Velocity structure of the Whataroa Valley using Ambient Noise Tomography

By

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Abstract

This thesis applies ambient noise tomography to investigate the shallow structure of the Whataroa Valley. Ambient noise techniques are applied to continuous seismic recordings acquired on 158 geophones deployed during the Whataroa Active Source Seismic Experiment. Despite only having four days of data, a robust shear-wave velocity model is calculated using a phase-weighted stacking approach to improve the cross-correlation functions' signal-to-noise ratios, allowing for robust velocity measurements to be obtained between periods of 0.3 and 1.8 s. This yields a database of 12,500 vertical component cross correlation functions and the corresponding Rayleigh wave phase and group velocity dispersion curves. Linearised straight-ray tomography is applied to phase and group velocity dispersion measurements at periods ranging from periods of 0.3 to 1.8 s. The tomography reveals a velocity that decreases from the vicinity of the DFDP-2B borehole to the centre of the valley. This is interpreted to be the geologic basement deepening towards the centre of the valley. A Monte-Carlo inversion technique is used to jointly invert Rayleigh-wave phase and group velocity dispersion curves constructed from phase and group velocity tomography maps of successively higher periods. Linear interpolation of the resulting 1D shear-wave velocity profiles produces a pseudo-3D velocity model of the uppermost 1,000 m of the Whataroa Valley. Using sharp increases in velocity to represent lithological change, we interpret two velocity contours at 1,150 and 1,250 m/s as potential sediment-basement contacts. Depth isocontours of these velocities reveal that the basement deepens towards the centre of the valley, reaching a maximum depth of 400 or 600 m for the 1,150 and 1,250 m/s velocity contours respectively. These depths indicate strong glacial over-deepening and have implications for future drilling projects in the Whataroa Valley. A sharp velocity increase

of 200 m/s also occurs at 100 m depth at the DFDP-2B borehole. We interpret this to be a change in sedimentary rock lithology from fluvial gravels to lacustrine silty sands, related to a change in sedimentary depositional environment.

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1 Introduction

1.1 Motivation and Thesis Aims

The Alpine Fault is a major continental transform fault that runs the length of the South Island, New Zealand. Large earthquakes (M_w 7.6-8.2) occur along the fault approximately every 329 ± 68 years (Berryman et al., 2012) with the last event occurring in 1717 (Sutherland et al., 2006). The fault is late in its seismogenic cycle, with a 27% probability of a large earthquake occurring over the next 50 years (Biasi et al., 2015). A unique opportunity exists to study the processes of a fault capable of producing a large earthquake before it ruptures (Townend et al., 2009).

Located on the central Alpine Fault, the Whataroa valley is an ideal location to study fault zone structure and processes as it provides access to the hanging wall at a range of distances from the fault. The site has hosted multiple scientific studies aimed at further understanding the seismogenic processes of the Alpine Fault and its fault mechanics (Townend et al., 2009). In January 2016, as part of the Deep Fault Drilling Project (DFDP), the Whataroa Active Source Seismic Experiment deployed a suite of downhole and surface instruments to image the Alpine Fault in the Whataroa Valley. During the deployment, 158 3-component geophones were continuously recording from the 17th to 21st of January.

Previous seismic studies aimed at characterising the Alpine Fault in the vicinity of the Whataroa Valley have mostly used reflection and refraction methods and have encountered a strong scattering of seismic waves and out-of-plane reflectors (Davey, 2010). The use of passive and tomographic methods to construct a velocity model

has shown promise in areas where it is difficult to use traditional methods (Bren-
guier et al., 2007). Lay et al. (2016) successfully used first arrival compressional-wave
travel-time tomography to characterise the Alpine Fault in the Whataroa Valley. In
this thesis we extend velocity characterisation of the Whataroa Valley using surface
wave tomography, performed using continuous recordings of seismic ambient noise.

This thesis aims to answer the following questions:

1. Can we make robust velocity models be made from from four days of contin-
uous seismic ambient noise recordings using ambient noise tomography?
2. What is the Shear-wave velocity structure of the Whataroa Valley?
3. How deep is the geologic basement in the Whataroa Valley?

1.2 Geologic Setting

New Zealand sits astride the boundary of the Pacific and Australian tectonic plates.
Interaction between these two plates has created a geologically diverse landscape of
natural processes such as volcanism, earthquakes and mountain building. The plate
boundary within New Zealand can be split into three main sections, the Hikurangi
Margin, Alpine Fault and Puysegur Trough [Figure 1.1]. The Hikurangi Margin is
the southern continuation of the Tonga-Kermadec subduction zone and is formed
by the Pacific Plate subducting westward beneath the Australian Plate. Inter-plate
motion transitions through the northern South Island Marlborough Fault System
and into the Alpine Fault, an oblique dextral transform fault along the South Island's
western coast (Sutherland et al., 2006). Near the bottom of the South Island, the
plate boundary motion has the opposite polarity to the Hikurangi Margin, creating
the Puysegur Trough off the coast of the South Island (Berryman et al., 1992).

Figure 1.1: Map of New Zealand showing the approximate plate boundary between the Pacific and Australian Plates. Relative motion between the two plates is represented by the black vectors. HT refers to the Hikurangi Margin, AF refers to the Alpine Fault and PT refers to the Puysegur Trough, all referenced in the text. Whataroa Valley, the study area, is marked by the red star. The arrows represent relative plate motion between the Pacific and Australian Plate (Altamimi et al., 2012). The black lines are surface fault traces from the New Zealand active fault database (Langridge et al., 2016).

1.2.1 Alpine Fault

The Alpine Fault was first identified from its topographic expression (Wellman and Willett, 1942); as a linear feature extending along the Western edge of the Southern Alps. Presently, the fault has an average strike of 055° (DeMets et al., 1994), and accommodates 75 % of intra-plate motion and 100 % compression in the central part of the fault (Norris and Cooper, 2001). Over the last 23 Ma a total of 460 km of dextral offset has occurred, recorded by offset of the Dun Mountain Ophiolite Belt (Sutherland et al., 2000). Reconstructions of plate motion show a total offset of 850 km between the western North Island and eastern South Island (Molnar et al., 1999). The 390 km difference in offset may be taken up by shear on other structures (Norris and Cooper, 2001). Vertical motion of the Alpine Fault was first recognised in regional offset of terrace surfaces and the long-term formation of the Southern Alps (Wellman, 1953, Wellman et al., 1952). Total uplift of the Southern Alps is estimated to be between 9 and 22 km (Cooper, 1980).

The dip character of the Alpine Fault changes down its length. The northern section dips steeply to the east (Cox and Sutherland, 2007), the central section dips $45\text{--}60^{\circ}$ to the east (Stern et al., 2007) and the southern section is nearly vertical (Sutherland et al., 2007). While fault motion is oblique for most of its length, some sections are considered dextral strike-slip and reverse slip-dip (Norris and Cooper, 2001, Sibson et al., 1979). Average horizontal motion of the fault is estimated to be 21–27 mm per year, with the southern most section having an estimated 31 mm per year (Sutherland et al., 2006).

New Zealand's basement geology is made up of a series of terranes and igneous suites (Cox and Sutherland, 2007). The basement terranes are grouped into Eastern and Western Provinces (Mortimer, 2004). The Western Province forms the foot-wall of the present day Alpine Fault and contacts the Eastern Province at the Alpine Fault (Mortimer, 2004). The Western Province is comprised of quartzose

sediments that have undergone metamorphism from plutonic intrusions (Cox and Sutherland, 2007). Along the West Coast of the South Island, the Western Province is overlain by 2000-4000 m of Cenozoic sediments (Nathan et al., 1986). Rocks of the Eastern Province are dominantly lithic and feldspathic greywackes, at differing levels of metamorphism (Cox and Sutherland, 2007). Immediately adjacent to the Alpine Fault is the Alpine Schist which formed during collision and metamorphism of the Caples and Rakaia Terranes (Mortimer, 2004). The Alpine Schist grades into proto-mylonite, mylonite, ultramylonite and finally cataclasite at the Alpine Fault Damage zone (Norris and Cooper, 2007).

1.2.2 Whataroa

Geologic mapping has revealed that the Alpine Fault changes geometry in the vicinity of the Whataroa Valley (Norris and Cooper, 2007). To the south of Whataroa, the fault is broken up into north-striking thrust zones and easterly striking, sub-vertical strike-slip zones (Norris and Cooper, 1995). North of Whataroa, the Alpine Fault is mostly a straight trace, with an oblique-dextral reverse slip sense of motion (Norris and Cooper, 2007). In the Whataroa Valley region, vertical displacement rates of 8-12 mm/yr are estimated and while horizontal displacement rates are estimated to be 22-27 mm/yr, (Herman and Braun, 2006) suggest in 100-200 m of uplift since the last glaciation (14 ka) (Herman and Braun, 2006). West of the Alpine Fault in the Whataroa region, basement rocks are overlain by a 2.5-3 km thick sedimentary sequence (Davey, 2010).

The Whataroa Valley is a glacial carved U-shaped valley characterised by river flats and terraces that record past earthquakes of the Alpine Fault (Herman and Braun, 2006). A mix of fluvioglacial and alluvial deposits characterise the surface geology of the valley floor [Figure 1.2]. The valley walls show a typical Alpine Fault hanging wall rock sequence where schist grades into mylonite towards the Alpine Fault [Figure 1.2]. The DFDP-2B borehole encountered a 240 m sedimentary sequence of

glaciolacustrine deposits overlain by deltaic sands and post-glacial alluvium (Sutherland et al., 2015). The thicker than expected sedimentary sequence suggests glacial over-deepening of the valley and rapid sedimentation since the last glacial cycle (Sutherland et al., 2015).

Figure 1.2: Geology map of Whataroa Valley and surrounding area. The study area is marked by the black box [Projection: NZTM].

1.3 Previous Studies

The Alpine Fault has been the subject of a broad range of research due to its importance as a seismic hazard and international relevance as a large fault exposed on-land. This section will only outline studies relevant to this thesis.

Deep Fault Drilling Project

The Deep Fault Drilling Project (DFDP) is a multicomponent project investigating the Alpine Fault late in its seismic cycle. The main objective of the project has been a series of boreholes aimed at drilling through the Alpine Fault to make observations in the hanging wall, principle slip zone and foot wall (Townend et al., 2009). Two boreholes were drilled at Gaunt Creek, 8 km southwest of the Whataroa Valley [Figure 1.2], reaching a vertical depth of 101 m and 151 m (DFDP-1A and DFDP-1B respectively). The principle slip zone of the Alpine Fault was recovered at 128 m in DFDP-1B (Townend et al., 2009). In mid 2014, a second phase of drilling began that aimed to intersect the principle slip zone of the Alpine Fault at appropriately 1100 m depth. DFDP-2 drilled two boreholes to a vertical depth of 211 m and 813 m (DFDP-2A and DFDP-2B respectively). Strongly foliated schist resulted in DFDP-2B deviating North, giving a total drilled depth of 893 m (Sutherland et al., 2015). Hostile drilling conditions including a hydrothermal gradient of $125 \pm 55^\circ\text{C}/\text{km}$ (Sutherland et al., 2017), a deeper than expected sedimentary sequence and highly foliated schist culminated in broken casing, leading to the well being abandoned before it reached its principle target (Sutherland et al., 2015). Despite not reaching its target, valuable geologic and geophysical observations were made, including higher than expected temperatures. Wireline logging was undertaken including the installation of a optical fibre cable outside the borehole casing to 893 m (Sutherland et al., 2015).

Active Source Seismic

Multiple active source seismic studies have been conducted in and around Whataroa Valley. The 1996 South Island Geophysical Transect (SIGHT) was a large 2D seismic line across the South Island that went through the Whataroa Valley. Offshore and onshore seismic lines were deployed with the aim of investigating the styles of lithospheric deformation associated with continent-continent collision at a major plate

boundary (Davey et al., 1998, Van Avendonk et al., 2004). In 1998, Whataroa98 was completed as an extension of the 1996 SIGHT survey, with the aim of imaging the Alpine Fault and crustal structures along the Whataroa Valley and across the coastal plain (Davey, 2010). Using reflection and refraction data, 300-500 m thick sedimentary sequences overlying basement were imaged in the Whataroa Valley and across the Alpine Fault, before deepening sharply 6 km North-west of the fault (Davey, 2010). The basement and sedimentary sequence contact was tied to the Harihari-1 and Waiho-1 exploration wells, dating it at Early Miocene (Davey, 2010).

In preparation for the Deep Fault Drilling Project, the WhataDUSIE survey acquired a 5 km 2D seismic line down the Whataroa Valley and a small 3D survey at the valley mouth with the aim of imaging the upper few kilometres with borehole site characterisation in mind. Lay et al. (2016) used the 2D line from WhataDUSIE and Whataroa98 to produce reflection and velocity models of the upper 5 km. They observe a strong reflector at depths of 1.5 to 2.2 km with an apparent dip of 48° interpreted as the Alpine Fault. Parallel reflectors above the main reflector are interpreted as a 600 m wide damage zone. Velocity modelling shows sediments overlying the basement to be 300-600 m thick with highly variable P-wave velocity of 600-2300 m/s. Basement P-wave velocity ranges from 4500 to 5500 m/s with a low velocity zone extending from 800 to 2000 m within the basement (Lay et al., 2016).

