial polarisation and ϕ is not between 20° and 70° , the result is considered null as no splitting is expected [Savage, 1999]. Null measurements can be unrepresentative of the mediums anisotropy and thus need to be identified so they can be neglected from interpretation [Savage, 1999].

2.4.6 Cycle skipping

Another potential source of error in shear wave splitting measurements is due to cycle skipping. This refers to the mismatch of slow and fast polarisations by an integer number of half cycles. Cycle skipping can potentially lead to the larger δt than is representative for the sampled anisotropy and a 90° rotation of ϕ . It

also can result in two or more distinct clusters of measurements in $\phi, \delta t$ space. Measurements with two or more distinct clusters are assigned a lower grade by MFAST's automatic grading system (section 2.4.4); so by only using grade A and B splitting measurements cycle skipping can be avoided. Also, to avoid cycle skipping MFAST uses time windows that include some of the signal before the S-wave onset, this allows the first peaks of the fast and slow polarisations to be matched accurately.

2.4.7 Converted phases

When a propagating shear ray is incident on a layer boundary, part of the incident wave energy is partitioned into a transmitted P-phase, this is called a S-P converted phase. S-P converted phases can affect shear wave splitting measurements and thus need to be avoided. This is achieved by defining a shear wave window in which incoming shear rays are not susceptible to P-S conversion, is angles of incidence in the range of vertical incidence $\pm$ the conversion phase critical angle. Any splitting measurements that are made using rays with incidence beyond the shear-wave window are rejected.

2.4.8 Incidence angles

For this study the angle of incidence for each wave arrival was computed by MFAST, which uses the TauP program of [Crotwell et al. \[1999\]](#), and a regional one dimensional velocity model of [Henrys et al., 2013](#). To calculate the critical angle we apply Snell's law on a regional average V_p/V_s ratio of 1.75 [[Eberhart-Phillips and Reyners, 2012](#)]. This gives a critical angle of approximately 35° , and a shear wave window of $\pm 35^\circ$. Prior to analysing the splitting results the raw splitting measurements are filtered by incident angle and measurements beyond 35 degrees incidence from vertical were taken to be outside of the shear wave window and excluded from the averages and any further analysis.

2.5 Delay time tomography

In order to constrain the regions of high anisotropy, obtain the strength of anisotropy and compute spatially averaged fast directions, we use the TESSA (Tomography Estimation of shear-wave Splitting and Spatial Averaging) method developed by [Johnson et al. \[2011\]](#). TESSA uses the splitting measurements output from MFAST to apply a two dimensional tomographic inversion to the delay times over a gridded region and infer the anisotropic structure of the study region. The tomography is run in three main steps:

1. Creation of a two dimensional rectangular box grid over the study region.
2. Computation of a matrix of synthetic distances of the incoming rays in each box.
3. Inverts the synthetic data to find the best fitting anisotropy model that reproduces the observed delay times.

2.5.1 Regional gridding

To run the tomography, the region must be gridded into two dimensional rectangular boxes. To determine the polygon size for the data inversion we use a quadtree gridding system. This method counts the number of rays that cross the study region and if there are more rays than 10 rays the then the grid box (the grid region for the first iteration) is divided into four smaller boxes of equal size. This process is carried out iteratively for each newly generated box until the number of rays in each box is either equal to 10, between 10 and 65, or smallest box dimension is 5 km, as been excluded from further analysis (less than 10 ray crossings).

There are a few advantages of using the quadtree gridding method over using a regular square grid for the TESSA inversion:

1. Quadtree gridding helps gain resolution in well-constrained areas and minimises block size of under-constrained areas

2. Reduced inversion computation time
3. The quadtree gridding system does not require a square region to obtain an invertible grid run produce area to grid

However, we note that regular gridding can be more useful for than quadtree gridding for making comparing delay time tomographies of the same region with a different set of data. If the tomography is run for an identical region using two different data sets, the quadtree grids produced for each study may differ because the gridding of the region is dependent on the ray paths of the input data set, and may influence the tomography results. Whereas setting a regular grid for each data set does not influence the results in this way.

2.5.2 Delay time inversion

[Johnson et al. \[2011\]](#) makes the first order approximation that the shear-wave splitting delay time of a measurement, linearly accumulate along the path of a shear ray and the total delay time for each grid is given by:

$$\delta t_r = \sum_{b=1}^n (s_b \times L_{rb}) \quad (2.3)$$

Where

- n is the total rays travelling through the box
- δt_r is the measured delay time of the ray r
- s_b is strength of anisotropy in a grid block
- L_{rb} is the length of the ray path through a grid block

Johnson et al. [2011] linearise the tomographic inversion problem and express it as:

$$G^T C^{-1} \underline{d} = (G^T C^{-1} G) \underline{m} \quad (2.4)$$

Where:

- G^T is the design matrix constructed by calculating the distance the ray travels in each block
- $\underline{d}$ is a vector containing the δt measurements for each ray
- $\underline{m}$ is the model solution of strength of anisotropy per block
- C is the error covariance matrix

This problem can be simplified to equation 2.5:

$$\begin{aligned} G^T C^{-1} \underline{d} &= (G^T C^{-1} G) \underline{m} \\ (G^T)^{-1} G^T C^{-1} \underline{d} &= (G^T)^{-1} (G^T C^{-1} G) \underline{m} \\ C^{-1} \underline{d} &= C^{-1} G \underline{m} \\ C \dot{C}^{-1} \underline{d} &= C \dot{C}^{-1} G \underline{m} \\ I_d \underline{d} &= I_d G \underline{m} \\ I_d (G \underline{m} - \underline{d}) &= 0 \end{aligned}$$

$$(G \underline{m} - \underline{d}) = 0 \quad (2.5)$$

The best possible fit to the above equation 2.5 is solved by finding a least squares solution. This is done using the *lsqlin* function in Matlab^{TM} subject to the bounding constraints that: $0 < m_i < \delta t_{max}$ for all values of the vector $\underline{m}$ [Johnson et al.,

2011]. This means that the minimum strength of anisotropy can not be negative or exceed the maximum observed delay time.

2.5.3 Checker board test

To test where structures are reliably resolved by the tomography the TESSA method incorporates a checkerboard test. Johnson et al. [2011] create a checkerboard grid with blocks of alternating anisotropy strengths of 0.01 km/s and 0.02 km/s and use the source to receiver ray paths of the splitting measurements to calculate synthetic delay times. By comparing the initial checkerboard grid and the model obtained by applying the delay time inversion to the synthetic delay times, we can assess where the delay time is resolved.

2.5.4 Spatial averaging of fast direction

In addition to the delay time tomography, the TESSA package spatially averages on each of the ϕ measurements within a grid block, using the same spatial gridding as the δt tomography. A weighting factor is applied to each ϕ measurement within a given grid block based on the inverse square of the distance from the station at which the measurement was made to the grid node. Once the inversion is performed circular histograms are output for each grid block and plotted on the grid node. If the standard deviation is less than 30° and the standard error of the mean is less than 10° degrees then the circular mean direction is also plotted.

2.5.5 Amendments to the TESSA methodology

To obtain a output checker board inversion grid, the synthetic design matrix for the checkerboard must be inverted. We found that for some iterations of the tomography that we ran that the design matrix for the synthetic data contained null rows and/or columns meaning the determinant of the design matrix is zero and no unique inverse exists. Typically the design matrix in question only contained four

null rows for all the synthetic data. To work around this issue we multiplied the average value of the design matrix by a random scale factor between 0 and 1, and inserted the value at a random position along the non zero row or column. This process was done once for each empty row and column, insuring that the design matrix had a unique inverse. For the tomographic inversions used in this study the design matrix was typically a square with 100 rows and columns; Given the scale of the design matrix it seems unlikely that the addition of a few randomly generated random points would have much effect on the matrix inverse, and thus the output resolution grid. To test if this work around had significant influence on the checkerboard tests multiple iterations of the tests were run. Because the positions of the random values were assigned at random points along the null vector, if the assignment of these values significantly effects the output checkerboard then the output should be significantly different for multiple iterations of the inversion with the same input parameters; however, we during testing we found no significant difference between the outputs.

2.6 Gravitational stress modelling

The Earth's surface topography arises due to additional or less mass above or below a fixed datum. This results in variations in the gravitational potential energy (GPE) and the strength of gravity, leading to differential stresses in the Earth's crust [Flesch et al., 2007]. We refer to these stresses this as topographic induced stresses. The fast directions (ϕ) measured in the crust are indicative of the stress contributions from all sources, including contributions from both topographic stresses and tectonic stress loading from the subducting slab (eg [Stern et al., 1992]). Therefore, when trying to understand the stress contributions from tectonic processes the effects of topographic induced stress must be understood. To do this we use the topographic stress model of [Hirschberg et al., 2015].

The overburden pressure, (also known in a geological context as the lithostatic stress) is the pressure exerted at depth by the weight of overlying material [Davis and Reynolds, 1996]. Because of the effect of lithostatic stress, absolute stresses in the Earth are compressional, and are defined as negative, and the stress due to gravity is described by a deviatoric stress tensor field (eg [Hirschberg et al., 2015]).

Flesch and Holt [2001] devise a method to solve force balance equations for vertically averaged deviatoric stresses associated with variations in GPE. Because the gravitational stress is vertically averaged, only the components of the tensor that are independent of the vertical component are of interest. Therefore, the stress tensor can reduced to two dimensions as follows:

$$\sigma = \begin{bmatrix} \sigma_{xx} & \sigma_{xy} \\ \sigma_{yx} & \sigma_{yy} \end{bmatrix} \quad (2.6)$$

The principle horizontal stress axes are the eigenvectors of equation 2.6, and the eigenvector corresponding to the most negative eigenvalue corresponds to the direction of maximum horizontal compressive stress. By evaluating the stress tensor in this way S_{Hmax} can be evaluated at each grid node for the was used to find S_{Hmax} at each point. This method was implemented using finite difference methods in

Matlab^{TM} by [Hirschberg et al. \[2015\]](#). Gravitational stresses are also influenced by variations in the density of materials in the Earth [[Telford et al., 1990](#)]. The stress modelling method of [Hirschberg et al. \[2015\]](#) do not account for stresses that arise due to density variations.

Chapter 3

Results and Analysis

3.1 Shear wave splitting results

Using 425 earthquake events we make 13,807 grade A and B shear wave splitting measurements on 50 stations. We present here tables of circular mean fast directions, mean delay times, and associated statistics for each station calculated using the MATLAB package of [Berens \[2009\]](#) (tables [3.1](#), [3.2](#), [3.3](#)). For this analysis, we divide our study area into three analysis regions based on the directional trend of mean splitting at each station, and fault structures which appear to delineate the locations of changes in fast direction azimuth. This particular binning is useful for seeing changes across the study region, which will become apparent later in the chapter.

Stations west of the Wairarapa Fault comprise the data set for the West Region, (figure [3.2](#), W, table [3.1](#)). The Basin Region is approximately bounded by the Wairarapa Fault to the west and the Dry Creek Fault to the east (figure [3.2](#), B, table [3.2](#)). The stations east of the Dry Creek fault make up the East Region (figure [3.2](#), E, table [3.3](#)).

FIGURE 3.1: Map of the faults in the SAHKE study area. We define the Wairarapa fault as the approximate boundary between the West and Basin regions. Dry Creek fault approximately marks the eastern extent of the boundary between the Basin and East regions.

FIGURE 3.2: Map of the SAHKE study area showing the spatial the West, Basin, and East regions (labelled W, B, E) that the map is broken into for analysis.

Station	N	$\mu(\phi)$	$\sigma_M(\phi)$	$\alpha_{95}(\phi)$	$R_M(\phi)$	$\mu(dt)$	$\sigma_M(dt)$
HOWZ	36	-31.315	8.715	17.081	0.373	0.068	0.005
LTW1	476	-89.362	1.712	3.356	0.505	0.158	0.007
LTW2	936	15.87	1.387	2.719	0.451	0.133	0.003
LTW3	1076	36.995	0.835	1.637	0.643	0.122	0.002
S023	148	46.261	2.137	4.189	0.667	0.158	0.01
S025	712	85.133	1.41	2.764	0.502	0.127	0.004
S026	280	42.773	8.11	15.895	0.148	0.127	0.006
S027	488	-4.273	4.298	8.425	0.211	0.065	0.002
S028	404	84.126	4.055	7.949	0.245	0.114	0.004
S029	160	57.736	6.309	12.366	0.25	0.109	0.007
S030	212	53.255	3.359	6.584	0.397	0.127	0.006
S031	176	14.049	3.028	5.935	0.474	0.173	0.013
S032	168	-70.456	2.144	4.202	0.637	0.158	0.006
S033	52	-59.904	5.428	10.639	0.485	0.183	0.017
S037	508	-55.091	10	19.601	0.09	0.128	0.005
S039	1172	-82.881	3.63	7.114	0.162	0.119	0.003
WAIK	116	51.413	2.498	4.896	0.652	0.168	0.012

TABLE 3.1: Average splitting measurements for each station in the West region shown on figure 3.2. N: number of measurements within the shear wave window, $\mu(\phi)$: circular mean of the fast direction, $\sigma_M(\phi)$: standard error of $\mu(\phi)$, $\alpha_{95}(\phi)$: 95% confidence interval of $\mu(\phi)$, $\mu(dt)$: mean delay time, $\sigma_M(dt)$: standard error of the delay time.

Station	N	$\mu(\phi)$	$\sigma_M(\phi)$	$\alpha_{95}(\phi)$	$R_M(\phi)$	$\mu(dt)$	$\sigma_M(dt)$
LE2	312	-41.377	4.096	8.028	0.275	0.145	0.006
LE4	3	64.333	1.395	4.439	0.999	0.297	0.002
LTN6	36	22.461	3.373	6.611	0.775	0.302	0.017
S010	792	-77.8	4.131	8.097	0.173	0.169	0.005
S014	144	-66.49	2.966	5.814	0.526	0.151	0.009
S015	556	4.038	4.964	9.729	0.172	0.168	0.003
S016	196	-48.727	6	11.76	0.238	0.187	0.008
S017	60	-4.579	12.126	23.767	0.213	0.179	0.018
S018	448	57.236	4.487	8.795	0.211	0.162	0.004
S019	260	-70.952	2.64	5.174	0.45	0.136	0.007
S020	93	88.562	8.306	16.28	0.249	0.128	0.009
S021	246	-53.956	7.863	15.412	0.163	0.175	0.009
S022	220	-72.887	3.622	7.098	0.364	0.097	0.005
S040	80	3.291	4.857	9.52	0.442	0.22	0.014
S043	88	33.353	5.305	10.397	0.391	0.142	0.01
S044	580	12.249	2.807	5.502	0.293	0.143	0.004

TABLE 3.2: Average splitting measurements for each station in the Basin region shown on figure 3.2. N: number of measurements within the shear wave window, $\mu(\phi)$: circular mean of the fast direction, $\sigma_M(\phi)$: standard error of $\mu(\phi)$, $\alpha_{95}(\phi)$: 95% confidence interval of $\mu(\phi)$, $\mu(dt)$: mean delay time, $\sigma_M(dt)$: standard error of the delay time.

Station	N	$\mu(\phi)$	$\sigma_M(\phi)$	$\alpha_{95}(\phi)$	$R_M(\phi)$	$\mu(dt)$	$\sigma_M(dt)$
LE3	364	64.73	3.335	6.537	0.31	0.141	0.082
MSWZ	39	73.002	4	7.839	0.685	0.31	0.178
PAWZ	33	-69.956	4.964	9.729	0.625	0.18	0.07
PLWZ	15	-72.494	11.499	24.504	0.432	0.068	0.018
S001	171	-7.546	11.91	23.343	0.13	0.129	0.081
S002	268	-80.529	5.231	10.254	0.233	0.135	0.094
S003A	192	29.907	6.974	13.669	0.207	0.142	0.099
S003	39	-23.957	8.446	16.554	0.37	0.17	0.078
S004	376	-18.252	8.661	16.976	0.12	0.116	0.074
S005	86	55.411	7.414	14.532	0.288	0.14	0.09
S006	105	81.936	2.253	4.416	0.72	0.195	0.077
S007	60	84.781	5.701	11.174	0.435	0.211	0.108
S008	153	55.096	4.396	8.616	0.36	0.254	0.172
S009	116	59.58	3.284	6.437	0.529	0.141	0.148
S012	268	-81.318	2.629	5.153	0.446	0.131	0.076
S013	272	53.564	1.553	3.045	0.674	0.196	0.099
TRWZ	16	65.113	7.889	16.725	0.579	0.147	0.07

TABLE 3.3: Average splitting measurements for each station in the East region shown on figure 3.2. N: number of measurements within the shear wave window, $\mu(\phi)$: circular mean of the fast direction, $\sigma_M(\phi)$: standard error of $\mu(\phi)$, $\alpha_{95}(\phi)$: 95% confidence interval of $\mu(\phi)$, $\mu(dt)$: mean delay time, $\sigma_M(dt)$: standard error of the delay time.

3.2 Splitting by region

To examine the spatial trend of splitting directions, we group the incidence-filtered splitting measurements into 10° bins. We plot circular histograms made at the station, with the mean fast direction superimposed (figure 3.3).

FIGURE 3.3: Map of all of the measured fast directions of grade B or greater for earthquakes of all depths. The purple plots show circular histogram of fast direction measurements in 10° bins. Yellow blocks show the circular mean fast direction as measured at each station. The red block indicates the circular histogram and mean for station S022 (discussed in section 3.3).

Overall the $\mu(\phi)$ of the stations in the West and East regions tends to trend NE-SW, approximately parallel with the Wairarapa Fault, whereas stations in the Basin region tend to trend NW-SE, perpendicular to the fault (figure 3.3). Station S022 (figure 3.3, red box) exhibits populations of splitting measurements along both of these trends.

3.3 Splitting populations

Many of the stations in figure 3.3 exhibit random scattering of ϕ measurements; this is likely to be caused by slight variations in the anisotropy seen by different ray paths, however some of the stations appear to have distinct populations of splitting measurements. To investigate whether ϕ is related to properties of the incoming ray, scatter plots were constructed for the parameters: back azimuth from the receiver to the source (BAZ), ϕ , δt , and incoming S-wave polarisation ($spol$). Visual inspection of the scatter plots for each station showed that most of the stations produced no clear clustering patterns, except for station S022, which exhibits clustering in (ϕ, BAZ) parameter space and $(spol, BAZ)$ space (figure 3.4). Also, we note that station S022 located close to the Wairarapa fault trace (figure 3.3).

FIGURE 3.4: Scatter plots of splitting parameters BAZ , ϕ , δt , $spol$ for station S022

At station S022 the majority of ϕ measurements are from a BAZ range of -60° - 40° (figure 3.4). This is reflected by the close alignment of the mean ϕ orientation and the fault perpendicular cluster (figure 3.3). Earthquakes from a BAZ of approximately 120° produce a cluster of ϕ measurements with an orientation of 60° , aligning with the strike of the Wairarapa Fault. Plot b shows the incoming

polarization pol is relatively independent of the ray back-azimuth BAZ ; However, there is a cluster at a BAZ of 120° that is related to a pol of $\sim 120^\circ$.

We think that the fault align cluster may result from rays incoming at 120° (approximately perpendicular to the fault), being polarised along the BAZ back azimuth and resulting in shear wave splitting measurements with the fast direction aligned with the Wairarapa Fault. It is unclear why this effect occurs on one side of the fault and not the other. Inspection of the scatter plots for other stations located along the Wairarapa Fault (S021, S023, LTN6, S039, HOWZ) did not show clustering of this nature. We therefore consider that the clustering observed by station S022 is a localized effect, which arises due to the stations close proximity to the Wairarapa Fault.

3.4 Dependence of delay time on ray path and frequency

Visual examination of scatter plots of the δt vs raypath shows that the delay time is not cumulative along the path length of the ray (figure 3.5). As the TESSA delay time tomography assumes the delay time is cumulative along the path length, this is an important consideration to make when interpreting the TESSA tomography results. Also, inspection of scatter plots of δt vs the dominant frequency of the incoming waveforms used to make the splitting measurements. This suggests there is no dependency of δt on the frequency of the incoming rays (frequency dependent splitting).

FIGURE 3.5: Scatter plots of delay time versus hypo-central depth (A), ray path (b) for splitting measurements made at station S022. Plot C shows the dominant S wave frequency vs delay time. This is a typical scatter plot for all of the stations used in the study.

3.5 Splitting by depth

In order to visualize the source locations of earthquakes for which we made high quality shear wave splitting measurements, we plot locations in relation to the subducting slab (see figure 3.6). The earthquake locations used in this analysis are preliminary GeoNet locations, which have no associated depth uncertainty estimate in the downloaded GeoNet meta data; however, even if the depths are inaccurate by as much as ± 5 km the features of figure 3.6 are expected too be similar, as most of the seismicity is clustered around the slab interface [Reyners and Eberhart-Phillips, 2009].

In order to constrain the location of the anisotropic sources, the splitting measurements were broken up into two depth bins: 1) Measurements from earthquakes above 30 km depth, 2) Earthquakes beneath 30 km depth. A third depth bin using earthquakes above 15 km depth was considered, as the main source of anisotropy in the crust is generally caused by stress aligned cracks in the upper 15 km (as explained in chapter 1); however, we think there are too few measurements made using earthquakes in this depth range to be representative of this depth range

(figure 3.6). Note that in the East region there are no events deeper than 30 km with an associated high quality shear wave splitting measurement.

FIGURE 3.6: Top plot: Map of earthquakes locations and depths in relation to the subducting slab. Bottom plot: Earthquake locations projected onto a cross sectional model of the slab interface [Williams et al., 2013]. Earthquakes above 30 km depth are coloured orange. Earthquakes below 30 km are coloured green. Red lines show the approximate bounds of the regions shown on 3.2. Region labels are indicated on the bottom plot. Note that only earthquakes that produced shear wave splitting measurements within the shear wave window (section 2.4.8) are plotted

We consider that in the Basin and West regions there is a visible agreement between the mean ϕ above and below 30 km, and observe that the regional circular mean ϕ for the above and below 30 km measurements are within 14° for the Basin, and 10° for the West region. Furthermore, for both regions the two sets agree within one standard deviation of their respective means. Also, we note that some of the deeper measurements tend to align better with the trend of the Wairarapa Fault (figure 3.7). Due to the similarity of mean ϕ for splitting above and below 30 km, it is likely that the measurements reflect the same source of anisotropy, i.e. the anisotropic body is located above 30 km depth. We therefore include measurements of all depths in any further analysis steps.

FIGURE 3.7: Map of station mean fast directions calculated using measurements obtained from earthquakes above 30 km (orange) and below 30 km (green). The earthquake locations of the events used for calculating these averages are shown in figure 3.6. Note that there are no events shallower than 30 km with high quality splitting measurements in the East region.

3.6 Gravitational stress due to topography

As discussed in chapter 3, the weight of topography on the lithosphere can significantly contribute to the stress field. To calculate the S_{Hmax} stress field for the study region (Hirschberg 2016, personal communication) construct a 20 km vertically averaged GPE grid with a node spacing of 5km for the Wellington region using the model of Lamb [2015] and applied the method of Hirschberg et al. [2015]. To enable comparisons between topographically induced S_{Hmax} (S_{Hmax}^{grav}) and azimuthal mean splitting direction (ϕ) at each station, we use a nearest neighbour search to obtain a single S_{Hmax}^{grav} for each mean splitting direction mean (figure 3.8, table 3.4).

FIGURE 3.8: Map of the nearest neighbouring S_{Hmax}^{grav} azimuths due to topographic gravitational stress loading for each station. There are changes in S_{Hmax}^{grav} that broadly correlate within the spatial bounds defined for the E, B, W regions.

We observe that in the East and West regions the the mean S_{Hmax}^{grav} orientations tend to align parallel to the trend of the Wairarapa Fault, and in the Basin region S_{Hmax}^{grav} orientations to trend perpendicular to the fault. These observations agree with the ϕ trends observe in section 3.2.