Lepine (2016), a Master of Science student from the University of Otago, conducted a hammer drop and weight drop experiment, producing several 2D reflection profiles that image the sediment overlying the basement in the Whataroa Valley. Weak reflectors are observed at 250 m near DFDP-2B and are interpreted to be basement (Lepine, 2016).

Gravity

Several regional gravity studies have been conducted in the Whataroa area, mostly with the aim of characterising the Alpine Fault footwall. Brikke (2007) combined gravity and seismic measurements to model basement depth of the Whataroa Plains. An average depth to basement in the Whataroa Valley was constrained to appropriately 300 m (Brikke, 2007). Davy et al. (2013) added a further 361 stations to the area surveyed by Brikke (2007) and identified large glacial channels previously identified by Brikke (2007) as a large negative gravity anomaly. Jenkins (2017) added a further 446 high precision gravity measurements in Whataroa Valley and the floodplain. A steep residual Bouguer gravity gradient near the DFDP-2B borehole is suggested to be a moderately dipping reverse fault or sub-vertical erosional wall. A negative Bouguer gravity anomaly south-east of DFDP-2B was interpreted as glacial lake deposits (Jenkins, 2017).

1.4 Dataset

Data used in this project are from the three-component geophone array deployed during the 2016 Whataroa Active Source Seismic Experiment [Figure 1.3]. This section introduces the dataset and outlines the Whataroa Active Source Seismic Experiment.

In January 2016, the Whataroa Active Source Seismic Experiment was conducted in the Whataroa Valley, Westland with the aim of imaging the hanging wall of the Alpine Fault. The experiment was the latest phase of the Deep Fault Drilling Project and utilised the optical fibre installed down to 893 m and open casing to 400 m. The experiment aimed to:

1. Directly image the Alpine Fault in order to determine its geometry and seismic characteristics and thus inform the targeting and technical design of future

Figure 1.3: Map of the Whataroa Valley and instruments used in this thesis. The orange triangles mark the geophone locations. The yellow circle marks the DFDP-2B borehole and location of the recorder and base operations of the active source survey. The Whataroa River is marked in blue, roads are marked as white lines and the Alpine Fault surface trace is marked as black lines.

Deep Fault Drilling Project drilling activities,

2. Test new optical fibre Distributed Acoustic Sensing methods in the borehole and compare results to those obtained using traditional down-hole geophones and seismometers, for both active source and earthquake seismology,
3. Determine the acoustic and elastic properties of the hanging-wall rocks in the vicinity of the Alpine Fault for comparison with geophysical logs collected in the well

An extensive pool of geophysical equipment was used in the experiment, which can be split into three categories:

1. Surface instruments - The main surface instruments were 158 3-component geophones with a GFZ DSS cube3 4.5 Hz recorder and a ARIES Multichannel Reflection System of 1-component geophones. An array of Reftek 3-component seismometers was also deployed.
2. Borehole instruments - Heterodyne Distributed Vibration Sensing (hDVS) system was deployed to 893 m and a SERCEL downhole Vertical Seismic Profiling (VSP) system was deployed to 400 m.
3. Sources - The primary energy source was a IVI EnviroVibe Minivibe supplied by the University of Calgary that provided 16 s sweeps from 10 to 200 Hz. A trailer mounted 250 kg weight drop was used to orientate the VSP.

The sparse geophone array was active from 16th to 21st January, 2016, and is used in this thesis. Due to field logistics, geophones were not correctly orientated to North, so this project only uses vertical component data as calculating orientations is beyond the scope of the project. The raw data were extracted from the cubes by staff from the University of Freiburg in a binary format, with each file containing p0, p1, and p2 channels, corresponding to EHZ, EHN and EHE. Gipptools, a program provided by the geophone supplier, was used to convert the files into miniSEED format. Obspy Python libraries were then used to archive each file into day long continuous

waveforms with the Buffer of Uniform Data (BUD) structure (Beyreuther et al., 2010).

1.5 Ambient Noise in Seismology

The benefits of using surface waves to study the Earth have long been recognised (e.g. Stein and Wyssession, 2009). However, until recently, using surface waves extracted from the ambient noise field was an under utilised method (e.g. Stein and Wyssession, 2009). Inspired by the work of Aki (1957) and cross-correlation of random wave-fields in acoustics (Lobkis and Weaver, 2001, Weaver and Lobkis, 2004), Shapiro and Campillo (2004) showed ambient noise can be cross-correlated to extract the Green's Function, and thus calculate surface wave dispersion. The use of ambient noise as a source overcomes a major spatial limitation of using earthquakes, which as a general rule mostly occur near plate boundaries and rifts. Increasingly, ambient noise is also used as non-intrusive complementary method to active source studies at local scales (e.g. De Ridder and Biondi, 2013, Hannemann et al., 2014, Lin et al., 2013). Previous studies using ambient noise tomography on a similar target frequency range have used longer recordings of ambient noise. Lin et al. (2013) used three weeks of recordings to make measurements between 0.5 and 4 Hz, and Hannemann et al. (2014) used two weeks and frequency range of 1.6 to 9 Hz.

1.5.1 Surface-wave Propagation and Nature of the Ambient Noise Wave-field

Seismic surface waves propagate along earth's surface, decaying over a two-dimensional surface, while body waves decay over a three-dimensional space, making surface waves the dominant phase in most earthquake seismographs (Stein and Wyssession, 2009). Rayleigh and Love waves are the two main types of surface waves. Rayleigh waves have similar particle motion to compressional-waves (P-waves) and vertically

polarised shear-waves (SV-waves), with motion on the vertical and radial components of a seismograph (Stein and Wysession, 2009). Love waves have similar particle motion to horizontally polarised shear-waves (SH-waves), with particle motion on the transverse component of a seismograph (Stein and Wysession, 2009). Surface waves are also dispersive, meaning different frequencies travel at different velocities. Surface waves also travel at two speeds due to their dispersive nature; the single wave velocity is called phase velocity and the wave envelope velocity is called the group velocity which is the derivative of the phase velocity (Stein and Wysession, 2009).

The dispersive nature of surface waves and their sensitivity to different depth underpins their usefulness. Different periods sample the extent of different depths, allowing a dispersion curve to be constructed that can then be inverted into a profile of shear-wave velocity with depth (Stein and Wysession, 2009).

Rayleigh and Love waves are commonly split into three different period bands; >1 , $1-0.2$ and <0.2 s (Bonnefoy Claudet et al., 2006). The period band above 1 s is produced by natural sources such as changes in air pressure or by ocean waves impinging on coast lines. Between 0.2 and 1 s, noise can be either natural or cultural from sources such as rivers, while below 0.2 s it is usually cultural noise such as road traffic, human footsteps or machinery (Bonnefoy Claudet et al., 2006).

1.5.2 Reconstructing the Green's Function

The extraction of surface wave information from the ambient noise wavefield was first demonstrated by Aki (1957), who showed the theoretical relationship between diffuse surface waves and ambient noise cross-spectra. Using ultrasonic acoustics, Lobkis and Weaver (2001) showed that cross-correlating diffuse wave-fields recorded at different stations can reconstruct the Green's Function, the response at each station to an impulse at the other. They then expanded the theory to open systems

such as the Earth (Weaver and Lobkis, 2004).

The work by Lobkis and Weaver (2001) set the framework for a resurgence in interest in the use of surface waves as a tool for imaging the earth. Campillo and Paul (2003) used earthquake coda wave cross-correlations to extract the surface wave component of the approximated Green's Function between two stations. They showed that all nine components of the Green's tensor were estimated by comparing synthetic tensors to real data. Shapiro and Campillo (2004) were the first to cross-correlate seismic ambient noise waveforms to retrieve the approximated Green's function. They then showed fundamental mode Rayleigh wave group velocity dispersion curves can be calculated from Cross-Correlation Functions (CCF). Using the same method, Shapiro et al. (2005) used Rayleigh wave group velocity dispersion curves to construct velocity maps of California at 7.5 and 15 s period. In doing this, they revealed the potential of using ambient noise to image the subsurface by correlating their surface wave velocity maps with well known geological features.

The CCFs from a noise source distribution, representing the ambient noise field, can be approximated using the equation derived by Yao and Van Der Hilst (2009). The cross-correlation C_{xy} of the wave-field recorded at station x and y, from sources emitting harmonic wave pulses of amplitude ρ and angular frequency ω , is

$$C_{xy}(\Delta t, \omega, W) = \int W(\Delta t_d) \rho(\Delta t_d, \omega) \cos[\omega(\Delta t - \Delta t_d)] d\Delta t_d \quad (1.1)$$

where Δt is the lag-time, Δt_d is the travel-time difference of wave phases travelling to stations x and y, and W is a windowing function describing the shape of the phases (Yao and Van Der Hilst, 2009). Figure 1.5 show examples of equation 1.1 using different source distributions with each source emitting a Gaussian-shaped phase of amplitude 1, period of 5 s and a travel-time of 10 s. The two stations are located 300 km apart. For a single source, the CCF signal corresponds to a correlation gather of source-position angle [see Snieder and Wapenaar (2010); Figure 1.5a].

Integrating over sources from all directions obtains two symmetric signals, representing a wave travelling from station x to y (virtual source to virtual receiver) at positive lag-time and y to x at negative lag-time at 3000 m/s [Figure 1.5c]. This illustrates the stationary phase argument of Snieder (2004) where signals originating from outside of the stationary phase zone destructively interfere while signals from a source within the stationary phase zone will constructively interfere [Figure 1.4].

Figure 1.4: Schematic showing stationary phase zone for stations A and B where signal from S' will constructively interfere and signal from S will destructively interfere. The grey area marks the stationary phase zone. Figure taken from Snieder and Larose (2013).

Asymmetric CCF are a common occurrence when cross-correlating seismic ambient noise, where sources have an uneven distribution, such as near the ocean (e.g. Lin et al., 2013). Wapenaar (2006) showed that it is possible to retrieve the approximated Green's Function from a one-sided distribution of noise sources. In the case of a one-sided source distribution, the approximated Green's Function is only retrieved from one side of the CCF [Figure 1.5b]. Ambient noise cross-correlations can also have symmetric signal in terms of lag-time, but with differing amplitude (e.g. Lin et al., 2013). This is shown in Figure 1.5d, where the number of sources for an acausal signal is five times the number of sources for the causal signal. The strength of signal in this situation will also control asymmetry of the CCF amplitude.

1.6 Thesis Structure

This section outlines the thesis structure by briefly describing the contents of each chapter. This thesis is written in the context of 'we' rather than 'I', reflecting the amount of time my supervisors and others have contributed to this thesis.

This thesis is structured as follows:

- **Chapter 1:** An introduction to the motivation and objectives of this study. The geologic setting of New Zealand and Alpine Fault is introduced before we describe the geology of the Whataroa Valley that is relevant to this study. The dataset used in this thesis was acquired during the Whataroa Active Source Seismic Experiment so we introduce the experiment and the dataset together. The underlying theory of turning noise into signal is described, along with examples of ideal and non-ideal noise source distributions and resulting CCF.
- **Chapter 2:** We outline methods used to turn ambient noise recordings into a CCF representing an approximated Green's Function and subsequent dispersion measurements that can be calculated from CCF. Inversion techniques used to obtain surface-wave and shear-wave velocity models are introduced. We also introduce techniques used for error analysis.
- **Chapter 3:** Results calculated using the methods outlined in Chapter 2 are presented in this chapter. We present our observations from each result, including CCFs, back-projection for persistent localised source location, dispersion curves, surface-wave velocity maps and shear-wave velocity models.
- **Chapter 4:** We assess our results by first discussing the errors and considerations of each key finding. Our velocity models are then interpreted to answer the aims of this thesis stated at the start of this chapter.
- **Chapter 5:** Concluding remarks are stated in this chapter, including key findings and recommendations for future work.

Figure 1.5: Noise source distributions and the summed Cross-Correlation Function. The left panels show the locations of noise source (red circles) signal recorded at stations x and y (yellow triangles) and then cross-correlated. The right panels show the sum of all cross-correlation functions from the corresponding source configuration on the left (after Behr, 2011).

2 Methods

The methods used to calculate the velocity structure of the Whataroa Valley are introduced in this chapter. We first discuss processing techniques used to obtain surface wave dispersion curves from cross-correlation of seismic ambient noise recordings, followed by inversion procedures to obtain surface-wave velocity maps and shear-wave depth profiles.

2.1 Measuring Surface Wave Dispersion from Ambient Noise

To calculate reliable surface wave velocities from ambient noise recordings, several processing steps must be undertaken: waveform preparation, cross-correlation, measuring surface wave dispersion curves and quality control [Figure 2.1]. Before waveforms are cross-correlated, they must be processed using signal processing routines with the aim of obtaining a set of ambient noise waveforms that best represent a stable noise-field in both space and time (Roux, 2009). Due to the short recording time of the dataset, we test techniques that differ from conventional processing of ambient noise recordings, namely phase cross-correlation and Phase-Weighted Stack (PWS).