West	S_{Hmax}^{grav}	Basin	S_{Hmax}^{grav}	East	S_{Hmax}^{grav}
HOWZ	-85.15	LE2	16.261	LE3	34.225
LTW1	71.399	LE4	47.266	MSWZ	13.475
LTW2	63.504	LTN6	-14.82	PAWZ	47.37
LTW3	30.816	S010	37.966	PLWZ	20.93
S023	23.812	S014	-58.36	S001	45.349
S025	31.089	S015	-43.85	S002	8.4299
S026	44.902	S016	-0.95	S003A	59.529
S027	44.902	S017	-12.43	S003	-33.48
S028	46.404	S018	-22.74	S004	15.774
S029	80.017	S019	-21.91	S005	-26.74
S030	58.204	S020	-39.38	S006	78.768
S031	66.076	S021	19.505	S007	-31.58
S032	34.911	S022	14.029	S008	68.529
S033	86.137	S040	-44.39	S009	48.327
S037	62.208	S043	30.417	S012	46.63
S039	-31.32	S044	-32.44	S013	47.372
WAIK	-69.47			TRWZ	47.266

TABLE 3.4: Orientation of the maximum horizontal compressive stress (S_{Hmax}^{grav}) due to gravitational loading at each station. Columns 1, 3, 5 correspond to the regions West, Basin, and East; orientations are presented for stations where a station mean ϕ was obtained

3.7 Region Average delay times

For each station a mean delay time is computed for all measurements that are within the shear wave window (column 7 of tables 3.1, 3.2, 3.3). The station δt averages are then used to compute δt averages for the regional blocks (table 3.5). We found that the region average delay times for the East and Basin regions matched within uncertainty, but differed from the West region. However, the range of values for the West region matches the range of 0.01 - 0.2 s measured by Gledhill and Stuart [1996] on the temporary Tararua array, which was deployed west of the Wairarapa fault. By combining and re-averaging the δt for the East and Basin regions, we find a eastward increase of $0.034 \pm 0.017s$ across the Wairarapa Fault, which is an order of magnitude smaller than the delay time change observed by Karalliyadda [2014].

Region	N	$\mu(\delta t)$	$\sigma_M(\delta t)$	$\alpha_{95}(\delta t)$
W	17	0.132	0.0078	0.016
B	17	0.167	0.0117	0.025
E	16	0.165	0.0133	0.028
B + E	33	0.166	0.00895	0.018

TABLE 3.5: Average delay time for each region E: East, B: Basin, W: West, E + B: East and basin regions combined. The averages are computed by averaging the $\mu(\delta t)$ for the stations that are within the corresponding region.

3.8 Topographic stresses versus mean fast directions

We compare ϕ , S_{Hmax}^{grav} , and focal mechanism stress inversions (S_{Hmax}^{focal}) from [Townend et al., 2012] (figure 3.9).

FIGURE 3.9: Map of mean ϕ (yellow bars), Nearest neighbour S_{Hmax}^{grav} (blue bars), and S_{Hmax}^{focal} (orange bars).

From figures 3.9 we observe:

- Visible agreement between the regional trend of ϕ and S_{Hmax}^{grav} within each region
- Visible agreement between the trend of ϕ in the West and East regions
- A $\sim 90^\circ$ difference in the trend of ϕ and S_{Hmax}^{grav} in the Basin Region to the East and West regions

To test the significance of these observations we calculate the mean ϕ ($\mu(\phi)$) and S_{Hmax}^{grav} ($\mu(g)$) for each region using the station means in each subsequent region (see figure 3.10), and their associated 95% confidence intervals (see table 3.6). We also calculate a mean S_{Hmax}^{focal} orientation as 63.6° with a 95% confidence interval of 8.95° , which agrees with the West region $\mu(\phi)$ and $\mu(g)$ orientations.

FIGURE 3.10: Comparison of the regional mean ϕ (R. mean phi) and regional mean S_{Hmax}^{grav} (R. mean grav)

We find that within 95% confidence interval, there is no statistically significant difference between the region mean ϕ and S_{Hmax}^{grav} in the Basin and West regions

(see table 3.7); however, there is a significant difference in the East region. Using Monte Carlo simulations of the sample cumulative probability density functions for each set of ϕ and S_{Hmax}^{grav} means, we estimate a 95% percent confidence interval of $39.0 \pm 19.5^\circ$ for the difference of the two means in the East region. Note that the because of the circular nature of the azimuthal measurements the averages and difference in averages presented in tables 3.6 and 3.7 respectively have 180° ambiguity (i.e measurements with a difference of 180° are the same).

Region	N	$\mu(\phi)$	$\alpha_{95}(\mu(\phi))$	$R_M(\phi)$	$\mu(g)$	$\alpha_{95}\mu(g)$	$R_M(g)$
West	18	65.6	35.3	0.268	59.4	13.7	0.606
Basin	15	-44.3	70.7	0.144	-12.7	16.4	0.566
East	18	74.4	22.0	0.418	35.4	16.2	0.532

TABLE 3.6: Table of region averages for fast directions and S_{Hmax} azimuths from the gravity model. μ is the mean, α_{95} is the 95% confidence interval, and R_M is the length of normalized vector sum. The symbols that are denoted with a subscript ϕ are the fast directions, and those with g are S_{Hmax} azimuths. All values are given to 3s.f.

Region	$\mu(\phi) - \mu(g)$	$\alpha_{95}(\mu(\phi)) + \alpha_{95}(\mu(g))$	significantly different
West	6.20	49.0	no
Basin	-31.6	87.1	no
East	39.0	38.2	yes

TABLE 3.7: table

Difference of $\mu(\phi)$ and $\mu(g)$ for the three study regions, $\mu(\phi) - \mu(g)$ is the difference of the mean ϕ and S_{Hmax} for each region, $\alpha_{95}(\mu(\phi) - \mu(g))$ is the 95% confidence interval for the difference. The forth column states whether there is a significant difference between the means.

Comparing the calculated region mean ϕ orientations to the ϕ orientations calculated by Gledhill and Stuart [1996] for the Tararua array (figure 1.10) we find:

- The West region mean ϕ orientation agrees with the mean fast direction of $51^\circ \pm 18^\circ$, calculated using all Tararua array stations except for one outlying station (LTN4)
- The Basin region ϕ agrees with the $120^\circ \pm 17^\circ$ ϕ value of station LTN4

3.9 Delay time tomography

We use the TESSA delay time tomography and spatial averaging method of [Johnson et al. \[2011\]](#) (described in chapter 3) in attempt to reduce the scattering of ϕ measurements and constrain the regions where high delay times are observed. To determine the best spatial weighting for the tomography, iterations were run for weighting factors: $1/d$, $1/d^2$, and no weighting. For this testing, a quad tree gridding system was used with a 10 km minimum grid block size and 1 km point node spacing. All output models are consistent in their overall trend; however, the $1/d^2$ weighting model best replicates the station mean ϕ trend. Also, $1/d^2$ weighting results in the most pronounced variations across the Wairarapa Fault, and was therefore chosen for further analysis. To better resolve the changes across the Basin region the tomography was run on a subregion of the study area using a 5 km minimum grid and checker board block size with a 1 km node spacing. Higher resolution models were tested but the inversion became unstable using smaller grid blocks.

FIGURE 3.11: Ray plot of all of the events used for the TESSA tomography. Earthquake ray paths are indicated by a straight red line from the earthquake epicentre to a receiving station. Inverted blue triangles indicate station locations.

Figure 3.11 shows that the Basin region and the east part of the West region, have the a large number of ray crossings and a wide range of crossing azimuths. We therefore expect to have good resolution over this region.

FIGURE 3.12: Input checker board for delay time tomography. Inverted white triangles indicate station locations.

FIGURE 3.13: Output checker board for delay time tomography. Inverted white triangles indicate station locations.

Checkboard testing shows that the eastern part of the West region (which is close to the Wairarapa Fault), the Basin region, and most of the onshore part of the East region is well resolved. This is also shown in the tomography resolution map (figure 3.14).

FIGURE 3.14: Delay time tomography resolution map. The thick black line bounds the limits of the region that we can confidently resolve. Inverted white triangles indicate station locations.

FIGURE 3.15: Delay time tomography map. The thick black line bounds the limits of the region that we can confidently resolve (as shown in figure 3.14). Inverted white triangles indicate station locations. Numbered areas bounded by pink lines are discussed further in the text

Figure 3.15 shows regions of very high anisotropy (red and black patches); however, these regions are not well resolved (see figures 3.12 and 3.13). This is probably due to tomographic edge effects and are not interpreted as representing physical anisotropy. We therefore only consider delay time tomography results in the region bounded by the thick black line in figures 3.14 and 3.15. Figure 3.11 shows poor ray coverage for regions three and four, and a limited range of azimuths crossing the regions. We therefore consider that these regions are an artefact of the tomography.

We compared the spatially averaged ϕ means (figure 3.16) to the nearest neighbour S_{Hmax}^{grav} measurements and find no improvements upon the agreement between the fast ϕ measurements and S_{Hmax} from gravity models.

FIGURE 3.16: Spatially averaged ϕ measurements computed using a $1/d^2$ weighting factor on a quadtree grid with a minimum grid size of 5 km. Red plots are circular histograms of the spatially averaged splitting measurements. Yellow bars indicate the station mean fast direction calculated using the spatially averaged ϕ measurements. Inverted blue triangle indicate station locations.

Chapter 4

Discussion

In this chapter we discuss:

1. Crustal anisotropy in the southern North Island and its physical origin
2. Deviation of the total stress fields maximum horizontal stress component from the maximum horizontal stress due to gravitational loading
3. The origin of the observed stress deviations

4.1 Origin of anisotropy

We think that the main source of anisotropy in the crust of the southern North Island is due to the stress induced dominant alignment of fluid-filled micro-cracks in the crust. There are three reasons for this:

1. The agreement between the fast directions (ϕ) for splitting measurements made using earthquakes with hypocentre depths below 30 km and above 30 km (figure 3.6). This suggests that the main source of anisotropy is above 30 km depth, which is likely to be due to stress aligned cracks [Crampin, 1994]. Note that there is no relationship between the delay time and depth (figure

- 3.5). This implies that delay time is not consistent for different depth ranges and doesn't imply any limits on the depth extent of anisotropy.
2. In the West region there is a statistically significant agreement between the mean ϕ , S_{Hmax}^{grav} , and S_{Hmax}^{focal} values, implying that ϕ is related to S_{Hmax} .
  3. There is a generally a good agreement between the ϕ and S_{Hmax}^{grav} (from gravitational loading studies of [Hirschberg et al. \[2015\]](#)) azimuths for each station, and a no significant difference the region mean azimuths for the Basin and West regions (section 3.8).

However, we must also consider potential contributions of intrinsic anisotropic sources to the measured fast direction ϕ . In the Basin and West regions, 95% confidence intervals show there is no significant difference between the region mean orientations of ϕ and S_{Hmax}^{grav} , implying that the orientations of the means match (section 3.8). This suggests that the contributions of gravitational stresses dominate the contributions from intrinsic sources, or that any intrinsic sources are parallel to S_{Hmax}^{grav} . Thus we interpret the West and Basin regional average ϕ azimuths to represent the mean S_{Hmax} orientation within the upper 15 km of the crust for each region.

In the East region there is a significant deviation of ϕ from S_{Hmax}^{grav} of $39.0 \pm 19.5^\circ$. We have not been able to identify a media with strong anisotropic domains (eg high grade metamorphic rocks with their anisotropic domains aligned parallel to ϕ or faults oriented parallel to ϕ) that would develop the anisotropy observed in the East region. We therefore interpret the East region ϕ average azimuth to also represent the mean S_{Hmax} orientation.

We interpret that ϕ measurements represent the maximum horizontal stress component of total stress field from all contributed sources. This includes stresses due to tectonic loading and gravitationally induced stress due to topography (this is discussed in greater detail later in this chapter). We assume that over the time interval between consecutive megathrust earthquakes on the subduction interface, which we define here as one earthquake cycle, the change in S_{Hmax}^{grav} due to

changes in topography is negligible in comparison to the changes in the tectonic stress state. Thus we regard S_{Hmax}^{grav} as constant throughout the earthquake cycle. We also assume that after each subduction zone earthquake the tectonic loading stress is totally relieved. Therefore at the beginning of an earthquake cycle the total stress field represented by ϕ should align with S_{Hmax}^{grav} .

We propose that any significant deviation of the region mean ϕ from S_{Hmax}^{grav} is caused by tectonic loading stresses which have accumulated since the beginning of the earthquake cycle. To test this hypothesis we compare the expected orientation of S_{Hmax} , due to the topographically induced gravitational stress and tectonic loading since the last known subduction earthquake at the southern Hikurangi margin, to the observed stress deviation. By comparing the expected and observed deviations, we can infer whether the deviation could be caused by tectonic loading.

4.1.1 Comparison to deep splitting measurements

As detailed in chapter 1, [Karalliyadda \[2014\]](#) uses SKS, ScS, and Teleseismic (Tel. S) S-phases (all of which travel through the mantle) recorded on the SAHKE array to measure deep anisotropy in the southern North Island. [Karalliyadda \[2014\]](#) observes a dominant NE-SW trend of fast directions across the southern North Island, which is attributed to trench parallel flow. The NE-SW trend of the deep splitting studies is in good general agreement with the fast directions measured in the East and West regions of this study; however, the approximately 90° change in ϕ in the Basin region is not observed by the non local S phases of [Karalliyadda \[2014\]](#).

[Karalliyadda \[2014\]](#) also measures a ~ 0.5 s change in station average delay times across the Wairarapa Fault, from 1 – 1.5 s west of the fault (typical values for upper mantle anisotropy) to 2 – 2.5 s east of the fault. [Karalliyadda \[2014\]](#) proposed that this change is due to lateral variations in strength of anisotropy caused by structural and intrinsic properties of the subducting slab, shallow-wedge, and upper plate. We calculate a 0.034 ± 0.017 s westward increase in delay times across

the Wairarapa Fault (section 3.7), which is approximately an order of magnitude smaller than the delay time change observed by Karalliyadda [2014].

The discrepancies between the two sets of observations could be explained by the SKS, ScS, and Tel. S phases used by Karalliyadda [2014] sampling a deeper anisotropic body, which is not sampled by the shallower local phases; however, because the relative position changes in delay time for each study match, it seems likely that the delay time changes are due to the same anisotropic body. This discrepancy could also be explained by one anisotropic body whose strength of anisotropy increases with the period of the phases sampling it. This is supported by the Karalliyadda [2014]’s observation of frequency dependent splitting for ScS and teleseismic S phases periods up to 11 s (table 4.1). We consider that this is the most likely cause of the differences in observations between the two studies.

Phase	Signal period (s)	$\delta t(s)$	Inc in δt with T
Local S	0.17 – 0.5	0.16	No
ScS	3 – 7	1.1	Yes
Tel. S	8 – 13	2.2	Yes
SKS	9 – 15	1.8	No

TABLE 4.1: Table showing the S phase periods (column 2) and mean delay times ($\delta t(s)$) from deep S-wave phases from [Karalliyadda, 2014], and whether the phase exhibits frequency dependent splitting (Inc in δt with T)

Numerical models of Grechka and Mcmechan [1995] show that tilting a layered horizontally anisotropic medium results in a hexagonal symmetry with non-horizontal and non-vertical symmetry. This causes the apparent heterogeneity of the material to decrease as the dominant frequency of the signal increases [Grechka and Mcmechan, 1995], i.e. frequency dependent splitting. Using SKS phases Marson-Pidgeon and Savage [1997] observed frequency dependent splitting on the Wellington station, and attribute it to anisotropy within the dipping subducting slab. Frequency dependent splitting was also observed by [Karalliyadda, 2014] using ScS and teleseismic phases (table 4.1); however, in this study we did not observe frequency dependent splitting (figure 3.5). We hypothesis that this is because the range dominant periods for the local S phases is not large enough to observe the frequency dependent splitting, which is observed over a much larger frequency range

by Karalliyadda [2014] (table 4.1). However, we think that slab anisotropy alone does not explain the sudden change in delay times observed across the Wairarapa fault. To better understand the anisotropic structure causing these discrepancies, three dimensional modelling of the anisotropic bodies would be required.

4.1.2 Delay time tomography

Using the delay time tomography method of Johnson et al. [2011] we obtain a map of the two dimensional strength of anisotropy for our study region and identify six patches of high anisotropy (figure 3.15). Here we discuss the possible physical origin of anisotropy of the identified regions.

We think that the high anisotropy labelled region 1 (figure 3.15) is due to highly deformed and fractured fluid rich sediments, which are being driven by the subducting slab under the Aorangi ranges (see figure 1.3). As observed, we expect that the affect of this anisotropic body would decrease westward as it is dragged deeper beneath the overriding wedge by the subducting plate, and overburden pressure increases forcing the cracks to close dehydrating the sediments.

In section 3.7 we observe a 0.034 ± 0.017 s delay time increase westward across the Wairarapa fault (table 3.5); however, this change in delay times is not reflected by change in strength of anisotropy resolved by the delay time tomography (figure 3.15). In the West region the average hypocenter depth of the earthquakes that are used to obtain splitting measurements is deeper in the Basin and East regions (figure 3.6), thus the average source to receiver ray path length is longer. Given the longer average ray path length and higher anisotropy in the West region (resolved by the delay time tomography), we expect longer delay times in the West region than in the other two regions; however, this does not reflect the station delay time averages observed in section 3.7. The discrepancy between the delay time tomography and region average delay times probably due to the underlying assumptions of the TESSA model, i.e. The approximating a three dimensional system with an observed non-linear relationship between δt and ray path length

(section 3.4), with a two dimensional system with a linear relationship between δt path length [Johnson et al., 2011]. Due to this discrepancy we think that regions two, five and six probably do not have a physical origin, and that figure 3.15 is not representative of the anisotropic structure of the Basin and West study regions.

4.2 Strength of anisotropy

To estimate the strength of anisotropy, k , for the upper crust we use the following relations. (Note that we assume that a difference between propagation speed of the S-wave along the fast and slow axis is small):

$$\begin{aligned}
 \delta t &= L \left(\frac{1}{V_{min}} - \frac{1}{V_{max}} \right) \\
 \delta t &= L \left(\frac{V_{max} - V_{min}}{V_{min} V_{max}} \right) \\
 \delta t &= L \left(\frac{\Delta V_S}{V_{min} V_{max}} \right) && \text{assuming } V_{min} \approx V_{max} \approx V_S \\
 \delta t &= \frac{L}{V_S} \left(\frac{\Delta V_S}{V_S} \right) && \text{where } k = \frac{\Delta V_S}{V_S} \\
 \delta t &= \frac{Lk}{V_S} && \text{rearranging for } k \\
 k &= \frac{V_S \delta t}{L}
 \end{aligned}$$

where:

- δt is the observed station average delay time
- V_{max} is the speed of a S-wave travelling along the fast axis of anisotropy
- V_{min} is the speed of a S-wave travelling along the slow axis of anisotropy
- V_S is the mean shear wave velocity in the over riding wedge
- L is the average ray path length

For the East and Basin regions we use a delay time $\delta t = 0.166$ s, and $\delta t = 0.132$ s for the West region. We assume that the main source of anisotropy is approximately limited to the upper 15 km of the crust ($L = 15 \pm 5$ km). Using the

model of [Eberhart-Phillips and Reyners \[2012\]](#) we estimate the average $V_S = 3.5 \text{ kms}^{-1}$. For the East and Basin regions we use a delay time $\delta t = 0.166 \text{ s}$, and $\delta t = 0.132 \text{ s}$ for the West region. Substituting these values into equation 4.2 gives $k = 3.9 \pm 1.3\%$ for the Basin and East regions, and $k = 3.1 \pm 1.0\%$ for the West region. Both of these estimates agree with the estimated 4% anisotropy calculated by [Gledhill and Stuart, 1996](#) using the Tararua array, which is located within the bounds of the West region. Averaging the delay times for all of the regions gives $k = 3.7 \pm 1.2\%$.

4.3 Strain loading

As detailed in chapter 1, tectonic forces within the Earth can result in deformation at the Earth's surface. By observing such surface deformation, details of the orientations of tectonic forces within the Earth can be inferred. It is therefore important to consider the direction of strain when interpreting potential stress deviations, such as those described in the previous section.

[Walcott \[1978\]](#) use geodetic observations and earthquake slip vectors to produce horizontal strain rates for a 200 km wide deformation zone between the converging Australian and Pacific plates, which are averaged over a 50 year period at the southern end of the Hikurangi margin. [Lamb and Vella \[1987\]](#) decompose these horizontal strain rates into two orthogonal components: $\dot{\gamma}_1$ the shortening rate normal to the structural trend, and $\dot{\gamma}_2$, the shear rate parallel to the structural trend. Figure 4.1 shows that their inferred principal axis of compression (PAC) arcs across the southern North Island, indicating that there is a change in the direction of maximum compressive strain across the southern North Island. This pattern matches the deformational trends from more recent geodetic studies using cGPS [\[Wallace et al., 2009\]](#), which are averaged over a shorter time period of approximately 10 years. [Lamb and Vella \[1987\]](#) found that by considering complications due to local variations in the structural trend and accumulating elastic shear strain within the study area, the $\dot{\gamma}_1$ measurements are compatible with the

shortening strain rates derived from geological structures over the last 100 Ka. This suggests that the 50 year average geodetic strain rates are a good approximation to the 100,000 yr (100 Ka) geological strain rates, therefore we elect to use the strain observations of [Lamb and Vella \[1987\]](#) to compare to our observations of stress across the study region.

The calculation of horizontal strain rates shows that east of the Wairarapa Fault $\dot{\gamma}_1 > \dot{\gamma}_2$, indicating the region is deforming predominantly by compression oriented normal to the structural trend [[Lamb and Vella, 1987](#)]. However, west of the Wairarapa Fault $\dot{\gamma}_2 > \dot{\gamma}_1$, indicating that here the region is predominantly deforming in by shear parallel to the fault [[Lamb and Vella, 1987](#)]. This change in strain regime across the Wairarapa Fault matches our observations in section 3.8 of changing S_{Hmax} trend across the Fault and supports the hypothesis that the shear wave splitting measurements are indicative of the regional maximum horizontal stress orientation.

FIGURE 4.1: Map of the southern part of the North Island showing networks used in retriangulation studies, orientation of principal axis of compression (heavy dashed line), maximum shear strain rates, and the direction of shortening in the study area. The standard error in the orientation of the principal axis of compression is shown by the double fan shape, and the thin lines extending from each fan give an estimate of the standard error in the maximum shear strain rate. The light dashed lines parallel to the Wairarapa Fault show limit of area accumulating strain which could be released (figure from [Lamb and Vella, 1987]).

4.4 Stress analysis

We identify three major contributors (and potential contributors) of stress to the total stress field at the southern Hikurangi Margin:

1. Stress transmission into the upper crust due to flexure and upward bending of the subducting plate
2. Gravitational loading associated with topographic overburden
3. Tectonic loading due to the subducting Pacific plate

The net stress acting on a body is the sum of the stress acting on a particle [Serway and Jewett, 2007]. We express the total accumulated stress in the East region as:

$$\bar{\bar{S}}_{Total} = \bar{\bar{S}}_g + \bar{\bar{S}}_f + \left(\frac{d\bar{\bar{S}}_L}{dt} \times T\right) \quad (4.1)$$

Where:

- $\bar{\bar{S}}_i$ denote tensors
- $\bar{\bar{S}}_{Total}$ is the total stress in the crust since the last subduction earthquake
- $\bar{\bar{S}}_g$ is the average gravitational stress in the East region due to topographic loading
- $\bar{\bar{S}}_f$ is the average stress in the East region due to plate flexure
- $\frac{d\bar{\bar{S}}_L}{dt}$ is the average tectonic loading rate since the last subduction earthquake
- T is the time since the last subduction earthquake

If the stress components on the right hand side of equation 4.1 are estimated then the total expected stress field can be calculated as $\bar{\bar{S}}_{Total}$. Below we discuss and estimate each of the three stress contributions outlined above.