We use a Python package, Monitoring Seismic Noise (MSNoise) (Lecocq et al., 2014), to process raw data into cross-correlations before using the dispersion curve package Alternate Frequency-Time Analysis (AFTAN) (<http://ciei.colorado.edu/Products/>, last accessed May 2017). The advantage of MSNoise is its modular design, allow-

ing for easy modification of codes and the ability to extract data part way through processing. It also has multi-thread processing allowing for more efficient computation, especially when performing cross-correlation, which can be a computationally expensive task.

2.1.1 Single Station Data Preparation

Waveforms are merged into day-long records and then demeaned, detapered and resampled to 20 Hz to reduce the file size. A bandpass filter with 0.01 and 8 Hz corners is applied before the day-long records are cut into shorter records and then a cosine taper is applied to both ends of the waveform to allow for smooth transition between the time and frequency domain [Figure 2.2]. We do not remove the instrument response because the geophones are all identical (Bensen et al., 2007). Standard procedure is to reject day data with less than 80 % of the day filled, however we have a limited recording time so this procedure is rejected during processing of CCF, but it is assessed during quality control (Bensen et al., 2007)[see Section 2.1.4].

We then apply time-domain normalisation to reduce bias in the cross-correlation from earthquakes, stationary noise sources and instrument spikes (Yao and Van Der Hilst, 2009). Bensen et al. (2007) outlined five methods for time-domain normalisation; one-bit normalisation (e.g. Larose et al., 2004, Yao et al., 2006); Root-Mean-Square (RMS) clipping (Sabra et al., 2005); automated event detection; absolute mean normalisation; and water-level normalisation. The methods are data dependant and should be tested to obtain the best result (Bensen et al., 2007). Absolute mean normalisation and water-level normalisation are the preferred methods of Bensen et al. (2007), however they are computationally expensive and give similar results to the simpler one-bit normalisation. A disadvantage of one-bit normalisation is dominant frequency bands such as a high frequency localised source may over-print non-dominant frequency bands. This study uses noise from both cultural and natural sources making one-bit normalisation problematic as the presence of

Figure 2.1: Simplified processing flow for measuring surface wave dispersion after Bensen et al. (2007).

high frequency noise from local sources may over print useful signal in the 1 Hz range. Instead we clip the amplitudes by three times the Root Mean Square (RMS) of the amplitude and find satisfactory results [Figure 2.2].

The Ambient noise wave-field contains an uneven distribution of spectral energy (Bensen et al., 2007). Over a broadband scale, three distinctive peaks occur; the first at 7.5s, the second at 15s, and a third peak at 50s referred to as the Earth hum (Rhie and Romanowicz, 2004). In this project we are concerned with periods below 2s, so any peaks in spectral energy are likely from localised cultural and natural sources. Spectral whitening equalises energy across a specified frequency band, ensuring the same spectral energy between the whitened frequency bounds (Stehly et al., 2009). To whiten, the RMS clipped waveform is transformed into the frequency domain where its spectral energy is whitened between 0.1 and 5 Hz (10 and 0.2s period) [Figure 2.2]. We limit the upper corner of the frequency band-width during whitening to 5 Hz to remove signal from the Vibroseis. The lower corner frequency of 0.1 Hz is intentionally left smaller than reasonable when considering rational wavelengths so we can explore the lower frequency signals that may be present in CCF of longer inter-station distances.

2.1.2 Calculating Cross-Correlations

Efforts to produce the best representation of the Green's Function usually focus on waveform preparation (Bensen et al., 2007). But due to the short recording time of the geophone dataset, we also test alternate cross-correlation and stacking methods to try to improve our Cross-Correlation Functions (CCF).

Cross-Correlation

We compute cross-correlations using two methods: standard cross-correlation and phase cross-correlation (Schimmel, 1999). Standard cross-correlation measures the coherence between the amplitude of the phase using the equation,

Figure 2.2: Signal processing routines applied to ambient noise waveforms prepare the traces for cross-correlation. Left panel shows 120 s long waveforms in the time domain at a) pre RMS clipping, b) pre-whitening and c) post whitening. Right panel shows the waveforms in the frequency domain at a) pre RMS clipping, b) pre-whitening and c) post whitening. Pre-whitening spectral energy is mostly above 2 Hz; however once whitened there is an even energy distribution between the whitened bounds of 0.1 and 5 Hz.

$$c_{ccf}(t) = \sum_{\tau=\tau_0}^{\tau_0+T} u_1(t+\tau)u_2(\tau) \quad (2.1)$$

where u_1 and u_2 are the waveforms, T is time series length and t is the lag time. This type of cross-correlation is amplitude biased which is why so much processing is required to prepare the waveforms prior to cross-correlation (Bensen et al., 2007). To calculate cross-correlations using equation 2.1, we use $T = 120$ second and a lag-time of 40 s.

An alternative cross-correlation method is called the phase cross-correlation and it is un-biased by amplitude (Schimmel, 1999, Schimmel et al., 2011). It measures the coherence between the instantaneous phase of the waveforms, meaning no time domain normalisation or spectral whitening needs to be performed (Schimmel et al., 2011). Instead we apply a bandpass filter with 0.1 and 5 Hz corners to keep the frequency spectrum of both cross-correlations consistent. Phase cross-correlation uses analytic signal theory where the analytic signal $s(t)$ of a real time series $u(t)$ is defined as $s(t) = u(t) + iH[u(t)]$, where $H[u(t)]$ is the Hilbert transform and $s(t)$ the imaginary component in the complex domain. In exponential form $s(t) = a(t)\exp(i\phi(t))$, the envelope function $a(t)$ and instantaneous phase $\phi(t)$ contain neighbouring waveform information and a coherence measure respectively. The phase cross-correlation (c_{pcc}) is defined as

$$c_{pcc}(t) = \frac{1}{2T} \sum_{\tau=\tau_0}^{\tau_0+T} \left\{ |e^{i\phi(t+\tau)} + e^{i\psi(\tau)}|^v - |e^{i\phi(t+\tau)} - e^{i\psi(\tau)}|^v \right\} \quad (2.2)$$

where $\phi(\tau)$ and $\psi(\tau)$ are the instantaneous phases of time series u_1 and u_2 , as a function of lag-time t . The factor of $1/2T$ normalises the phase cross-correlation, where $c_{pcc} = 1$ is a perfect correlation and -1 an anti-correlation. To control the sensitivity, the power v is changed and generally leads to improved signal-to-noise ratio but decreased absolute correlation values (Schimmel et al., 2011).

We tested several v values and find that the default value of 1 is sufficient as the

use of phase-weighted stack gives us sufficient signal-to-noise ratio [see Stacking for explanation of phase-weighted stack]. Several time series lengths are tested using the phase correlation method and we found the phase CCF is almost identical when using time series lengths of 120-960 s. To increase computational efficiency, we opt to use 960 s time-series as calculating phase cross-correlations is computationally expensive (~ 300 s per job for ~ 53000 total jobs using 120 s time lengths versus ~ 80 s for 960 s time lengths).

We decide to use the standard cross-correlation method defined in equation 2.1 as the phase cross-correlation limits the range of inter-station distances that can be used for reliable dispersion measurements; see Section 4.1 for a discussion on the two techniques.

Stacking

Individual cross-correlations are then stacked into the Cross-Correlation Functions (CCF) representing each Julian day of data to improve signal-to-noise ratio (Bensen et al., 2007). Stacking improves the signal-to-noise ratio by calculating the mean of each point in the CCFs trace, with coherent signal that is similar through time 'adding', while non-coherent signal will cancel out [Figure 2.3]. The standard method for stacking is the linear stack where the amplitude of the wavelet is used and there is no weighting to any individual trace (Bensen et al., 2007). The linear stack ($stack_{lin}$) is defined as

$$stack_{lin}(t) = \frac{1}{N} \sum_{j=1}^N s_j(t) \quad (2.3)$$

where N is the number of signals s_j to be averaged. Recent studies have shown the phase-weighted stack gives a higher signal-to-noise ratio than the traditional linear stack (Schimmel and Paulssen, 1997, Thurber et al., 2014). The principle of the phase-weighted stack is to down-weight components of signals that do not share the

same instantaneous phase while up-weighting signals that do (Thurber et al., 2014).

Phase-weighted stacking first calculates the mean coherence between the instantaneous phase in all CCFs. The coherence is then used as a weighting factor to calculate the mean of each point on the trace. Phase-weighted stacking uses analytical signal theory in a similar fashion to the phase cross-correlation, where a CCF trace $s_j(t)$ and its Hilbert transform $H[s_j(t)]$ are combined to form the complex trace $S_j(t)$, defined as $S_j(t) = s_j(t) + iH[s_j(t)]$. $S_j(t)$ is expressed in exponential form as $S_j(t) = A_j(t)\exp(i\phi_j(t))$ where $A_j(t)$ is the amplitude and $i\phi_j(t)$ the instantaneous phase of the signal. Since no amplitudes are explicitly involved, for every t , the phase stack ($weight_{phase}$) is defined as

$$weight_{phase}(t) = \frac{1}{N} \sum_{j=1}^N e^{i\phi_j(t)} \quad (2.4)$$

where N is the number of traces. $weight_{phase}(t)$ is normalised to have an amplitude range between 0 and 1, with 1 meaning the signal is coherent and 0 is non-coherent. The phase stack is then used as a lag-time-dependant weighting factor for the linear stacking method ($stack_{lin}$) (Schimmel and Paulssen, 1997). This is performed by multiplying equations 2.3 and 2.4, so that together the phase-weighted stack ($stack_{pws}$) is defined as

$$stack_{pws}(t) = \frac{1}{N} \sum_{j=1}^N s_j(t) \left| \frac{1}{N} \sum_{j=1}^N e^{i\phi_j(t)} \right|^u \quad (2.5)$$

The power u controls the weight of $weight_{phase}$. We find a u value of 2.0 gives the best result as using a higher value results in unrealistic signal. The phase-weighted stack method is used as our stacking method in this thesis; see section 4.1 for discussion of the two methods.

Each daily CCF is then stacked into one stack using the phase-weighted stack method. We chose to remove Julian day 16 from our final stack as it only contains 4 hrs of data. Julian day 20 CCFs only contain ~60-80% of daily recording

Figure 2.3: Comparison of traces stacked using a linear and phase-weighted stack method. The plot marked single CCF is produced using equation 2.1. The left and right panels show the stacked CCF using the linear and phase-weighted stack methods. The plots marked full-stack show the final CCF using equation 2.3 and 2.5 respectively. The causal and acausal components are then averaged to produce a symmetric CCF resulting from linear and phase-weighted stack methods (marked as folded-stack in the plots).

time so we choose to also reject these daily CCFs from our final stack in accordance with the guidelines of Bensen et al. (2007). Using the principle of reciprocity, we then average causal and acausal components of final CCFs to reduce the effect of a heterogeneous noise-field (Shapiro and Campillo, 2004). The resulting CCFs are referred to as symmetric CCFs and are considered to be the final cross-correlation.

2.1.3 Calculating Surface Wave Dispersion Curves

We measure Rayleigh wave group and phase velocity by applying frequency-time analysis to all symmetric cross-correlations (Bensen et al., 2007, Levshin et al., 1972, 1992). The frequency-time analysis we use is called AFTAN (<http://ciei.colorado.edu/Products/> last accessed May 2017). Frequency-time analysis is an efficient method for mea-

suring dispersion from a high number of CFF due to its automatic nature (Bensen et al., 2007). To measure group velocity, the CCF is transformed into the frequency domain and then filtered with narrow-band Gaussian filters. For each filter, the function is transformed back into the time domain, resulting in a smooth envelope function and phase function. The group velocity travel-time for each filter is defined as the peak of the envelope function. Phase velocity is then measured using the method outlined by Lin et al. (2008), who showed that phase velocity (c) at period t_{max} can be calculated with the equation

$$c = \frac{\omega}{\kappa} = \frac{r\omega}{\phi(t_{max}) + \omega(t_{max}) - \frac{\pi}{4} - 2\pi N} \quad (2.6)$$

where ω is the angular frequency, κ the wave number, r the inter-station distance of the station pair, ϕ the phase of the filtered cross-correlation function and N the number of wave cycles at ω between the station pair. To calculate N , a previous dispersion curve is required to act as a reference curve (Lin et al., 2008). As we are only measuring vertical-vertical components we can keep the $-\frac{\pi}{4}$ phase ambiguity constant as shown by (Lin et al., 2008). The phase velocity reference curve used to calculate N is created using the software package Gpdc (<http://geopsy.org/wiki/index.php/Gpdc>, last accessed May 2017). Input into Gpdc requires layer thickness, shear and compressional-wave velocity, attenuation factor of compression and shear-wave velocity, and density (Figure 2.4). The parameters used in AFTAN are explained in Table 2.1.