4.4.1 Stresses due to slab flexure and topography

We consider that there are two potential sources of stress associated with the subducting slab that may be acting on the over riding wedge: The first is upward forces applied to the upper wedge by the upward bending of the subducting plate; The second potential stress contribution is transmission of flexure stresses into the over riding plate (eg [Babeyko and Sobolev, 2008]). Both of these sources may contribute stresses into the overriding wedge, which will have an effect on the total stress field and thus the orientations of S_{Hmax} .

To analyse the effects of in-plane compression stress of the subducting Pacific plate on the overriding Australian plate at the Whanganui basin North Island New Zealand, Stern et al. [1992] model a three dimensional thin elastic sheet with spatially varying rigidity, which has a downward "slab pull" force applied to one end. This modelling showed that significant uplifting stress is applied to the overriding Australian plate and may cause the observe uplift around the Whanganui basin margins [Stern et al., 1992]. It is plausible that this process may occur further south in the Wellington region applying vertical and horizontal stresses to the overriding Australian plate; however, there been no investigation into whether this phenomena is occurring beneath our study region.

The transmission of flexure stresses into the over-riding wedge is dependent on the strength of the boundary between the wedge and the subducting plate. As detailed in chapter 1, the composite reflection-refraction tomographic model of Henrys et al. [2013] images underplated structures on top of the subducting slab, which are thought to be remnants from offshore sediments that are brought down by the subducting slab. Henrys et al. [2013] also speculate that these structures may form a fluid trap inhibiting fluid release from the subducting oceanic plate. This would cause an accumulation of fluid in the upper plate to the west of the underplated structure, and explain a low Qp high Vp/Vs zone seen by Eberhart-Phillips et al. [2005]. We propose that these structures would act to reduce the mechanical strength of the transition zone from the slab into the over-riding wedge,

severely reducing the magnitude of flexure stresses transmitted into the over-riding wedge.

Without producing detailed models of the subducting slab, we are unsure how stress is transmitted into the over-riding plate due to in-plane plate compression and plate flexure. Thus to estimate the expected stress deviations we neglect these stress contribution from our analysis and simplify equation 4.1 to:

$$\bar{\bar{S}}_{Total} \approx \bar{\bar{S}}_g + \left(\frac{d\bar{\bar{S}}_L}{dt} \times T \right) \quad (4.2)$$

Because we only consider stresses and strain rates in the horizontal plane we simplify equation 4.2 from a three dimensional tensor equation, to a two dimensional vector problem where the stresses are represented by stress forces in the horizontal plane:

$$\bar{S}_{Total} \approx \bar{S}_g + \left(\frac{d\bar{S}_L}{dt} \times T \right) \quad (4.3)$$

As discussed in section 3.8 in the East region the mean S_{Hmax}^{grav} gravity is 35.4° . Averaging the stress magnitudes of the S_{Hmax}^{grav} stress vectors from the GPE stress model of Hirschberg et al. [2015] gives a stress magnitude of 10.0 MPa. This vector is used as $\bar{S}_g$ in equation 4.3.

4.4.2 Tectonic loading stress estimation

To calculate the tectonic loading stress applied to the over-riding by the subducting slab, we model the Earth as a one dimensional elastic spring that is shortening parallel to the plate motion of the Pacific plate, in a plane horizontal to the Earth's surface. We use Hooke's Law to calculate the accumulated elastic stress since the last known rupture event ~ 500 yr BP [Clark et al., 2015] and use this value as the magnitude of the tectonic loading vector. For this calculation we assume that:

1. The loading vector is oriented normal to the strike of the subducting slab at an angle of -50° (since the subduction interface accommodates the normal component of motion).
2. The motion of the subducting slab is predominantly in the horizontal plane (Note that slab dip is approximately 8° beneath the east region [[Henry et al., 2013](#)])
3. The interface between the subducting slab and over riding plate has been locked since the last rupture.
4. There has been negligible loss of elastic potential energy since the last rupture event.
5. The last rupture event relieved all of the accumulated stress in the crust associated with loading stress due to the subducting slab.

Hooke's Law states that the force required to compress an elastic spring, of initial length L_0 , is proportional to the distance the spring is compressed by ΔL [[Stein and Wyss, 1991](#)]. This relation can be expressed by the modulus of elasticity E (also known as Young's modulus), axial stress σ , and strain ϵ :

$$\sigma = E\epsilon \quad (4.4)$$

The nominal strain ϵ is defined as the total strain normalised by the length over which the strain occurs and is expressed as:

$$\epsilon = \Delta L / L_0 \quad (4.5)$$

ΔL can be estimated as the product of plate motion convergence rate (from geodetic studies) and the time lapsed since the stress accumulation was last relieved.

$$\Delta L = vt \quad (4.6)$$

Where v is the tectonic stress loading rate normal to plate motion, $\frac{d\bar{S}_L}{dt}$, in equation 4.3.

Substituting equations 4.5 and 4.6 into 4.4 gives total exerted stress on the overriding plate since the last subduction zone earthquake in terms of measureable parameters.

$$\sigma = E(vt/L_0) \quad (4.7)$$

We use a deformation width of 200 km, which is the width of the emergent part of the deformation zone used by Walcott [1978] and Lamb and Vella [1987] to make the strain calculations discussed in section 4.3. Nicol and Beavan [2003] give a subduction rate of 40-43 $mmyr^{-1}$. Wallace et al. [2009] estimated that 80% of this is accommodated by the subduction thrust, which equates to a normal plate motion of approximately 34 $mmyr^{-1}$. Note that by using a 200 km deformation width gives a stress estimate over the entire study area, not just the East region where we observe significant stress deviations. The variation in stress deviations is discussed later in section 4.5.

- $v = 34 \text{ } mmyr^{-1}$
- $t = 500 \text{ yr}$
- $L_0 = 200 \text{ km}$: southern North Island width parallel to plate motion
- $E = 7.5 \times 10^{10} \text{ Pa}$: estimate for a Poisson solid earth [Stein and Wysession, 1991]

Substituting these values into gives a horizontal stress of exertion of 6.4 MPa (2 s.f.) since the last rupture event. Note that this estimate accounts only for stress components normal to the plate motion.

4.4.3 Expected vs. observed stress deviation

The addition of the gravitational loading and tectonic loading vectors gives a total stress vector $\bar{S}_{Total}$ (equation 4.3) of 12.0 MPa at 4.4° , resulting in a difference between the East region mean S_{Hmax}^{grav} and $\bar{S}_{Total}$ of 31.0° (figure 4.2 below).

FIGURE 4.2: Schematic figure of expected stress deviation. The blue vector represents the average stress due to topographic loading. It is oriented at an azimuth of 35.4° (see table 3.6) with a magnitude of 10.0MPa . The tectonic loading vector is oriented normal to the subduction margin at an angle of 310.0° (50.0° west of north) with a magnitude of 6.4MPa . The total S_{Hmax} stress vector is shown in black with an azimuth of 4.4° and magnitude of 12MPa .

The black dashed lines align north south.

There is significant uncertainty associated with these estimations. The three main sources are: Assuming no stress relief since the last subduction earthquake, not accounting for stress due to plate flexure and upward bending, and simplifying a complex three dimensional problem to two dimensional vector addition.

To observe how the addition of slab induced stresses may affect the expected stress deviation, we add in a slab bending contribution to the total horizontal stress field (S_{Hmax}^{slab}) of equal magnitude to S_{Hmax}^{grav} . We model this stress contribution as acting

in the same direction as S_{Hmax}^{grav} , effectively doubling the magnitude of the S_{Hmax}^{grav} vector. The Re-calculated expected stress deviation is 17° , which is close to the lower bounds of our observed stress deviation (section 3.8). Based on these first order approximations it is conceivable that tectonic loading stresses could cause the observed deviation.

4.5 Variation in stress loading

In the previous section we show how one component tectonic loading due to the subducting Pacific plate could cause the observed deviation between S_{Hmax}^{grav} and ϕ in the East region. We consider that there are four potential reasons for this, which we discuss respectively:

1. There is stronger plate coupling beneath the East region
2. The concentration of basal shear stress in the overriding block due to the subducting Pacific Plate is higher in the East region than the Basin and West regions
3. There is less stress release by seismicity and inelastic deformation in the East region than the Basin and West regions
4. The results in the Basin are not well determined and have a large confidence intervals (section 3.8)

4.5.1 Plate coupling

As detailed in chapter 1, geodetic studies indicate that there is a strong interseismic coupling on the subduction interface [Wallace et al., 2004]. This coupling is strongest east of the Wairarapa fault and reduces towards zero west of the fault (see figure 1.9), implying that the subduction interface west of the Wairarapa fault can slip relatively freely. We therefore expect that the transfer of basal shear stress

into the overlying crust due to the subducting slab to be greater on the eastern side of the Wairarapa fault (i.e in the Basin and East regions), than on the western side.

4.5.2 Shear stress concentration in the over-riding block

The relative concentration of stresses in the overriding plate influences the relative amounts of accumulated stress per volume of overlying material. We demonstrate this in terms of the ratio of a finite stress element (S) on the base of the plate to the depth to the plate interface (D):

$$X = \frac{S}{D} \quad (4.8)$$

Assuming the same basal stress S along the dip of the subducting plate

$$X \propto \frac{1}{D} \quad (4.9)$$

Due to the westward dip of the subducting slab the depth to the slab increases westward. Following equation 4.5.2 the basal shear stress exerted by the subducting Pacific plate, per kilometre of overlying rock, decreases westward (see figure 4.3). That is $X_e < X_b < X_w$ (see figure 4.3).

FIGURE 4.3: Schematic of the strain distribution in the over-riding plate (Not to scale). Pink lines illustrate the depth increase westward the depth; S elements represent finite elements of basal shear stress (from equation 4.5.2); dashed black lines so the depth to the plate interface in each region.

4.5.3 Release of stored elastic energy

In addition to a larger basal shear stress per kilometre of overburden rock, and stronger interseismic coupling of the plate interface, the East region has less active upper plate faulting (see 1.3) and an-elastic deformation than the Basin and West regions. The most active faults are the Wairarapa Fault and those west of it (data.gns.cri.nz/af, last accessed: 10/02/2016); We therefore expect a larger portion of the accumulated stress to be released by seismicity on active faults in the West region than in the Basin and East regions. The Basin region is a compressional zone region and is characterised by active folding and thrust faults [Lamb and Vella, 1987]. Deformation on these thrusts results in the release of elastic stress energy stored in the over riding plate. Because the East region has lower seismic activity than the Basin and West regions, more energy is stored in the over-riding plate, and is able to significantly deviate the gravitational stresses (due to topography).

Based on the estimated stress deviations, which match the observed deviations and since we have a physical mechanism to explain why there is a difference in stress deviation, we consider that there are higher levels of accumulated stress in the over-riding crustal wedge of the East region. Therefore, we infer that the

subduction interface beneath the East region may be later in its earthquake cycle (section 4.1) than the Basin and West regions.

4.5.4 Confidence in the deviation estimates

In section 3.8 we calculate the difference in the means values of S_{Hmax}^{grav} and ϕ for each region (table 3.6), and find that the difference in the means are similar in size for the East and Basin regions. However, in the Basin region the scatter parameter $R_{mean}(\phi)$ is small, indicating a large amount of scatter in ϕ . This results in a large 95% confidence interval for the mean of ϕ and no significant difference between ϕ and S_{Hmax}^{grav} for the Basin. However, if there was less scatter in the data we may see a significant difference between S_{Hmax}^{grav} and ϕ . This would indicate a stress deviation of similar magnitude to that observed in the East region, which would match the change in strain regimes across the Wairarapa Fault [Lamb and Vella, 1987]. However, since the size of the 95% confidence interval of the mean ϕ is relatively large the scatter would have to be reduced significantly.