Table 2.1: Parameters used in AFTAN to calculate dispersion measurements for group and phase velocity. The Gaussian filters centre on a range of periods between T_{min} and T_{max} . The window of the symmetric CCF is defined by velocity bounds V_{min} and V_{max} . We choose a SNR value of 2 to allow a full range of measurements, we then apply stricter criteria at the next stage of processing.

Parameters	Value
Factor to multiply phase ambiguity ($\frac{\pi}{4}$)	-1
Velocity min ($V_{min} - km/s$)	0.4
Velocity max ($V_{max} - km/s$)	2.0
Period min ($T_{min} - sec$)	0.2
Period max ($T_{max} - sec$)	2.0
Second derivative threshold for group velocity jump ($tresh$)	20.0
Factor of Gaussian filter width ($fact$)	0.1
Factor of left end taper ($taperl$)	1.0
SNR cut-off (SNR)	2
Factor to compress signal width ($fmatch$)	1.0

Figure 2.4: a) Compressional/shear-wave velocity model, and b) synthetic phase velocity dispersion curve used in AFTAN. The reference V_p/V_s model is calculated using the V_p model at the DFDP-2B borehole of Lay et al. (2016). To calculate V_s we use a ratio of 1.75. Density is calculated using Gardners relation $1.74 \times V_p^{0.25}$ (Gardner et al., 1974) and we assume a constant attenuation factor of 100 for both Q_p and Q_s .

We chose to omit phase-matched filtering from dispersion curve measurements because of the narrow frequency range measured. As mentioned in Bensen et al. (2007), short period signals are contaminated by prominent side-lobes when the Gaussian filtered signal is collapsed, contaminating the signal. For this reason Bensen et al. (2007) only advocate the use of phase match filters for broadband signals.

2.1.4 Quality Control

Error is not directly measured when calculating surface wave dispersion curves with AFTAN so selection criteria are applied to identify acceptable measurements. Bensen et al. (2007) advocate the use of signal-to-noise ratio, a minimum wavelength criteria and dispersion curve measurements of sub-stacks of the full stack.

Dispersion measurements with a signal to noise ratio below 15 are discarded, defined as the maximum amplitude of the signal divided by the RMS value of the noise (Bensen et al., 2007). The signal to noise ratio of 15 was chosen as the value that gave the best quality dispersion measurements whilst maintaining an acceptable number of measurements for further analysis e.g. path coverage for tomography. The lag-time window of each CCF, specified as t_{min} and t_{max} , is defined using the inter-station distance of each CCF, and velocities V_{min} and V_{max} specified in Table 2.1. The velocities are chosen as minimum and maximum values that provide a window that best represents the velocity range that signal may travel at between stations. All signal outside of the window defined above is considered noise.

The minimum wavelength criterion varies between papers; Bensen et al. (2007) proposes a minimum inter-station distance of three wavelengths, Shapiro et al. (2005) use two and Brenguier et al. (2007) use one. Studies using broadband signals tend to require a higher number wavelengths while studies using higher frequency bandwidths use less (Brenguier et al., 2007). We require one wavelength because our dispersion measurements appear to remain stable, allowing us to have a higher number of measurements at longer periods.

Phase velocity dispersion measurements at discrete periods are plotted as a function of travel-time and distance to look for outliers and cycle skipping of phase velocity measurements (Lin et al., 2008). This occurs when the number of wave cycles calculated using the reference dispersion differs from the real value (N in equation 2.6).

To test the reliability of our dispersion curve measurements we adapt the method of Bensen et al. (2007) and measure dispersion curves from CCFs made of sub-stacks representing one day of data. Following Behr et al. (2011), we measure a lower and upper dispersion curve for every dispersion curve measurement by taking the lower and upper envelope of the daily dispersion curves. This allows us to capture the variability of our dispersion measurements. While it is not a direct measure of error, it provides an estimate of dispersion curve error.

2.2 Locating Persistent Localised Sources

Inspection of CCF virtual source moveout plots show a linear feature of a similar frequency and slope to the direct wave signal [see Figure 3.3b for example]. Zheng et al. (2011) found that stationary sources, termed persistent localised sources, located within an array can generate a signal similar to what is observed in our CCFs. The presence of this signal in CCFs when measuring surface wave velocity using AFTAN will bias the group velocity measurement. If the stationary source signal is at a different frequency to the ambient noise wave-field frequencies then frequency filtering can remove the signal. But signal at a similar frequency can only be removed by discarding affected CCF traces. To identify these traces the persistent localised sources must be located.

To locate the sources, we follow Zheng et al. (2011) who used a back-projection method similar to Ishii et al. (2005), but adapted to CCF. First, all points from a 10 m spaced grid are considered as potential source locations where the lag-time of the CCFs is the difference in distance from potential source to virtual source and receiver, divided by apparent velocity. A constant apparent velocity is assumed, which is calculated from the slope of the persistent localised source phase on virtual source moveout plots. To counter the effects of positive and negative amplitudes

cancelling, the positive envelope function is taken for each CCF [Figure 2.5b]. The amplitude of each lag-time for every potential source is then mapped through space to give a grid of hyperbolas that map the possible source location for every CCF lag-time [Figure 2.5c]. We repeat this for every CCF with a stationary source phase present and then stack the amplitude grids to give a final location [see Figure 3.8 for affected stations].

The persistent localised source signal is in a similar spectral range (0.4 to 0.8 s period) to our CCFs, making it difficult to remove via frequency filtering. To ensure unbiased dispersion curve measurements, acausal or causal time lags are removed from a CCF if it is contaminated by a persistent localised source phase. Using the back-projection location, the distance from the persistent localised source to both virtual source and receiver is calculated and if it is located within a great circle connecting both station and receiver, then the CCF is considered contaminated. The persistent localised source is located on the positive lag if it is closer to the virtual source and on the negative lag if it is closer to the virtual receiver. For CCFs where both stations are identified as contaminated, the function is discarded if the persistent localised source is located within the great circle connecting both virtual source and receiver. For stations not identified as contaminated, the causal and acausal functions are stacked into a symmetric CCF.

2.3 Inversion for Velocity Structure

Surface wave dispersion curves can be inverted for tomographic maps of surface velocity at discrete periods and for shear wave velocity with depth. Inverting a surface wave dispersion curve calculated from a CCF for shear-wave velocity will give the average 1D velocity structure between two stations (e.g. Behr et al., 2010). To better model the velocity structure, surface-wave tomography maps of discrete periods can be used to create interpolated dispersion curves from each grid point

Figure 2.5: a) CCF, b) positive envelope function of CCF, c) back-projection of positive envelope function. The CCF is first bandpass filtered between 1 and 5 Hz and then a positive envelope function is calculated. A backprojection (c) is then applied to the envelope function (b). The red and blue arrows show the location of the stationary phase signal on the envelope function (b) and each locations potential source location is mapped (c). Real amplitudes are used for the stacked back-projections despite being normalised in this example. The blue star is the virtual source and white triangle is the vertical receiver

within the tomographic model (e.g. Behr et al., 2011, Lin et al., 2013). The interpolated dispersion can then be inverted for shear-wave velocity, giving the 1D velocity structure for a grid point. Using 1D inversions from each grid point, a pseudo-3D shear-wave velocity model is constructed by interpolating the 1D profiles into 3D space.

2.3.1 Surface Wave Tomography

The calculation of dispersion curves from CCFs make these measurements suitable to tomography methods as both the source and receiver locations are known. A popular method was proposed by Barmin et al. (2001) which minimises a penalty function, assuming no ray-bending. More recently, eikonal tomography has become a common method which solves the eikonal equation for apparent phase velocity

by tracking a wave-front propagating across an array from a virtual source and calculating the travel-time gradient and its direction at each grid point (Lin et al., 2009). We choose to not use the Barmin et al. (2001) method because no codes were available that would suit this dataset and writing new ones is beyond the scope of this thesis. Eikonal tomography uses bent-ray theory so we decide to use a simpler method considering the array geometry.

To calculate the spatial variation of phase and group velocities within the study area, we use a linear straight ray travel-time tomography method (Uieda et al., 2013). The problem is in the linear form of

$$d = G(m) \tag{2.7}$$

where d is the data, m is the model and G is a linear function linking d and m . The problem minimises the objective function ϕ of travel-time misfit between the observed and modelled travel-time by finding the parameter vector $\bar{p}$ that solves the linear system

$$\bar{\bar{H}}\bar{p} = -\bar{g} \tag{2.8}$$

where $\bar{\bar{H}}$ is the Hessian matrix of ϕ and $\bar{g}$ is the gradient vector of ϕ . Minimising the objective function ϕ through a straight-forward linearisation is very sensitive to outliers, resulting in a unstable and unrealistic solution (Uieda et al., 2013). To overcome this we apply a regularizing function that uses a 1st order Tikhonov regularization that requires adjacent parameters have similar values. The strength of smoothing is controlled by a scalar value chosen by finding the value where misfit is stabilised.

Checker-board Tests

To test for lateral resolution of velocity maps we employ checker-board tests (e.g. Rawlinson et al., 2014). The checker-board test models how well a synthetic model made up of blocks with varying velocity can be reconstructed using the ray-paths

of the observed dataset. Travel times are first calculated for each ray-path using the synthetic velocity model, where the ray-paths are same used for each velocity map calculation. Tomography is then performed on the travel-times as outlined in section 2.3.1. The level of similarity between the input synthetic model and reconstructed synthetic model gives an estimate of resolution. Typical features observed are smearing of checkers and the damping of velocity (Rawlinson et al., 2014).

2.3.2 Inversion for Shear-Wave Velocity

Inversion of surface wave dispersion curves is a non-linear optimization problem that gives non-unique solutions (Dunkin, 1965). Popular techniques include linearised methods and direct search techniques (Wathelet et al., 2004). Direct search methods are well suited to problems with a limited number of unknowns (Wathelet, 2008).

We use the Neighbourhood Algorithm (NA), a stochastic direct search method that is steered towards a best fitting solution (Sambridge, 1999a,b). The software package Dinver is used which implements a modified version of the NA where a parameter space with irregular boundaries can be defined (Wathelet, 2008, <http://www.geopsy.org/wiki/index.php/Dinver>, last accessed May 2017). We start by defining a parameter space of shear velocity, thickness, compressional velocity (linked to shear-velocity by Poissons Ratio) and density that is divided up into a number of Voronoi cells (n_{s0}) by a pseudo-random generator. For each Voronoi cell sample, a forward model dispersion curve fitting the sample parameters is computed using the method of Dunkin (1965). A misfit value between the forward modelled dispersion curve and the input data (in our case a dispersion curve(s)) is calculated using the equation

$$misfit = \sqrt{\sum_{i=0}^{n_F} \frac{(x_{di} - x_{ci})^2}{\sigma_i^2 n_F}} \quad (2.9)$$

where x_{di} is the velocity of the observed curve at frequency f_i , x_{ci} is the velocity of the theoretical curve at f_i , σ_i is the uncertainty of frequency samples and n_F is the number of frequency samples considered (Wathelet, 2008). We do not give uncertainty for observed dispersion curves so replace σ_i with x_{di} (following Wathelet et al., 2004). A new random generation of samples is guided by the misfit value from previous samples. For all Voronoi cells, n_r number of cells with the lowest misfit are chosen and n_s number of new samples are randomly generated to better sample promising area of low misfit. The inversion is then repeated for the new and old cells, and for each new point a misfit value is calculated. The process continues for it_{max} number of iterations (Wathelet, 2008). For a discussion of the parametrisation used in this thesis, see Section 4.1.4.

2.3.3 1D and pseudo-3D Shear-Wave Velocity Modelling

The benefit of using the Neighbourhood Algorithm for our shear-wave velocity inversions is the ability to model the results based on misfit values. Instead of accepting the model with the lowest misfit, we consider all models below a misfit value and calculate the mean at 10 m depth increments using misfit values as a weighting scheme [see Section 2.3.4 for how the misfit value is chosen]. This method serves to smooth the shear-wave velocity profile. We can also test if the inversion solution has converged on a model by plotting a mesh of cells that shows the number of accepted models going through each cell. If a sufficient number of models is contained in a narrow zone of cells then the inversion has converged.

In ambient noise tomography, a common method for modelling 3D variations of shear-wave velocity is to project 1D shear-wave velocity profiles into 3D space using linear interpolation (e.g. Behr et al., 2011, Hannemann et al., 2014, Lin et al., 2013). This type of model is referred to as a pseudo-3D velocity model as it is not a 3D inversion (Behr et al., 2011). To calculate a velocity model using a 3D inversion, the 3D wave equation must be solved (Tape et al., 2010) or 3D sensitivity kernels

can be used (Fichtner, 2010, Zhou et al., 2006). Previous studies have shown that the linear interpolation method is a reasonable approximation of 3D structure at a local scale so we decide to use this as it is a simpler method (Breguier et al., 2007, Hannemann et al., 2014). The limitation of this method is that it involves two inversions, the first being surface wave tomography maps and the second the shear-wave velocity inversion; hence we interpret our results carefully (Behr, 2011).

2.3.4 Error Analysis

To calculate error in our surface-wave tomography and shear-wave inversions we follow Behr et al. (2011). Surface-wave tomography error for phase and group velocity is estimated by taking the low and high bound of a set of daily dispersion measurements described in Section 2.1.4 and performing tomography on both sets, giving three surface wave maps for each period at which tomography is performed. We then construct dispersion curves at each 100 m cell in the tomography map from the high and low bound tomography models for both velocity types [see Figure 3.21 for example]. We then use these bounds to provide an estimate of the misfit value in the 1D shear-wave inversions that constrains all acceptable models in each inversion (Godfrey et al., 2017). These accepted models are then used to calculate the mean shear-wave profile using misfit as a weighting.

The non-linear relationship between surface wave dispersion curves and shear-wave velocity profiles makes it difficult to link CCFs and dispersion curve uncertainty to shear-wave velocity models. Therefore all our error analysis is empirical (Behr, 2011).

By using a weighted mean for the 1D shear-wave profiles we are able to estimate the standard deviation of the velocity and depths of each profile. To do this, we calculate a histogram of shear-wave velocity at 10 m depth intervals and fit a Gaussian curve to each histogram. Following Behr et al. (2011), we define one standard deviation of every curve as the measure of uncertainty. To evaluate depth uncertainty we

follow a similar procedure except that the histograms of the depth layers set in the inversion parametrisation are used. The resulting standard deviation of each depth layer is then interpolated at 10 m intervals.

3 Results

This chapter presents the results of our investigation into the velocity structure of the Whataroa Valley using a temporary geophone array. Cross-Correlation Functions (CCF) of the seismic ambient noise wave-field are used to calculate surface wave dispersion curves. We show back-projected CCFs that reveal a stationary source present in the study area. Rayleigh wave phase velocity maps are then constructed from phase velocity dispersion curves. A pseudo-3D shear-wave velocity model is calculated by inverting dispersion curves constructed from phase velocity maps over an evenly spaced grid and then interpolating the 1D profiles across 3D space.

3.1 Cross-Correlation Functions

We computed 53,428 daily CCFs that are stacked into 12,247 final CCFs representing all pairwise station combinations, made up of three days of data; Julian day 017, 018 and 019. The CCFs can be plotted in terms of their lag-time and inter-station distance, which we refer to as moveout plots. To view all CCFs on a single plot, we stack them into 25 m inter-station distance bins to facilitate interpretation [Figure 3.1]. By plotting CCFs as a function of inter-station distance, we observe several phases in the CCFs that originate from both zero and non-zero lag-times. Phases originating from zero are called direct waves and propagate from the virtual source to the receiver in positive lag-time, and from receiver to the source in negative lag-time. A high-frequency phase (>1 Hz) travelling at an apparent group velocity of 500 m/s is present in CCFs at distances up to 3 km. A large reduction in amplitude is apparent after 3 km [Figure 3.1].

Lower-frequency (<1 Hz) phases appear at shorter lag-times, with a peak occurring at zero. These phases may be a pseudo auto-correlation that forms from a high similarity between longer-period surface waves at close inter-station distances (Chávez García and Rodríguez, 2007). Body waves may also form a peak at zero; these can be removed by selectively removing individual CCFs containing such peaks. Our CCFs contain these peaks for all inter-station distances, and therefore we suspect these features to be pseudo auto-correlations of surface waves rather than cross-correlations of body waves. Several other phases with a similar moveout slope to the direct phase appear at longer lag-times in CCFs further than 2 km station separation.

Figure 3.1: Moveout plot of all raw CCFs stacked into 25 m distance bins.

The pseudo auto-correlation feature present in most CCFs can be removed with a 1 Hz high-pass filter [Figure 3.2]. The strongest feature is the moveout from zero lag-time of the direct wave at 500 m/s. There are also phases parallel to the direct

wave at both shorter and longer lag-times. The parallel phases appearing at longer lag-times are referred to as coda. The direct wave phase has a large reduction in amplitude at 3 km inter-station distance in Figure 3.2. The attenuation in amplitude of the direct wave is stronger on the positive side than the negative side of lag-time; however due to the orientation of CCFs we cannot distinguish direction in Figure 3.2 and 3.1.

Figure 3.2: Moveout plot of all CCFs stacked into 25 m distance bins and high-pass filtered at 1 Hz. The red arrows marks the direct wave and the blue polygon show the phases we consider coda.

To investigate spatial variations of CCF features, we plot a selection of different stations as a virtual source against all other stations [Figure 3.3]. These plots allow for the direction of wave propagation energy to be distinguished, giving insight into the symmetry of CCFs and the direction of noise sources. We do not refer to the causal and acausal components when describing plots in Figure 3.3, rather CCFs are plotted as positive or negative inter-station distance if they are north or south of the virtual source. This means that direct wave signals in causal CCFs with a posi-

tive inter-station distance are travelling away from the virtual source, and signals in the causal CCFs with negative inter-station distance are travelling towards the virtual source. Amplitudes of each CCF are normalised so cannot be compared directly.

Virtual source moveout plots of stations located in the northern part of the array generally exhibit lower quality CCFs [Figure 3.3b and c]. Multiple phases of a similar slope to the direct wave that cross the zero lag-time but do not originate from zero appear in most of these stations; we refer to these signals as the persistent localised source phase [green box in Figure 3.3c and d; see Section 3.1.1]. Although stations east of the Whataroa River are contaminated by this signal, a clear direct wave is usually observed with moderate symmetry [Figure 3.3b]. Within the northern part of the array, stations located north of State Highway 6 [see transparent red box in Figure 3.5b for road location], along the western edge of the Whataroa River are the lowest quality stations in the array [Figure 3.3c]. Weak amplitude direct waves and strong contamination from the persistent localised source phase described above are typical characteristics.

At stations located near the DFDP-2B borehole, a strong direct wave is observed travelling away from the drill-pad [Figure 3.3a]. Strong asymmetry is also observed in most virtual source moveout plots in this area. In other parts of the central area of the array, stations generally have a moderate to strong amplitude direct wave and good symmetry [Figure 3.3f]. The persistent localised source phases are also present in these stations, especially at longer inter-station distances. We observe a causal relationship in virtual source move-out plots where the inter-station distance of the stationary source phase intersection with the direct phase is dependant on the distance of the virtual source from the DFDP-2B site area. As the distance increases, the inter-station distance where the intersection occurs increases [See Figure 3.3d,e and f].

A moderate amplitude direct wave is observed in virtual source move-out plots lo-

cated in southern end of the array [Figure 3.3d and e]. Most plots have a weak symmetry with the strongest phase travelling towards the virtual source.

In some plots a coda-like arrival is observed at inter-station distances above 1000 m [Figure 3.3b, c, d and f]. The general relationship observed is that the length of the coda, defined as the difference between the maximum lag-time of the coda and the lag-time of the direct phase, tends to increase with inter-station distance [e.g. Figure 3.3c].

To investigate the signal-to-noise ratio of the CCFs and how it relates to the distribution of noise sources, we plot the signal-to-noise ratio of virtual source move-out plots in Figure 3.3 [see Section 2.1.4 for a definition of the signal-to-noise ratio]. For each move-out plot we calculate the signal-to-noise ratio of the causal and acausal component, where the axis of separation is perpendicular to the down valley axis [Figure 3.4]. Noise source direction and strength can be discriminated in terms of which hemisphere it is coming from. The strong asymmetry observed in Figure 3.3a is evident in Figure 3.4a. In stations down valley (north) of the virtual source there is very low signal-to-noise ratio in CCFs from signal travelling from the down valley direction but high signal-to-noise ratio is observed in these CCF from signal travelling from up valley (South). For stations up-valley of the virtual source, the opposite relationship holds true whereby a high signal-to-noise ratio occurs in the CCFs component for signals travelling from the virtual source to the virtual receiver [Figure 3.4a]. A decay in signal-to-noise ratio as inter-station distance increases is also observed [e.g. Figure 3.4a, c and d]. Localised bands of high signal-to-noise ratio (above 20) are observed in some virtual source move-out plots [Figure 3.4b, c and f]. A common feature of these high signal-to-noise ratio bands are that they line up with the area around the DFDP-2B borehole [located near the black star in Figure 3.4a]. In general we observe higher signal-to-noise ratios in direct waves travelling from down valley; however a source within the array appears to be present.

Figure 3.3: Moveout plots of station as a virtual source vs all other stations. Stations as virtual source a) 643, b) 664, c) 668, d) 684, e) 784, and f) 884. Each CCFs has been filtered with a 1 Hz high-pass. See Figure 3.4 for station locations. The red polygon marks the direct wave, green polygon marks the persistent localised source phases and blue polygon marks the coda. The coloured dots in b show the stationary source phase used in Figure 2.5.

Figure 3.4: Maps of signal-to-noise ratio of stations used in Figure 3.3 as a virtual source (black star). The black half of the beachball in each figure shows the direction the potential signal sources are coming from. For example, signal propagating towards the array from a source located north of the array would be plotted on the left panel of each subplot. By splitting the potential noise source direction in half, we can delineate between signal travelling down and up the valley.

3.1.1 Back-projection of Cross-Correlation Functions

Assuming a constant velocity of 450 m/s, we apply a back-projection to the CCFs to locate sources of coherent noise from within the array [see Section 2.2 for a description of this technique]. To estimate the velocity constant, we use the slope of the phase identified in Figure 3.3b with red dots. This velocity is consistent with other values observed in similar phase's from different move-out plots. Stacking virtual source back-projections contaminated by zero-crossing phase reveals a high-amplitude anomaly [Figure 3.5a], which we call a persistent localised source. To further investigate the location we plot normalised amplitudes above 0.8, a value found on the basis of trial and error to best represent the extent of the anomaly [Figure 3.5b]. We find the location occurs close to the Whataroa River. Despite evidence for more than one signal [See Figure 2.5 and 3.3b], we are only able to recover one location. Using the method outlined in Section 2.2, we construct symmetric CCFs with the stationary source removed.

Figure 3.5: a) Back-projection of all contaminated CCFs, and b) Colour scale limited to 0.8-1.0. To better locate the anomaly; stations, the Whataroa River, roads and the Alpine Fault are plotted in (b). Stations treated as a virtual source in the back-projection are plotted as orange circles while those not used are plotted as orange triangles. The transparent red box shows poor quality stations north the road, identified in section 3.1.

3.2 Rayleigh Wave Dispersion Curves

After correcting for the stationary source phase, all symmetric CCFs are used to measure dispersion curves, yielding a total of 7758 dispersion curves containing 64 group and phase velocity measurements each. From DFDP-2B well logs, we assume there is a sedimentary-basement contrast within the study area (Sutherland et al., 2015). To model the effect of basement depth on Rayleigh wave dispersion we compute fundamental and first-order higher mode synthetic dispersion curves [Figure 3.6]. The Gpdc software package is used, where our input model is one-layer over a half-space, representing the sedimentary layer and the basement (<http://geopsy.org/wiki/index.php/Gpdc>, last accessed May 2017). We vary thickness of the first layer from 0 to 300 m and use a shear-wave velocity of 420 m/s for the sedimentary layer and 1500 m/s for the basement. A V_p/V_s ratio of 1.75 is used

to calculate compressional-wave velocity and the density is calculated using Garners equation [see Figure 2.4].

Fundamental mode dispersion of group velocity shows a large jump due to the velocity contrast in the model [Figure 3.6]. The depth of this contrast controls the period where the jump occurs. A decrease in velocity at periods shorter than the jump period is observed which is consistent with simplified theoretical group velocity models (Stein and Wysession, 2009). Fundamental mode phase velocity dispersion shows a less abrupt increase in velocity than group dispersion although the period where the increase occurs also increases with basement depth [Figure 3.6]. Higher-mode dispersion is more complex, especially group velocity where rapid increases and decreases in velocity are separated by a constant velocity whose period appears to be controlled by basement depth. Phase dispersion for the first higher mode has a higher degree of similarity to the fundamental mode than group velocity does. The main difference between phase dispersion modes is that the fundamental mode shows an exponential increase in velocity and the first-order mode is cubic-like. The model with no overlying sediment [marked as 0 in Figure 3.6] is not present in higher mode group and phase velocity curves as there is no dispersion in a homogeneous half-space (Stein and Wysession, 2009).

Figure 3.6: Synthetic Rayleigh wave dispersion curves a) group velocity - fundamental mode, b) phase velocity - fundamental mode, c) group velocity dispersion - first order higher mode, and d) phase velocity - first order mode. Individual curve labels specify the depth in meters at which the velocity contrast occurs, representing the basement contact.

A selection of individual station pair dispersion curves, representing different ray-paths, is plotted in Figure 3.7 [see Figure 3.8 for ray paths]. Clean dispersion measurements from station pairs with a shorter inter-station distances are observed, allowing a broad range of azimuths due to the shorter periods of the surface wave train [Figure 3.7a and b]. Measurements from station pairs with an inter-station distance less than 1000 m generally have group velocities of 400-600 m/s and phase velocities of 600-800 m/s.