The physical origin of the scatter of ϕ is unknown; however, we speculate it may be caused by localised complex geological structures associated with the thrust and folding sequences in the Basin region [Lamb and Vella, 1987]. To reduce the scatter in the Basin the region could be divided into smaller more homogeneous subregions effectively further grouping sets of measurements. The smaller subregions created by this process would contain fewer measurements than the original West, Basin or East regions reducing the confidence of the regional circular means of ϕ and the inferred S_{Hmax} orientations. Also, we consider that subdividing the Basin region (or East and West regions) into smaller more homogeneous regions introduces more complexity into our proposed stress model without increasing our understanding of the underlying sources of anisotropy in the Basin region. Furthermore, the bounds of the East, Basin and West regions are based on the location of important physical structures as well as visible spatial trends. Because further divisions of the regions would be based on visible spatial trends alone, further subdividing is not as strongly justified as the original regional divisions. To better

understand the sources of anisotropy in the Basin region a detailed investigation of the earthquake focal mechanisms of earthquakes to measure S_{Hmax}^{focal} should be undertaken. This would allow a comparison between S_{Hmax}^{focal} , which measures the orientation of stress S_{Hmax} without the influence of intrinsic anisotropy, and the anisotropic fast directions.

4.5.5 Additional stress model limitations

Other than the uncertainties associated with not accounting for plate flexure and bending stresses, there are potential unaccounted stress contributions from the gravitational field. The gravity model of [Hirschberg et al. \[2015\]](#) does not account for variations in density resulting in changes in the GPE field which the S_{Hmax}^{grav} orientations are derived from. The largest variation in density beneath the study region is due to the subducting slab, which is shallowest beneath the East region at a depth of approximately 17 - 20 km. Because the GPE is averaged over the top 20 km, the influence of the subducting slab on the gravity field is small in terms of an equivalent elevation difference and is considered unlikely to have a significant impact on the regional average gravity direction.

Chapter 5

Conclusions

5.1 Summary

The aim of this project was to use shear wave splitting analysis to obtain measurements of seismic anisotropy at the southern end of the North Island, and from this infer the orientation of maximum horizontal compressive stresses (S_{Hmax}) in the crust. We used Multiple Filter Automatic Splitting Technique (MFAST) [Savage et al., 2010] to perform shear wave splitting analysis on 425 earthquakes recorded on 44 temporary and six permanent seismometers, where P-wave arrivals were picked using automated methods of SAC's APK function ([Goldstein et al., 2003], [Goldstein, P., Snoke, 2005]) and S-wave arrivals by SpickerC [Castellazzi et al., 2015]. We obtained 13807 measurements with a quality of at least grade B. In order to smooth the fast direction station splitting measurements (ϕ) and constrain the areas of high anisotropy, we used the TESSA delay time tomography method of Johnson et al. [2011] and observed:

1. An area of high anisotropy in the East region, which is thought to arise due to highly fractured fluid rich sediments that are subducted beneath the Aorangi mountain range.

2. Spatial averaging does not improve the agreement between station mean ϕ directions and S_{Hmax}^{grav} , thus the spatially averaged fast directions is not used for further analysis.

As explained in chapter 1 in the upper 15 km the main source of anisotropy is generally stress aligned cracks [Crampin, 1994]. Due to too few quality shear wave splitting measurements for earthquakes with hypocentres shallow than 15 km we were unable to determine whether the trends were the same for earthquakes above 15km and of all depths. However, the trends of the station mean ϕ azimuths for earthquakes with hypocentres above 30 km and hypocentres below 30 km are similar. This suggests that the dominant source of anisotropy is limited to the upper 30 km of crust. A nearest neighbour search was done on the a gravitational stress model to obtain an estimate of S_{Hmax}^{grav} at the station. We observe that there is generally a good agreement between the alignment of S_{Hmax}^{focal} and S_{Hmax}^{grav} orientations, and the mean ϕ azimuths measured at each station, which we go on to show is statistically significant in section 3.8. This agreement, along with the restriction of the main source to shallower than 30 km depth, suggests that the main source of anisotropy is due to stress aligned cracks. Also, in section 4.3 we show that changes in region stress orientation across the Wairarapa fault, made using shear wave splitting measurements, are consistent with the change in strain regime observed by Lamb and Vella [1987] across the fault. This observation further supports the hypothesis that the shear wave splitting measurements are indicative of the regional maximum horizontal stress orientation.

Based on the trend of station mean ϕ azimuths and the regional structure, we divided the map into three sections: 1) A West region, bound to the east by the Wairarapa Fault, 2) A Basin region, bound to the west by the Wairarapa Fault trace and is approximately bounded to the east by the Dry Creek Fault, 3) An East region, bounded by the eastern edge of the Basin region. For each region we calculated the mean of the regional average: delay times, ϕ azimuth, and S_{Hmax}^{grav} orientation, using the station means corresponding to each region.

Delay time means for each region show an increase of 0.034 ± 0.017 s eastward across the Wairarapa Fault, which aligns with the position of the ~ 0.5 s change observed by Karalliyadda [2014]. However, the difference in delay time we observe across the fault is approximately an order of magnitude smaller than the ~ 0.5 s change observed by Karalliyadda [2014]. We attribute this difference to differences in the dominant period of the signal phases in each study, which have also been observed by Karalliyadda [2014] and [Marson-Pidgeon and Savage, 1997].

For each region we calculated the difference between the region mean ϕ and S_{Hmax}^{grav} orientations. In the East region we found a difference of $39.0 \pm 19.5^\circ$, and no significant difference between the means in the West and Basin regions at 95% confidence. We think this is caused by a larger accumulation of stress due to tectonic loading in the East region than in the Basin and West regions effectively deviating the total stress field from the gravitational stress field. The proposed mechanism for this is a combination of higher shear strain and lower rates of stress release by seismicity and inelastic deformation in the overriding block of the East region, than in the Basin and West regions.

By using the mean splitting delay time for the West region and the combined Basin and East regions, we calculate an average strength of anisotropy of the overriding plate of: $K = 3.1^\circ \pm 1.0^\circ$ for the West region, $K = 3.9^\circ \pm 1.3^\circ$ for the Basin and East regions. Both of these values agree with the 4% anisotropy calculated by Gledhill and Stuart [1996] for the Tararua array, which is within the bounds of the West region. We also find that the West region average $\phi = 65.6 \pm 18.0^\circ$ matches the Tararua array average $\phi = 51 \pm 17^\circ$ value, which excludes the outlying station LTN4 [Gledhill and Stuart, 1996]. Furthermore, the Basin average $\phi = 135.7 \pm 70.7^\circ$ agrees with $\phi = 120^\circ \pm 17^\circ$ observed for LTN4.

5.2 Future work and recommendations

Ongoing work is taking place to develop a complete focal mechanism catalogue for the SAHKE data set (Dominic Evanzia personal communication, 2016). This

will be a good comparative data set to the stress model presented here. Although not shown here, comparison of the initial results of Evanzias' work are in good agreement with the splitting results presented here.

As explained in section 4.1.1, the changes in anisotropy measured across the Wairarapa fault do not fully explain the high delay time changes across the Wairarapa fault or observed frequency dependent splitting observed by [Karalliyadda \[2014\]](#). As explained in chapter 4, we think that these changes may be due to the dipping anisotropy of the subducting slab. In order to determine whether this is a plausible mechanism, three dimensional modelling of the anisotropic subducting slab should be undertaken to try and reproduce the splitting observations of a composite data set of measurements from: 1) [Karalliyadda \[2014\]](#), 2) [Gledhill and Stuart \[1996\]](#), 3) The measurements presented in this study, and [\[Marson-Pidgeon and Savage, 1997\]](#).

Appendix A

Appendix A

A.1 Seismometer orientation

Correct seismometer alignment is crucial for obtaining reliable shear wave splitting measurements. We undertake further testing to verify the orientations of the seismometers deployed SAHKE. This section describes the techniques trailed to verify the seismometer orientations, which are: 1) A newly developed method, which uses a combination of emergent ray coordinate transformation and maximisation of the radial to transverse energy envelopes (Adam Carrizales, Personal communication 2015), 2) Seismogram rotation to maximise the cross correlation between the radial and vertical components [Stachnik et al. \[2012\]](#), 3) The ambient noise analysis method of [Zha et al. \[2013\]](#).

Generally, when a seismometer is installed the north component of the seismometer is aligned with true north, this means any azimuthal measurements made in the seismometer are the same as the geographic reference frame. During seismometer deployment true north is generally obtained using a compass to find magnetic north and the measurement is corrected for the local declination (22° west of magnetic North, in the Wellington region). During the deployment of the SAHKE array, there was a shortage of experienced personnel on the field teams to ensure the sensors were correctly aligned and as a result, many of the sensors were misaligned

[Seward et al., 2010]. Records from the retrieval teams suggest that the majority of miss-orientations were due to aligning the sensor 22° east, rather than west, of Magnetic North. However, it was also noted that there were also several randomly oriented sensors. This is put down to the possibility of magnetic interference of the sensor or other nearby metallic objects with the compass needle. During the retrieval of the instruments, the sensor orientations were re-measured to obtain the sensor misalignment. The orientations were then validated by observing the initial particle motion of impulsive P-arrivals from teleseismic events [Seward et al., 2010]. The authors estimate that there obtained values are accurate to within 5° .

A.1.1 Misalignment calculation

To correct a splitting measurement for sensor orientation miss-alignment the misalignment angle, γ , needs to be added to the observed splitting direction, ϕ_m . This gives a splitting direction relative to true north (see figure A.1), which we call ϕ_o :

$$\phi_o = \phi_m + \gamma \quad (\text{A.1})$$

FIGURE A.1: Splitting measurement, shown in blue, relative to a geographic (True north and east) and seismometer coordinate systems (N and E axes)

The miss-alignment angle, γ , is obtained by finding the difference between the incident angle of a earthquake ray in the seismometer reference frame and the incident angle in a geographical reference frame (see fig A.1). The incident angle relative

to the geographical reference frame is given as the back azimuth, which is based on the epicentral location of the event. The incident angle in the seismometers reference frame is calculated using the particle motion of the incoming earthquake ray on event seismograms. Here we make a fundamental assumption that the angle of incidence from incoming particle motion as measured in the seismometers frame is equal to the back azimuth in the seismometers frame. To obtain the correction angle (γ) we find the difference between the back azimuth β , which is based on the epicentral location of the event, and the angle of incidence of the incoming ray using properties of the particle motion α (figure A.2). We define the equation:

$$\gamma = \beta - \alpha \quad (\text{A.2})$$

where:

- γ = The station azimuth relative to geographic north
- α = Angle of incidence relative to the seismometer
- β = Event back azimuth

FIGURE A.2: The particle motion associated with a p wave arrival (shown by the black double headed arrow) aligns along the earthquake *BAZ*. The motion associated rayleigh wave is in the plane formed by the *BAZ* and the vertical component of the seismometer

Each of the techniques listed above uses the same fundamental principle to obtain the misalignment, only the method for obtaining the incident angle relative to the seismometer varies. Each of the orientation methods were used tested using the broadband GeoNet stations: ANWZ, CMWZ and HOWZ, which were assumed to have the correct orientations. Also, because the stations are broadband rather than short period we expect that these stations should yield better results than the short period SAHKE stations, making them more reliable for testing. With each of the trialled methods we are unable to get concordant results with control groups, therefore the correction angles provided from the SAHKE field report were used without further verification. In tables A.1, A.2, A.3 below we present station locations and the misalignment correction values for each region [Seward et al., 2010].

Station	Array	Latitude	Longitude	Elevation (m)	Correction angle
LE3	SAHKE	-41.27628	175.61702	300	0
S001	SAHKE	-41.53508	175.22843	234	0
S002	SAHKE	-41.50888	175.37565	99	0
S003	SAHKE	-41.50997	175.49651	112	0
S003A	SAHKE	-41.48447	175.22524	239	0
S004	SAHKE	-41.38245	175.37131	275	-7
S005	SAHKE	-41.45107	175.402458	200	7
S006	SAHKE	-41.42861	175.51995	154	12
S007	SAHKE	-41.47318	175.57005	43	22
S008	SAHKE	-41.36764	175.55321	211	0
S009	SAHKE	-41.33713	175.51059	343	-78
S012	SAHKE	-41.30778	175.57074	361	44
S013	SAHKE	-41.33421	175.63959	37	0
MSWZ	GeoNet	175.24946	-41.41583	403	0
PAWZ	GeoNet	175.42689	-41.38156	567	0
PLWZ	GeoNet	175.25479	-41.56918	638	0
TMWZ	GeoNet	175.89047	-41.1066	529	0
TRWZ	GeoNet	175.6879	-41.39804	379	0

TABLE A.1: Station locations and correction values for stations in the East region (figure 3.2)

Station	Array	Latitude	Longitude	Elevation (m)	Correction angle
LE2	SAHKE	-41.23397	175.53958	91	32
LTN6	SAHKE	-41.10332	175.3238	103	0
S010	SAHKE	-41.30275	175.39665	122	0
S014	SAHKE	-41.37822	175.16967	74	8
S015	SAHKE	-41.35454	175.27789	230	0
S016	SAHKE	-41.29509	175.3586	207	44
S017	SAHKE	-41.22548	175.36084	6	0
S018	SAHKE	-41.19346	175.41391	46	44
S019	SAHKE	-41.16566	175.33335	88	-28
S020	SAHKE	-41.26381	175.21662	2	-48
S021	SAHKE	-41.29601	175.1357	147	-98
S022	SAHKE	-41.22775	175.15938	57	32
S040	SAHKE	-41.08353	175.62844	173	0
S043	SAHKE	-41.26653	175.45663	91	6
S044	SAHKE	-41.1408	175.39705	33	0

TABLE A.2: Station locations and correction values for stations in the Basin region (figure 3.2)

Station	Array	Latitude	Longitude	Elevation (m)	Correction angle
LTW1	SAHKE	-40.96497	175.01906	176	17
LTW2	SAHKE	-40.99176	175.12222	212	-8
LTW3	SAHKE	-41.06623	175.1996	199	0
S023	SAHKE	-41.16433	175.22222	90	0
S025	SAHKE	-41.15699	175.08725	182	0
S026	SAHKE	-41.26463	174.98012	105	0
S027	SAHKE	-41.29669	174.95461	126	31
S028	SAHKE	-41.35687	174.89925	16	22
S029	SAHKE	-41.20336	174.84597	291	0
S030	SAHKE	-41.13292	174.91334	121	0
S031	SAHKE	-41.07584	174.94353	88	0
S032	SAHKE	-41.11444	175.00145	428	30
S033	SAHKE	-41.0166	174.9291	246	68
S034	SAHKE	-41.18959	174.78145	136	34
S035	SAHKE	-41.25059	174.71462	56	0
S037	SAHKE	-41.04383	175.0682	213	0
S038	SAHKE	-40.91226	175.12906	250	12
S039	SAHKE	-41.02804	175.3993	139	0
WAIK	SAHKE	-40.85654	175.03351	3	0
HOWZ	GeoNet	175.51509	-40.89733	475	0
OGWZ	GeoNet	175.1703	-40.81669	206	0

TABLE A.3: Station locations and correction values for stations in the West region (figure 3.2)

A.1.2 Rotapy and Protate

To obtain the seismometer orientations we combined two methods to obtain the angle of incidence (α), defined in equation A.2 above. The first is method uses *Protate* [Teanby et al., 2004] to rotate seismograms into the emergent ray coordinates. The second method, *Rotapy* (Adam Carrizales, personal communication 2015), assumes that the maximum ratio of the radial to transverse energy envelopes (R/T ratio) aligns along the back azimuth, and rotates the seismograms through 360° degrees in one degree increments to find the angle which maximises the R/T ratio.

To transform a seismogram from the recording seismometer component frame into the emergent ray coordinates, Teanby et al. [2004] defines a transform (R):

$$\begin{bmatrix} a & b & c \end{bmatrix} = R \cdot \begin{bmatrix} e & n & z \end{bmatrix} \quad (\text{A.3})$$

Protate works by decomposing R into two rotations making to align parallel to the in coming P wave, from a local earthquake event, with the vertical component of the seismometer. Firstly by a rotation clockwise ψ about the z axis to give a b z (effectively aligning the north component with the projection of the vertical component into the horizontal plane). The second rotation ξ is a clockwise rotation about the E axis, which effectively rotates the vertical component down onto the ray. (For a full mathematical description of this process see Teanby [2006]).

FIGURE A.3: Rotation of the seismogram from the seismometer coordinate system (Z, E, N; Solid black arrows) into the ray coordinate system (a, b, c; Dashed black arrows). Blue dashed line earthquake ray path along BAZ; Red object: distance earthquake source; Double headed arrows indicate orientation of particle motion.

The correction angle, α , is assumed to be equal to ψ , the coordinate rotation in the horizontal plane. ψ is calculated by fitting a line trend to the horizontal components over the defined P-wave window. The tangent of ψ is then calculated as the inverse of the gradient of the trend line, and ξ is calculated from the gradient of the projection of east and north components onto the horizontal line defined by ψ . The uncertainties of the angles ψ and ξ are given as $d\psi$, $d\xi$. These values are found by calculating the angles for a gradient which has been adjusted so that

the variance of the trend line changes by 5% of the variance of the best fitting gradient.

The *Protate* program reads in a three component seismogram; a user then defines a pick window around the P-wave pick. This step was automated by reading in seismograms with P-picks already in the file header and defining a window length based on one period of the P-wave, corresponding to the dominant frequency of the P-wave arrival T_p . The start of the window P-wave window ts , is set equal to the p-wave pick and the end time tf was defined as one period on from ts . ($tf = ts + T_p$)

The rotapy method finds α by rotating a user defined P-wave seismogram window through 360° in 1° increments and calculating the ratio of the radial to the transverse energy envelope (R/T) for each angle. Because the motion associated with a P-wave is parallel to the propagation direction α , the angle at which R/T is maximised should be equal to the back azimuth of the incoming P-wave. The original Rotapy program requires the user to select window around the P-wave; however, we find it difficult to remain consistent when selecting the same part of the waveform for analysis. To make this process less subjective, the programme was automated by using the same window that was defined for the *Protate* program.

For testing purposes, only earthquake events that have a S-pick from the automatic splitting Spicker codes (section 2.3) were used, as this ensures only high quality seismograms are used. For each event we apply a 5 Hz - 45 Hz butterworth bandpass filter and calculate the SNR over the defined P-wave window. The program outputs an ascii file containing: event ID, BAZ , ψ , $\delta\psi$, ξ , $\delta\xi$, $ppick$, ts , tf , snr . From this we estimate the: incident angle as $I = \xi - \delta\xi$ and linearity factor as $lin = \frac{\delta\psi}{\psi}$.

For testing we used events that had a $SNR > 5$, $I > 20$ (so most of the P-wave energy was on the horizontal components), $lin < 0.15$. The program was tested for the three previously mentioned GeoNet control stations. (Note that because the GeoNet stations are correctly aligned, the back azimuth, β is approximately

equal to α for all of the events. This means that: and the average correction angle γ should equal zero, and a plot of α against β should show a positive linear trend with a gradient of one. However, the tested GeoNet stations do not definitively show this trend (e.g. figure A.4) below demonstrates this for the geonet station BSWZ.

FIGURE A.4: Scatter plot of the correction angle vs the earthquake back azimuth for measurements used to orient the GeoNet station BSWZ. Blue dots: Rotapy measurements; Red dots: *Protate* measurements

Station	Mean rotation	standard deviation	Number measurements
ANWZ	-25.9	50.1	25
BSWZ	0.8	57.9	46
HOWZ	21.4	50.1	23

TABLE A.4: Results from the rotapy program for three GeoNet test stations

Comparing the seismograms that have been rotated onto the back azimuth shows that there is still significant energy on the transverse component (figure A.5). When the energy on the transverse component is minimized, the energy difference

between the radial and transverse components still appears to be small. Figure A.4 demonstrates that the R/T ratio is not necessarily maximised when the seismogram is rotated, thus the north component is not parallel to the back azimuth of the incoming earthquake. This maybe due to an over simplification of the particle motion of the incoming P-wave.

FIGURE A.5: Station BSWZ. Left top: Aligned to geographic North, Right top: middle Aligned to BAZ, Bottom Left: Aligned to ϕ . The top trace for each seismogram is the the transverse component being minimized.

A.1.3 Stachnik method

Rayleigh waves are surface waves that exhibit retrograde elliptical particle motion in the vertical plane along the direction of wave propagation [Stein and Wyssession, 1991]. This means the particle motion associated with a Rayleigh wave should only be observed on the vertical and radial components of a seismometer, and not on the transverse component. Rather than by measuring ellipticity, Stachnik et al. [2012] perform a polarisation analysis over a Rayleigh wave window. A Hilbert transform is applied to the vertical component, which phase shifts it by 90° , aligning it with the radial component and giving a linear relationship between the vertical and radial components. The largest radial component is expected when

the seismometer is rotated to align parallel with the back azimuth; the amplitude on the vertical component is not expected to change with sensor alignment. [Stachnik et al. \[2012\]](#) rotate the seismogram through 360° in 1° increments and cross correlate the Hilbert transformed vertical and radial components. The maximum positive correlation coefficient should occur when the radial component aligns with the back azimuth of the earthquake.

For our analysis, earthquakes from the USGS catalogue that occurred within $30^\circ - 90^\circ$ of latitude of the study region, were cut into one hour long SAC files from the SAHKE continuous data set. This totalled eight teleseismic events, although not all of the events were visible on all of the stations. The GeoNet station WEL, was used as a control to test the method. WEL is a broadband station at a hard rock site in Kelburn Wellington (magma.geonet.org.nz/delta). The instrument is housed in a purpose-built vault with a concrete floor, so it should be subject to minimal background anthropogenic noise.

To find the optimal waveform for processing we adjusted the predicted Rayleigh wave speed, Rayleigh wave window parameters, and frequency filter; however, we find that the adjustments have very little effect on the final result. We therefore use default parameters of [Stachnik et al. \[2012\]](#) listed below. We also removed the instrument response, but this also had very little effect on the results.

- The Rayleigh wave speed was assumed to be 4.0 km/s (default value from [Stachnik et al. \[2012\]](#))
- Window set 20 s before and 600 s after the predicted event arrival time.
- 0.03 - 0.1 Hz (10 - 30 s) butterworth bandpass filter
- Waveform window with a 10 percent cosine taper

Four of the eight teleseismic events extracted from the continuous SAHKE data were clearly identified on the GeoNet WEL station. The calculated correction angles ranged from $220^\circ - 340^\circ$ with a mean of 250° , when the expected result was

a mean of 0° . This method was also trailed on other GeoNet stations with little success.

A.1.4 Ambient noise cross-correlation analysis

This technique was developed by [Zha et al. \[2013\]](#) for determining the orientations of ocean bottom seismometers. The method finds the optimal sensor orientation, α (see equation A.1 above), by obtaining the azimuth that maximizes the correlation between the Radial-Vertical and Vertical-Vertical cross components of the greens tensor for ambient noise propagating between pairs of stations.

This method was trailed by VUW post-doc Adrian Shelley and PhD student Francesco Civilini (personal communication, 2015) using Rayleigh waves in the frequency range 0.1-0.5 Hz. The method requires obtaining good quality cross-correlations, which we were unable to obtain for this data set. The method was thus neglected. Visual inspection showed that some of the continuous data was visually of poor quality with gaps and trace errors. It is also suspected that during pre-processing the data may have been down sampled destroying some of the record making it unusable for noise analysis of this frequency range. Also, most of the SAHKE stations had a 1 Hz natural frequency this would reduce the quality of the cross-correlations because the ambient noise signal is typically below 1 Hz. It also should be noted that a longer set of continuous is preferable for this method.

Bibliography

- Allen, R. V. (1978). Automatic earthquake recognition and timing from single traces. *Bulletin of the Seismological Society of America*, 68(5):1521–1532.
- Babeyko, A. Y. and Sobolev, S. V. (2008). High-resolution numerical modeling of stress distribution in visco-elasto-plastic subducting slabs. *Lithos*, 103(1-2):205–216.
- Babuska, V. and Cara, M. (1991). *Seismic Anisotropy in the Earth*, volume 10. Kluwer.
- Balfour, N. J., Savage, M. K., and Townend, J. (2005). Stress and crustal anisotropy in Marlborough, New Zealand: evidence for low fault strength and structure-controlled anisotropy. *Geophysical Journal International*, 163(3):1073–1086.
- Beavan, J. and Darby, D. (2004). Fault Slip in the 1855 Wairarapa Earthquake based on new and reassessed vertical motion observations: Did slip occur on the subduction interface? pages 31–41.
- Berens, P. (2009). CircStat: A MATAB Toolbox for Circular Statistics. *Journal of Statistical Software*, 31(10):1–21.
- Castellazzi, C. (2014). Manual for the use of the Automatic S-Wave Picker modified version from T. Diehl & al [2009] for MFAST application Claire Castellazzi Table of Contents. Technical Report July.