Dispersion measurements are also obtained from station pairs with inter-station distances of between 1000 and 2000 m [Figure 3.7c and d]. Both station pairs have a clear mode with an amplitude between 97 and 100 dB for the picked dispersion curve. Another mode may also be present at higher velocities, although the energy

of the amplitude is lower at 90 to 97 dB.

Longer periods are observed for station pairs with an inter-station distance greater than 2000 m, reaching up to 2.0 s for group velocity in Figure 3.7e and f. Group velocities of up to 800 m/s are observed in station pairs 670-785 and 672-690 [Figure 3.7d and f]. Higher phase velocities are also observed, with measurements up to 1700 m/s observed at longer periods [Figure 3.7e].

Figure 3.7: Rayleigh wave group (blue) and phase (black) velocity dispersion curves from individual station pairs a) 609-652, b) 609-658, c) 635-843, d) 670-785, e) 672-690, and f) 671-697. A Selection criteria of 1 wavelength and a signal-to-noise ratio minimum of 10 is applied to all measurements. See Figure 3.8 for ray-path locations.

We apply selection criteria of one wavelength and a Signal to Noise Ratio (SNR) minimum of 15 to all dispersion curves, resulting in 5981 group and 4560 phase velocity dispersion curves [Figure 3.9a and b]. Group velocity measurements show a range of velocities of between 400 to 2000 m/s and period ranges from 0.3 to 2.0 s. Phase velocity measurements generally have a higher velocity than group velocity with a maximum velocity of 2500 m/s observed. The maximum period observed is

shorter than that of the group velocity at 1.5 s, as expected due to the minimum wavelength criteria applied. A strong velocity variation is observed in group and phase velocity measurements [Figure 3.9a and b].

The broad range in group velocity values suggest the persistent localised source phase(s) has not been completely removed from the data [see Section 2.2]. To test this, we remove all dispersion measurements that use any station located in the northern part of the array [defined as grey triangles in Figure 3.8]. The total number of group and phase velocity measurements reduces to 2952 [Figure 3.9c] and 2260 [Figure 3.9d] respectively; we hereby call this the sub-array [Figure 3.8]. There is a clear reduction in the variation of velocities observed, suggesting that persistent localised source phases remain in symmetric CCF.

Figure 3.8: Map of station paths used in Figure 3.7 [red lines]. Triangles with orange fill are stations used in the sub-array and grey fill are rejected stations. The Whataroa River, roads, Alpine Fault surface trace, DFDP-2B and 50 m topographic contours are marked in blue, white/black, thick black, yellow and thin black colours respectively.

Figure 3.9: Rayleigh wave group and phase velocity dispersion curves a) group velocity - full array, b) phase velocity - full array, c) group velocity - sub array, and d) phase velocity - sub array. A selection criteria of one wavelength and a minimum SNR of 15 is applied. The top panels are measurements from the full array and the bottom panels are taken from station pairs not contaminated by noise from the river [see Figure 3.5].

There is a reduction in steep phase dispersion curves between the full-array and the sub-array, although a number still remain [Figure 3.9]. We test for cycle skipping in phase velocity measurements by plotting travel time vs. inter-station distance [Figure 3.10]. Cycle skipping will introduce a $2\pi N$ error when the phase velocity reference curve used to calculate the number of wave cycles, N in Equation 2.6, varies from the actual values Lin et al. (2008). As we only use one reference curve, and show through synthetics that variation in basement depth results in different dispersion curves [Figure 3.6b], it is important to test for cycle skipping. A clear lin-

ear trend is observed in dispersion measurements from the sub-array [Figure 3.10b]. This trend is less clear in measurements from the entire array [Figure 3.10a], with a high degree of scattering linked to the high velocity measurements observed in Figure 3.9a and b. The parallel branch structure of measurements of a similar slope to the main branch step out at near equal intervals of 0.6 s, suggesting these measurements have a $2\pi N$ error [Figure 3.10b].

Figure 3.10: Travel-time vs. interstation distance plot of all period measurements at 0.6 s period a) full array, and b) sub-array. The red lines are plotted at 0.6 s increments, showing the correct number of cycles N [blue line in figure] plus x , where x is $\pm 1, 2, 3$ the number of cycles skipped.

3.3 Rayleigh Wave Velocity Maps

Straight ray travel-time tomography is applied to discrete phase and group dispersion curve measurements from 0.3 to 1.0 s period and 0.3 to 1.8 s period respectively. We do not interpolate between measurements to recover the desired period and instead only select dispersion measurements that are within 0.02 s of the stated period. Due to the ambiguity of dispersion curve measurements using the full array [see Figure 3.10a], only dispersion curves from the sub-array are used for tomography. Stations used in the sub-array are marked as orange triangles in Figure 3.8 and rejected stations are marked as grey triangles. Measurements with cycle skipping are removed from phase measurements as we have enough path coverage in the cycle skipping range of 0.4 to 0.8 s period. A maximum period of 1.0 s is used for phase

velocity tomography as there are not enough measurements for complete ray-path coverage at higher periods. Due to slower velocities, group tomography can be extended to 1.8 s period while maintaining good ray-path coverage. From here on, we refer to the area covered by the sub-array as the study area.

3.3.1 Resolution

Resolution of the tomography results was tested using checkerboard tests for phase [Figure 3.11] and group [Figure 3.12] velocities. We also plot ray path density and azimuth coverage for each tomography cell. A two-velocity model is used in the checkerboard tests with a low velocity of 600 m/s and high velocity of 900 m/s in 200 m cells [Figure 3.11a].

Phase velocity checkerboard tests resolved velocities with a higher degree of accuracy in the centre of the valley [Figure 3.11]. Increased smearing and damping occurs towards the outer edge of the study area [Figure 3.11]. At 0.3 s period the input velocity is poorly recovered with only 3 cells resolved in the centre of the study area [Figure 3.11b]. The recovered cells are also smeared in a north-west/south-east direction [Figure 3.11b]. At 0.5 and 0.9 s period the centre of the study area is well resolved with evidence of a small degree of velocity damping [Figure 3.11c and d]. A small degree of north-west/south-east smearing in the centre of the study area is evident in both 0.5 and 0.9 s period tests. The outer edges of the 0.5 and 0.9 s period models are poorly resolved, particularly at the southern extent of the study area.

The best resolved test for group velocity is at 1.3 s period, with a small degree of smearing and damping in the centre of the study area, increasing towards the outer edges [Figure 3.12d]. Compared to phase velocity tests [Figure 3.11], the 1.3 s period is able to resolve more of the input model in the southern part. The 0.3 s period test is poorly recovered with only two cells resolved, and high degrees of damping and smearing occur in a north-west/south-east direction [Figure 3.12a]. The 0.5 s period

group velocity test [Figure 3.12b] appears as well resolved as the 0.5 s period phase velocity test [Figure 3.11c]. 0.9 s period is better resolved than 0.5 s period but still has a small degree of north-west/south-east direction smearing [Figure 3.12c].

The best ray-path coverage occurs at 0.5 s period for phase velocity [Figure 3.13b] and 0.9 s period for group velocity [Figure 3.13f]. All maps show low path densities at the edges of the study area, with density increasing towards the centre. Phase velocity path coverage at 0.3 s period is very low in the south of the study area, but path density increases up to 100 paths per cell in the north of the study area [Figure 3.13a]. At 0.5 and 0.9 s periods, the distribution of paths is less skewed than at 0.3 s period and is centred in the middle of the study area for both phase and group velocity [Figure 3.13b,c,e,f]. The phase velocity at 0.5 s period reaches a density of 150 paths per cell in the centre, in comparison to the 0.3 s period, fewer ray paths are located in the northern area. At 0.9 s period, phase velocity has the poorest ray path density with a density of less than 50 for much of the area [Figure 3.13c]. 0.3 s period for group velocity shows the same increase in path density in the north as the phase 0.3 s period, however path density for group peaks at 100 [Figure 3.13d]. The peak in path density of 0.5 s period group velocity moves south compared to 0.3 s period, occurring in the centre of the study area, reaching 150 at its peak [Figure 3.13e]. Group velocity at 0.9 s period has the highest path density which is greater than 200 in the centre and reduces to 20 around the edges [Figure 3.13f].

Figure 3.11: Rayleigh wave phase velocity checkerboard test a) input velocity model - blue is 600 m/s and red is 900 m/s, b) 0.3, c) 0.5, and d) 0.9 s period. The black polygon shows the bounds of the tomography model.

Figure 3.12: Rayleigh wave group velocity checkerboard test a) 0.3, b) 0.5, c) 0.9, and d) 1.3 s period. The input velocity model used is shown in Figure 3.11. The black polygon shows the bounds of the tomography model.

Figure 3.13: Rayleigh wave phase and group velocity tomography path density a) phase velocity - 0.3s, b) phase velocity - 0.5s, c) phase velocity - 0.9s, d) group velocity - 0.3s, e) group velocity - 0.5s, and f) group velocity - 0.9s. The black polygon shows the bounds of the tomography model.

The best azimuthal coverage occurs at a period of 0.5s for both the phase and group velocities [Figure 3.14b and e]. Both maps show coverage increasing towards the centre of the study area, where in two patches north and south of the centre, values reach almost 100% coverage. The 0.3s period for both velocities shows blanket coverage of around 25% [Figure 3.14a and d]. For the 0.9s period, the phase velocity reaches just under 50% coverage in the centre while the group reaches 80% in isolated cells [Figure 3.14c and f].

Figure 3.14: Rayleigh wave phase and group velocity tomography azimuth coverage a) phase velocity - 0.3s, b) phase velocity - 0.5s, c) phase velocity - 0.9s, d) group velocity - 0.3s, e) group velocity - 0.5s, and f) group velocity - 0.9s. Azimuth calculated using 10° bins and is plotted as a percentage where 100% corresponds to 180° coverage. The black polygon shows the bounds of the tomography model.

3.3.2 Velocity Maps

As period increases, Rayleigh wave phase velocity also increases [Figure 3.15]. At 0.3s period phase velocity is 600 m/s, with the exception of an area of increased phase velocity from the DFDP-2B borehole westward, where velocity is 700 m/s

[Figure 3.15a]. As the period increases a gradient of increasing velocity becomes apparent from the centre of the valley toward the valley walls [e.g. Figure 3.15b]. At the 0.7 s period, the largest range in velocity was found, with a peak in the north-western side at 1100 m/s, while the centre of the valley is 700 m/s [Figure 3.15c]. At this period the velocity only appears sensitive to the western valley wall. In contrast, at a period of 0.9 s the velocity begins to reflect both sides of the valley [Figure 3.15c and d]. At 0.9 s period a low velocity region localises in the centre of the valley [Figure 3.15d]. Velocity peaks at 1100 m/s along the western valley wall and decreases to 900 m/s towards the east [Figure 3.15d].

Group velocity maps show a smaller increase in velocity with increasing period compared to phase velocity. Velocities observed in periods less than 1 s are in the range of 400 to 650 m/s [Figure 3.16]. At 0.3 s period the velocity across the whole valley is 400 m/s, except for a slight increase to 450 m/s in the north-western corner [Figure 3.16a]. At the 0.5 s period, two lobes of low velocity appear at the valley floor with areas parallel to the valley walls increasing to 550 m/s [Figure 3.16b]. The velocity from the borehole to the west increases to 600 m/s while the high velocity area in the north-west increases to 650 m/s [Figure 3.16b]. At 0.7 s period both low velocity lobes decrease in extent, the southern lobe maintains a velocity of 650 m/s and is greatly reduced in extent, while the northern lobe increases to 600 m/s and is marginally reduced in size [Figure 3.16c]. A south-east to north-west gradient exists from 650 m/s in the centre of the valley to 700 m/s in the north west corner [Figure 3.16c]. At 0.9 s period the velocity at the center of the valley is 650 m/s and increases to 780 m/s towards to the valley walls [Figure 3.16d].

Group velocity periods greater than 1 s (1.1 and 1.3 s) have the largest range in velocity [Figure 3.17a and b]. From the centre of valley to the western edge, an increase of 500 m/s is observed [Figure 3.17a and b]. At 1.1 s period this velocity increase is distributed more evenly towards the valley walls, while at 1.3 s period this increase occurs over a shorter distance around the area of the borehole [Figure 3.17a

and b]. At 1.5 and 1.7 s period the velocity range decreases to 200 m/s, with both periods peaking at 700 m/s in the north-western corner [Figure 3.17c and d].

The velocity maps show a trend of lower velocity at the centre of the valley with higher velocities occurring in the north-western corner and towards the valley walls. The group velocity peaked at 900 m/s (1.1 and 1.3 s period) while phase velocity peaked at 1100 m/s (0.7 s). The lowest velocity found in group velocity measurements was 400 m/s (0.3 s period) and 600 m/s for phase velocity (0.3 s period). The phase velocities were overall higher than the group velocity. A change in velocity gradient is observed parallel to the valley walls [e.g. Figure 3.17b].

Figure 3.15: Rayleigh wave phase velocity maps a) 0.3 s, b) 0.5 s, c) 0.7 s and d) 0.9 s. The Whataroa River is plotted in blue, roads are plotted in white and DFDP-2B marked as a yellow dot. The topographic contours are spaced at 50 m.

Figure 3.16: Rayleigh wave group velocity maps a) 0.3 s, b) 0.5 s, c) 0.7 s and d) 0.9 s. The Whataroa River is plotted in blue, roads are plotted in white and DFDP-2B marked as a yellow dot. The topographic contours are spaced at 50 m.

Figure 3.17: Rayleigh wave group velocity maps a) 1.1 s, b) 1.3 s, c) 1.5 s and d) 1.7 s. The Whataroa River is plotted in blue, roads are plotted in white and DFDP-2B marked as a yellow dot. The topographic contours are spaced at 50 m.

A regularization smoothing factor of 4×10^6 is used to smooth the tomography maps. By plotting the smoothing factor against the travel-time misfit, we observe a steep change in misfit value [Figure 3.18]. In both phase and group velocity plots, a rapid increase in misfit is observed as the smoothing factor is increased [Figure 3.18]. Across all periods, misfit appears to stabilise at approximately 4×10^6 . We also plot velocity maps with a smoothing factor that is higher and lower than 4×10^6 [Figure 3.19]. A low smoothing factor of 4×10^5 results in unrealistic velocity anomalies in both phase and group velocity [Figure 3.19a and b]. A high smoothing factor of 4×10^7 results in a model that is too smooth [Figure 3.18a and b]. By visual inspection of velocity maps and plotting the travel-time misfit against the smoothing factor, a value of 4×10^6 is used as this is a good compromise between realistic velocities and a smooth velocity gradient across the maps.

Figure 3.18: Travel-time misfit and smoothing factor applied to Rayleigh wave a) phase and b) group velocity maps at 0.7 s period.

Figure 3.19: Regularization smoothing factor applied Rayleigh wave a) phase and b) group velocity maps at 0.7 s period. The number in each plots is the smoothing factor applied to the tomography map.

3.4 Shear-Wave Velocity Inversions

Phase and group velocity dispersion curves were constructed using a 100 m bin size from each velocity map [Figure 3.15 for phase and Figure 3.16 and 3.17 for group velocity; Note: we only show odd number periods but we include even number periods in the construction of dispersion curves]. We start by inverting phase velocity dispersion curves and then invert group and phase dispersion curves jointly, allowing us to probe the subsurface down to 1000 m compared to 300 m using phase dispersion measurements alone.

3.4.1 1D Shear-Wave Velocity Profiles

During the inversion process, 50000 models were computed as the inversion explores the full parameter space. Instead of using the model corresponding to the lowest misfit, we follow Behr et al. (2011) and compute a weighted mean profile using misfit values as weights. For phase velocity inversions, a misfit threshold value is chosen

based on the velocity error of the observed dispersion curves, as we consider all models below this value as valid (Godfrey et al., 2017). For the joint inversion, this method is somewhat difficult as the dispersion curves have different errors so we use a constant misfit value of 0.4 (decimal percentage) for all profiles.

Phase velocity dispersion curves from 0.3 to 1.0 s period are inverted using two layers over a half space parametrisation and 50 iterations [see Section 4.1.4 for discussion on parametrisation]. A selection of three inversions are presented in Figure 3.20, representing the north-west, central and south-east portions of the study area. Misfit values for inversions near in the north-west region are higher than other areas as illustrated in Figure 3.20a-c where a misfit of 0.03 is obtained for the north-west area compared to 0.01 for other areas. The observed dispersion curve error is also higher for this region when compared to the other areas [Figure 3.20a-c]. A 200 m/s error is observed in the north-west section observed dispersion curve at period of 0.5 s, reducing to 100 m/s for other periods. A smaller error is observed for other dispersions curves in the order of 10-50 m/s [Figure 3.20b-c].

The shear-wave velocity depth profiles are constrained to two-layers over a half-space. The bottom of the upper layer is in the upper 20 m in Figure 3.20d and f whereas it is located deeper at 80 m in Figure 3.20e. The bottom of the lower layer follows a similar pattern where it occurs at 220-240 m in Figure 3.20d and f and at 260 m in Figure 3.20e. Very low velocities of 150-250 m/s are observed in the upper layer of Figure 3.20d and f, before rapidly increasing to 800 and 700 m/s in the lower layer. A higher velocity of 500 m/s is observed in the upper layer of Figure 3.20e before an increase to 700 m/s to the lower layer. The half-space velocity is highest in the north-west profile at 1200 m/s [Figure 3.20d]. In the south-east profile it is slightly less at 1150 m/s [Figure 3.20e] and in the central profile it is lowest at 1000 m/s [Figure 3.20f].

Figure 3.20: Shear wave inversion results from selected geographic locations using phase velocity dispersion curves. Plots a-c show 1000 random theoretical dispersion curves for each inversion, coloured by velocity misfit (decimal percentage), and the observed dispersion curve and its error bar [black dots]. The random dispersion curves are drawn from the set of dispersion curves that correspond to each shear-wave profile. Shear wave depth profiles are plotted from d-f and correspond to plots a-c with misfit values of 0.35, 0.07 and 0.15. The shear-wave profiles are plotted as number of models per cell, with a cell size of 10 m by 10 m/s. The grey profiles are those above the defined misfit threshold. The location of each inversion profile is marked on Figure 3.22.

Group and phase dispersion curves from periods of 0.3 to 1.8 s and 0.3 to 1.0 s respectively are jointly inverted using a 4 layer over a half space model parametrisation and 500 iterations [See section 4.1.4 for discussion on parametrisation]. Higher misfit values are observed in the joint inversion compared to phase inversions, particularly in the northern area [Figure 3.21a]. Group velocity dispersion generally has a lower error than phase dispersion, although some cells show the opposite [e.g. Figure 3.21f].

In the shear-wave profiles, velocities between 500 m/s and 700 m/s are found in the upper 100 m of the north-west and central study areas [Figure 3.21g and h]. In south-east area this layer extends to depths of 140 to 160 m [Figure 3.21i]. A jump in velocity to 900 m/s in the second layer is observed in the north-west and south-west [Figure 3.21g and i]. The velocity of this layer in the central profile is 700 m/s [Figure 3.21i]. The depth of this layer is 300 m in the north-west and central profiles [Figure 3.21g and h], and 400 m in the south-east profile [Figure 3.21i]. There is a rapid increase in velocity of 500 m/s between the second and third layer in the north-west and south-east profiles [Figure 3.21g and i]. The increase in velocity from the second to third layer in the central area is 400 m/s [Figure 3.21h]. The third layer extends to a depth of 1000 m in the north-west profile and has a velocity of 1400 m/s [Figure 3.21g]. In the central profile it extends to 600 m at 1100 m/s, before a gradual increase in velocity to 2000 m/s occurs over a depth range of 200 m [Figure 3.21h]. The south-east profile is similar to the central profile, except the third layer velocity is the same as the north-west profile and the gradual velocity increase does not occur until 800 m depth [Figure 3.21i].

Figure 3.21: Shear wave inversion results from the same geographic locations as Figure 3.20, using phase and group velocity dispersion curves. The dispersion curves plots and shear-wave depth profiles are constructed using the same method as Figure 3.20. The only difference is a constant misfit threshold of 0.4 is used. Plots a-c and d-f show phase and group velocity dispersion curves for each inversion. Shear wave depth profiles are plotted from g-i and correspond to plots a-f. Note the observed phase dispersion curve is identical to Figure 3.20a-c

3.4.2 Pseudo-3D Shear-Wave Velocity Profiles

A pseudo-3D shear-wave velocity model is constructed from 129 weighted mean 1D depth profiles [e.g. Figure 3.21] which represent the average shear-wave velocity in each 100 m cell of the velocity maps. We choose to use the joint inversion over the phase inversion for the construction of our pseudo-3D model [see Section 2.3.2]. From this model, 2D profiles along west-east and south-north orientations are extracted to evaluate the velocity structure. To better visualise the velocity structure we construct a fence diagram made up of four west-east profile and three south-north profiles [Figure 3.22].

As our main aim in this thesis is to investigate the sedimentary-basement contact structure in the Whataroa Valley, we choose a colour scheme that best highlights sharp changes in velocity gradient, indicative of an strong impedance contrast. The consequence is we are not able to view shallow layers such as the layer at 100 m depth identified in Figure 3.21 [see Section 4.3]. The pseudo-3D model shows two sharp velocity transitions [Figure 3.22]. The first is present across the entire model and has a jump in velocity from 750 m/s to 1250 m/s in the north-west and south-east areas, with a depth range of 300 m to 400 m. The contrast is present in the central area; however its velocity increase is from 750 m/s to 1100 m/s. While the first velocity transition is relatively planar, the second is bulge-like and occurs at the lower depth limit of our model where a transition to velocities above 1600 m/s occurs. The peak of the second transition is along the north-south axis, reaching a minimum depth of 650 m [see central south-north profile in Figure 3.22]. The transition becomes deeper to south-east and north-west where it tapers to a thin slice along the edges of the study area. One exception to this is along the south-west edge of the study area where the transition maintains a depth range of 650 to 900 m. An interesting feature of the central area is the velocity zone between the two velocity transitions described above. The zone has a velocity range of 1100 to 1250 m/s which extends down to 800 m in some areas [Figure 3.22].

Figure 3.22: Fence plot of shear-wave velocity depth profiles of the Whataroa Valley using joint shear-wave inversions of phase and group velocity dispersion curves. Profile transects are marked on the map along with profile locations for Figure 3.20 and 3.21 marked as coloured circles and the red star is DFDP-2B. The blue circle refers to the north-west profile, yellow to the central profile and orange to the south-east profile.

4 Discussion

This chapter provides a discussion on the limitations and considerations of the results presented in the previous chapter. We then interpret the shear-wave velocity model before comparing our findings to those of previous studies.

4.1 Data Analysis and Considerations

Here we discuss the limitations and considerations of the four main components of this study; cross-correlations, dispersion curves, tomography and shear-wave velocity inversions.

4.1.1 Cross-correlation Functions

The data used in this study were not recorded with the intention of performing ambient noise analysis, therefore this study is in part a test of whether a short deployment of a few days can be used to perform this type of analysis. The length of recording time is an important consideration when cross-correlating seismic ambient noise. A CCF constructed from a single or small number of correlations will not be an accurate approximation of the Green's Function, whereas stacking all individual cross-correlations will start to represent an approximated Green's Function as a increasing number of individual Cross-Correlation Functions (CCF) are stacked (Lin et al., 2006). Although there is no way to empirically calculate the number of individual CCFs required, it is possible the four days of data we use in this thesis may not be enough to achieve a stable CCF. To counter this limitation we extend the standard processing techniques recommended by Bensen et al. (2007) to test phase

coherence based methods of phase-weighted stack and phase cross-correlation. We observe higher signal-to-noise ratio in CCFs by employing the phase weighted stack method [Figure 4.1]. For virtual sources located at the northern and southern extent of the array, a larger difference in signal-to-noise ratio is observed than for virtual sources located near the persistent localised source. We showed in Figure 3.4 that the persistent localised source location acts as a strong source which suggests the small signal-to-noise ratio difference observed in the central part of the array [for Figure 4.1a and b] is due to the strong source, as is the lower signal-to-noise ratio difference for virtual sources in the central part of the array [Figure 4.1c]. In some virtual source plots, such as Figure 4.1a, the signal-to-noise ratio difference can be related to the inter-station distance where the difference increases with distance. This suggests that with increasing inter-station distance, the phase-weighted stack improves the CCF at a greater rate than linear stacking. There is no evidence to suggest that the phase weighted stack allows us to observe new signals, but rather signals observed will be cleaner and thus result in more reliable dispersion curves.

Figure 4.1: Signal-to-noise ratio difference between linear and phase-weighted stacked CCF. Each plot shows a different station treated as a virtual source with the signal-to-noise ratio calculated from moveout plots similar to Figure 3.3 a) station 621, b) station 666, and c) station 636. Virtual source location is marked as a yellow star and its station written above the map scale. The Whataroa River is also marked for reference. Positive means phase-weighted stack has a higher signal-to-noise ratio than the linear stack.

Using both cross correlation methods, we observe the same signal propagating at an apparent group velocity of 500 m/s [Figure 4.2]. A comparison of CCFs at different inter-station distances shows phase cross-correlations have a higher correlation coefficient than conventional CCFs. The low frequency peak at zero lag-time is absent in phase cross-correlations, except for the 3.44 km example [Figure 4.2d]. Schimmel et al. (2011) show the phase cross-correlation method is better at correlating the entire wave-train as it is sensitive to small waveform changes whereas the conventional CCFs is biased by high amplitude phases and thus the correlation may have a peak at this phase. The presence of the high amplitude peak in our CCFs suggests that our preprocessing steps of RMS clipping and spectral whitening have not been entirely successful in removing high amplitude arrivals. It is likely the peak is a pseudo auto-correlation of long-wavelength arrivals that are very similar at both source and receiver (Chávez García and Rodríguez, 2007). We have attempted to mitigate this by limiting the spectral range of frequencies in our cross-correlation in the preprocessing steps, but the zero lag-time peak is 0.5 Hz, thus at the edge of

our spectral range defined as 0.5 to 5 Hz during whitening [see Figure 2.2]. An improvement in the processing flow could be to use dynamic whitening bounds chosen to correspond to a minimum frequency set by a wavelength criterion. The presence of a high amplitude direct phase in both types of CCF leads us to use the conventional method over the phase cross-correlation method. The conventional method provides a robust result at all inter-station distance whereas phase cross-correlation works very efficiently at short inter-station distances but appears to reduce in effectiveness as inter-station distance increases [Figure 4.2d]. Through the adoption of the non-conventional processing techniques of phase weighted stacking, we have shown that a short deployment of temporary instruments can be used to perform ambient noise analysis.