- Castellazzi, C., Savage, M. K., Walsh, E., and Arnold, R. (2015). Shear wave automatic picking and splitting measurements at Ruapehu volcano, New Zealand. *Journal of Geophysical Research: Solid Earth*, 120(5):3363–3384.
- Clark, K. J., Hayward, B. W., Cochran, U. A., Wallace, L. M., Power, W. L., and Sabaa, A. T. (2015). Evidence for Past Subduction Earthquakes at a Plate Boundary with Widespread Upper Plate Faulting : Southern Hikurangi Margin , New Zealand. *Bulletin of the Seismological Society of America*, 105(3):1661–1690.
- Crampin, S. (1994). The fracture criticality of crustal rocks. *Geophysical Journal International*, 118(2):428–438.
- Crotwell, H. P., Owens, T. J., and Ritesema, J. (1999). The TauP Toolkit : Flexible Seismic Travel-Time and Raypath Utilities. *Seismological Research Letters*, 70:154–160.
- Davis, G. and Reynolds, S. (1996). *Structural geology of rocks and regions. 2nd edition*. Wiley, 2 edition.
- Diehl, T., Deichmann, N., Kissling, E., and Husen, S. (2009). Automatic S-Wave Picker for Local Earthquake Tomography. *Bulletin of the Seismological Society of America*, 99(3):1906–1920.
- Eberhart-Phillips, D. and Reyners, M. (2012). Imaging the Hikurangi Plate interface region, with improved local-earthquake tomography. *Geophysical Journal International*, 190:1221–1242.
- Eberhart-Phillips, D., Reyners, M., Chadwick, M., and Chiu, J. M. (2005). Crustal heterogeneity and subduction processes: 3-D Vp, Vp/Vs and Q in the southern North Island, New Zealand. *Geophysical Journal International*, 162(1):270–288.
- Flesch, L. M. and Holt, E. (2001). Dynamics of the India-Eurasia Shortening-. *Journal of Geophysical Research*, 106:435–460.
- Flesch, L. M., Holt, W. E., Haines, A. J., Wen, L., and Shen-tu, B. (2007). The dynamics of western North America : stress magnitudes and the relative role

- of gravitational potential energy , plate interaction at the boundary and basal tractions. *Geophysical Journal International*, 169:866–896.
- Gledhill, K. and Stuart, G. (1996). Seismic anisotropy in the fore-arc region of the Hikurangi subduction zone, New Zealand. *Physics of the Earth and Planetary Interiors*, 95(3-4):211–225.
- Godfrey, H. J., Shelley, A., and Savage, M. K. (2014). Search for temporal changes in shear-wave splitting associated with the 2012 Te Maari Eruptions at Mount Tongariro , New Zealand. *Journal of Volcanology and Geothermal Research*, 286:277–293.
- Goldstein, P., Dodge, D., Firpo, M., and Minner, L. (2003). 85.5 SAC2000: Signal processing and analysis tools for seismologists and engineers. *International Geophysics*, 81(PART B):1613–1614.
- Goldstein, P., Snoke, A. (2005). Sac Availability for the IRIS Community.
- Grechka, V. Y. and McMechan, G. A. (1995). Anisotropy and non-linear polarization of body waves in exponentially heterogeneous media. *Geophysical Journal International*, 123(3):959–965.
- Hall, L. C. (2013). *Seismic structure beneath the Wellington region from receiver functions*. Masters thesis, Wellington University of Wellington.
- Henry, S., Wech, a., Sutherland, R., Stern, T., Savage, M., Sato, H., Mochizuki, K., Iwasaki, T., Okaya, D., Seward, a., Tozer, B., Townend, J., Kurashimo, E., Iidaka, T., and Ishiyama, T. (2013). SAHKE geophysical transect reveals crustal and subduction zone structure at the southern Hikurangi margin, New Zealand. *Geochemistry, Geophysics, Geosystems*, 14(7):2063–2083.
- Hirschberg, H., Lamb, S. S., and Savage, M. (2015). Gravitational Stress Calculation for New Zealand. Technical Report November 2014, Victoria University of Wellington, Wellington.

- Johnson, J. H., Savage, M. K., and Townend, J. (2011). Distinguishing between stress-induced and structural anisotropy at Mount Ruapehu volcano, New Zealand. *Journal of Geophysical Research: Solid Earth*, 116(12):1–18.
- Karalliyadda, S. (2014). *S-wave splitting in the transpressional zone of New Zealand plate-boundary : implications for deformation and dynamics*. Doctoral thesis, Victoria University of Wellington.
- Küperkoch, L., Meier, T., Br, A., Lee, J., and Friederich, W. (2012). autoregressive prediction : application to local and regional distances. *Geophysical Journal International*, 188(2):687–702.
- Kusznir, N. J., Vita-Finzi, C., Whitmarsh, R. B., England, P., Bott, M. H. P., Govers, R., Cartwright, J., and Murrell, S. (1991). The Distribution of Stress with Depth in the Lithosphere: Thermo-Rheological and Geodynamic Constraints [and Discussion]. *Philosophical Transactions of the Royal Society of London A: Mathematical, Physical and Engineering Sciences*, 337(1645):95–110.
- Lamb, S. (2015). Kinematics to dynamics in the New Zealand Plate boundary zone : implications for the strength of the lithosphere. *Geophysical Journal International*, 201:552–573.
- Lamb, S. H. and Vella, P. (1987). The last million years of deformation in part of the New Zealand plate-boundary zone. *Journal of Structural Geology*, 9(7):877–891.
- Lay, T., Kanamori, H., Ammon, C. J., Nettles, M., Ward, S. N., Aster, R. C., Beck, S. L., Bilek, S. L., Brudzinski, M. R., Butler, R., and Deshon, H. R. (2005). The Great Sumatra-Andaman Earthquake of 26 December 2004. *Science*, 308(5725):1127–1134.
- Little, T. A., Grapes, R., and Berger, G. W. (1998). Late Quaternary strike slip on the eastern part of the Awatere fault, South Island, New Zealand. *Geological Society of America Bulletin*, 110(2):127–148.

- Margheriti, L. (1998). SKS-wave splitting and upper mantle structure: results in the Southern Apennines. *Annali Di Geofisica*, 41(4):607–616.
- Marson-Pidgeon, K. and Savage, M. K. (1997). Frequency-dependent anisotropy in Wellington, New Zealand. *Geophysical Research Letters*, 24(24):3297–3300.
- Nicol, A. and Beavan, J. (2003). Shortening of an overriding plate and its implications for slip on a subduction thrust , central Hikurangi Margin , New Zealand. *Tectonics*, 22(6):1–14.
- Nur, A. (1971). Effects of Stress on Velocity Anisotropy. *Journal of Applied Geophysics*, 76(8):2022–2034.
- Reyners, M. and Eberhart-Phillips, D. (2009). Small earthquakes provide insight into plate coupling and fluid distribution in the Hikurangi subduction zone, New Zealand. *Earth and Planetary Science Letters*, 282(1-4):299–305.
- Russo, R. M., Silver, P. G., Franke, M., Ambeg, W. B., and James, D. E. (1996). Shear-wave splitting in northeast Venezuela , Trinidad , and the eastern Caribbean. *Physics of the Earth and Planetary Interiors*, 95(3-4):251–275.
- Savage, M. K. (1999). Seismic anisotropy and mantle deformation: What have we learned from shear wave splitting? *Reviews of Geophysics*, 37(1):65–106.
- Savage, M. K., Wessel, a., Teanby, N. a., and Hurst, a. W. (2010). Automatic measurement of shear wave splitting and applications to time varying anisotropy at Mount Ruapehu volcano, New Zealand. *Journal of Geophysical Research*, 115(B12):B12321.
- Serway, R. and Jewett, J. (2007). *Physics for Scientists and Engineers with Modern Physics*. Brooks Cole, 7 edition.
- Seward, A., Henderson, M., Stern, T., Mochizuki, K., Iwasaki, T., Bassett, D., Henrys, S., Sutherland, R., Savage, M., Sato, H., and Kennedy, C. (2010). Seismic Array HiKurangi Experiment (SAHKE) Field Acquisition Report. Technical Report 48, GNS Science.

- Silver, P. and Chan, W. (1991). Shear Wave Splitting and Sub continental Mantle Deformation. *Journal of Geophysical Research: Solid ...*, 96:16,429–16,454.
- Stachnik, J. C., Sheehan, a. F., Zietlow, D. W., Yang, Z., Collins, J., and Ferris, a. (2012). Determination of New Zealand Ocean Bottom Seismometer Orientation via Rayleigh-Wave Polarization. *Seismological Research Letters*, 83(4):704–713.
- Stein, S. and Wysession, M. (1991). *An Introduction to Seismology, Earthquakes, and Earth Structure*. Wiley.
- Stern, T. A., Quinlan, G. M., and Holt, W. E. (1992). Basin formation behind an active subduction zone : three-dimensional flexural modelling of Wanganui Basin , New Zealand. *Basin Research*, 4(3-4):197–214.
- Teanby, N. (2006). Shear Wave Splitting Programs Version 1.0. Technical report.
- Teanby, N. A., Kendall, J., and Baan, M. V. D. (2004). Automation of Shear-Wave Splitting Measurements using Cluster Analysis. 94(2):453–463.
- Telford, M. W., Geldart, L. P., and Sheriff, R. E. (1990). *Applied Geophysics*. 2nd edition.
- Townend, J., Sherburn, S., Arnold, R., Boese, C., and Woods, L. (2012). Three-dimensional variations in present-day tectonic stress along the Australia-Pacific plate boundary in New Zealand. *Earth and Planetary Science Letters*, 353-354:47–59.
- Walcott, R. I. (1978). Present tectonics and Late Cenozoic evolution of New Zealand. *Geophysical Journal of the Royal Astronomical Society*, 52(1):137–164.
- Wallace, L. M., Barnes, P., Beavan, J., Dissen, R. V., Litchfield, N., Mountjoy, J., Langridge, R., Lamarche, G., and Pondard, N. (2012). The kinematics of a transition from subduction to strike-slip : An example from the central New Zealand plate boundary. *Journal of Geophysical Research B: Solid Earth*, 117(B2).

- Wallace, L. M. and Beavan, J. (2010). Diverse slow slip behavior at the Hikurangi subduction margin, New Zealand. *Journal of Geophysical Research: Solid Earth*, 115(12):1–20.
- Wallace, L. M., Beavan, J., McCaffrey, R., and Darby, D. (2004). Subduction zone coupling and tectonic block rotations in the North Island, New Zealand.
- Wallace, L. M., Reyners, M., Cochran, U., Bannister, S., Barnes, P. M., Ellis, S., Nicol, A., McCaffrey, R., Townend, J., Robinson, R., Bell, R., Wilson, K., and Power, W. (2009). Characterizing the seismogenic zone of a major plate boundary subduction thrust: Hikurangi Margin, New Zealand. *Geochemistry, Geophysics, Geosystems*, 10(10).
- Williams, C. a., Eberhart-Phillips, D., Bannister, S., Barker, D. H. N., Henrys, S., Reyners, M., and Sutherland, R. (2013). Revised Interface Geometry for the Hikurangi Subduction Zone, New Zealand. *Seismological Research Letters*, 84(6):1066–1073.
- Zha, Y., Webb, S. C., and Menke, W. (2013). Determining the orientations of ocean bottom seismometers using ambient noise correlation. *Geophysical Research Letters*, 40(14):3585–3590.
- Zinke, M. Z. J. (2000). Structure-related and stress-induced shear-wave velocity anisotropy: observations from microearthquakes near the Calaveras Fault in central California. *Bulletin of the Seismological Society of America*, 90(5):1302–1312.
- Zoback, M. L. (1992). First- and Second-Order Patterns of Stress in the Lithosphere ' The World Stress Map Project. *Journal of Geophysical Research: Solid Earth*, 97(B8):11703–11728.