Figure 4.2: Comparison of sample CCFs at different inter-station distances calculated using conventional cross-correlation and phase cross-correlation methods a) 0.49 km. b) 1.33 km, c) 2.02 km, and d) 3.44 km. Conventional cross-correlation are in the left panel and phase cross-correlation are on the right panel with inter-station distance of CCFs marked in the top right corner of plots on left panel. The y-axis of each plot is the correlation co-efficient of the CCFs.

The quality of CCFs varies with station location in terms of contamination by phases such as the persistent localised source located in Section 3.1.1, and variations in amplitude. We find stations located in the northern section of the array have lower quality CCFs than those in the central and southern section. Lower amplitude CCFs may be affected by poor coupling and higher attenuation of seismic waveforms. The damage zone surrounding the Alpine Fault may be leading to high attenuation of surface waves due to fractured rock (Toy et al., 2015). In CCFs from stations deployed on the stop bank along the western side of the Whataroa River, North of state highway 16, a sudden reduction in amplitude is observed [Figure 3.1

and 3.2]. This may be due to either high attenuation of the Alpine Fault or poor coupling of geophones as the stop-bank is made up of river gravels so is likely to be poorly consolidated. Determining the attenuation coefficient of the Alpine Fault damage zone and principle slip zone is beyond the scope of this thesis, and thus at this point the cause of this reduction in phase amplitude is unresolved.

Using back-projection of CCFs we are able to locate one persistent localised source, but some moveout plots show multiple phases of a similar character, likely to be from multiple sources [see Figure 3.3c]. A comparison of moveout plots [Figure 3.3b and c] and their back-projection [Figure 4.3a and b], shows we are able to confidently locate stationary source phases with a clean signal in both amplitude and signal-to-noise ratio [Figures 3.3b and 4.3a]. In plots with a lower quality signal, the back-projection is less clear, resulting in a scattered distribution of high amplitude locations [Figures 3.3c and 4.3b]. The assumption of a constant velocity is likely a contributing factor to only being able to locate one stationary source. The use of a 2D velocity model may improve our locations and enable us to identify more than one persistent localised source. We interpret the stationary source located in Figure 3.5 as rapids in the Whataroa River. We know from field observation during the deployment that rapids are located east of this site. The offset of the high amplitude anomaly is probably due to the assumption of a constant velocity; however we deem it accurate enough to make the interpretation.

In a spatially inhomogeneous noise distribution, such as the seismic ambient noise wave-field, a non-symmetric cross-correlation is expected (Shapiro and Campillo, 2004). We observe this in our CCFs to varying degrees with some CCFs showing strong asymmetry [Figure 3.3a] and others moderate symmetry [Figure 3.3f]. Outside of the array, signal predominantly originates from the northern side of the array with higher signal-to-noise ratio observed for southward-travelling signals than for northward-travelling signals [e.g. Figure 3.4d]. The persistent localised source, located in a back-projection, also serves as a source to stations located nearby, at

which both high signal-to-noise ratio and strong CCF asymmetry are observed [e.g. Figure 3.4a]. More localised bands of high signal-to-noise ratio are observed for virtual sources located in the northern section of the array [e.g. Figure 3.4b and c]. The localised bands of high signal-to-noise ratio are generally inline with the persistent localised source and virtual source location, with nearly all other stations having lower signal-to-noise ratios. This evidence suggests that the persistent localised source is acting as a strong source when located in the stationary phase zone of an individual CCF (Snieder, 2004). The presence of the persistent localised source in our data shows that the river can act as a ambient noise source within the frequency range we are using, and as such the many rivers in the local area are probably important contributors to the signals recorded (e.g. Burtin et al., 2010). The weak symmetry observed in our CCFs shows that we have reasonable azimuthal distribution of signal sources. Shapiro and Campillo (2004) and subsequent studies by other authors have shown robust observations can be made from partially diffuse ambient seismic noise and we believe our CCFs fit this criterion.

Figure 4.3: Comparison of CCFs back-projection for virtual source move-out plots in Figure 3.3b and c. a) back-projection of virtual source station 664, and b) back-projection of virtual source station 668. The Whataroa River outline is plotted in white.

4.1.2 Dispersion Curves

As discussed above, we only find one persistent localised source location despite move-out plots suggesting the presence of more. The large variation in group velocity dispersion curves is likely due to this as we cannot confidently remove all persistent localised source phases from CCF component(s) with only one location. For this reason, we have chosen to use the sub-array for further analysis [see Figure 3.8].

The presence of a strong velocity contrast has been shown to excite higher modes in the surface wave-field (Boaga et al., 2013). From the DFDP-2B borehole measurements, we know that the basement-sediment boundary is a sharp density contrast (Sutherland et al., 2015), suggesting a high velocity contrast and thus the potential for higher modes to be present in our CCFs. At inter-station distances less than 1000 m we confidently identify our dispersion curves as being fundamental mode Rayleigh waves as the spectral amplitude of the frequency-time analysis shows a clear high-energy anomaly associated with group velocity dispersion [Figure 3.7a and b]. At longer inter-station distances (1000-2000 m), several high amplitude bands form, suggesting higher modes may be present in CCFs with an inter-station distance greater than 1000 m [Figure 3.7c-f]. To explain the jump in group dispersion velocity between two amplitude bands in Figure 3.7f, we model synthetic dispersion curves using a two-layer input model with varying basement depth. We found a good match between fundamental mode synthetic and observed dispersion, which suggests the velocity jump in group velocity is due to the velocity contrast between the sedimentary layer and the basement.

Daily dispersion curves can be used as a proxy for dispersion curve reliability and error [Figure 4.4]. We find a good fit between daily and full-stack dispersion, particularly in station pairs with no jump in group velocity [see Figure 3.7], as observed in Figure 4.4c. Station pairs with a jump in group velocity show a small degree of

variation in daily dispersion curves at periods where the jump occurs [Figure 4.4a and b]. At periods where there is a jump, most of our dispersion curves have missing measurements where we cannot measure reliable dispersion. The start, end and width of this gap has small variations between each station pair daily dispersion curves. As we interpret some of the noise source to be from river sources, small fluctuations in the spectral content of dispersion around the jump may be due to changes in the spectral content of the river source due to water-levels controlled by rainfall and snowfall (Burtin et al., 2010). As the west coast has very high precipitation, this exerts a control on river water level (e.g. Henderson and Thompson, 1999), which in turn could control the excitement of signal from rapids (Burtin et al., 2010).

Figure 4.4: Daily dispersion curves of phase (red) and group (blue) velocity a) Station pair 643-783, b) 627-785, and c) 624-687. The yellow squares mark periods where a measurement is recorded in the full stack dispersion curve that is within criterion outlined in Section 3.2.

4.1.3 Tomography

We are able to recover most of the input checker-board model for most periods in both phase and group velocity, with exception of 0.3s period where poor recovery occurs [Figures 3.11 and 3.12]. The area resolved usually coincides with areas of high ray path density and azimuthal coverage for both velocity types [Figures 3.13 and 3.14]. We can resolve structure down to 200m in the centre of the valley for periods above 0.3s. At 0.3s period we expect a high degree of smearing in phase and group velocity. At periods above 0.3s, a small degree of north-west south-east

smearing is expected at most periods. The correlation of cells with high ray density and azimuthal coverage suggest a high degree of path crossing, essential for good resolution in tomographic models (Rawlinson and Sambridge, 2003). Due to the array geometry, we have more ray-paths travelling down or up the valley than across it. This suggests that while we have good correlation between ray-path density and azimuth, the angle between crossing paths is likely to be less than the desirable 90° . We believe this is the cause of the small degree of north-west south-east smearing that occurs in most of checkerboard tests. Through our checkerboard tests we conclude that we can resolve structures in the central and northern sections of the array for phase velocity tomographic models with resolution increasing with period. For group velocity, we can also resolve parts of the southern section of the study area at periods above 0.9s. In both models, we cannot resolve the edges which is consistent with the lack of ray path density and azimuth in Figures 3.13 and 3.14.

The limitation of straight ray tomography is it does not take reflection and diffraction into consideration when calculating travel-time (Rawlinson and Sambridge, 2003). Through the centre of the valley we consider the straight ray approximation acceptable as we expect most geology structure to be sub-horizontal. Along the edges of the valley, particularly near the DFDP-2B borehole, a sub vertical geologic structure may exist (Jenkins, 2017). In this area we consider the straight ray approximation to be inappropriate. This limitation is reflected in the velocity variation of the observed dispersion curves constructed from the surface wave velocity maps [see Section 2.3.4, Figure 3.21a-f]. Around the DFDP-2B borehole and north-west extent of the study area, larger uncertainty in the observed dispersion curve is observed compared to other areas [Figure 3.21a and d]. The lower values in the centre of the study area give us confidence that the straight ray approximation is acceptable in most of the study area [Figure 3.21b,c,e,f], with the exception of around the DFDP-2B borehole.

We find good agreement between phase and group velocity tomography models [Figures 3.15, 3.16 and 3.17]. The main observation of an increasing velocity at

longer periods is observed in both types of velocity. The lateral increase in velocity from the western side to the centre of the valley is also observed in both. One variation between the velocity types is the low velocity zone that forms at longer periods in the centre of the valley. For phase velocity, the anomaly is circular and does not touch the edge of the study area, while for group velocity a similar shape is observed but it covers the eastern edge of the study area. The difference in geometry is likely from a combination of different depth sensitivity between phase and group velocity (Bensen et al., 2007), and low resolution along the eastern edge of the study area [see Section 3.3.1].

4.1.4 Shear-wave Velocity Inversions

The neighbourhood algorithm used to invert dispersion curves for shear-wave velocity explores the parameter space defined a priori to the inversion to find the best fitting model (Wathelet, 2008). Therefore it is important to integrate prior knowledge into the parametrisation as the inversion process explores the full space. Search ranges that are too large will result in unstable results while those that are too narrow might converge on a local minimum, missing the best solution and resulting in poor misfit (Wathelet, 2008). The geologic units, including the basement depth, within the DFDP-2B borehole are well defined, however gravity modelling suggests the lateral distribution of the basement depth changes towards the centre of the valley, becoming deeper than the 240 m depth in the borehole (Jenkins, 2017). With this knowledge we include with a large depth range of 50 to 400 m. The shallow depths this study probes are difficult to parametrise as we expect strong lateral and vertical variations, so we allow a broad search range of all parameters in all layers. The 2-D compressional-wave velocity profile modelled by Lay et al. (2016) provides an estimate of shear-wave velocity although the V_p/V_s ratio is unknown so we choose to give compressional-wave velocity a broad range.

For the inversion of phase velocity dispersion, we use a two-layers over a half space [Table 4.1]. The simple model used in the phase inversion gives a well constrained

Table 4.1: Parametrisation of phase velocity inversion. We link a single compressional-wave (m/s) layer to shear-wave layer two and allow Poisson’s ratio (ν) to range between 0.2-0.5. Density is fixed at 2000 kg/m as it is the least important parameter (Wathelet, 2008).

Layer	Vs range (m/s)	Vp range (m/s)	Depth range (m)
1	150-1000		1-150
2	500-2500	500-5000	50-400
Halfspace	500-3000	500-5000	

model with low misfit values [Figure 4.5a]. In simple models such as our phase parametrisation, the use of constant compressional-wave, Poisson’s ratio (ν) and density is acceptable, except in the case where a strong velocity contrast is expected (Wathelet, 2008). From DFDP-2B and our dispersion curves and velocity maps we expect a strong velocity contrast so choose to parametrise a compressional-wave layer linked to the shear-wave layer of the velocity contrast for our phase inversions. We find that using one layer over a half-space to explain how the basement depth led to the inversion process converging to velocities that were unrealistic at shallow depths, so we add a shallow layer to allow low velocities to be fitted. Lay et al. (2016) observed very low (<500 m/s) compressional velocity in the upper 100 m of their model which justifies integrating a shallow low velocity layer. In our models, this shallow layer occurs at very low depths of 20 m and has a very low velocity in the north-west and south-east regions, whereas in the central model it occurs at 60 ms and has a higher velocity. The difference in velocity and depth suggests that the layer has been forced to fit the north-west and south-east models but is likely to be real in the central model where it is consistent with depths observed by Lay et al. (2016).

For the joint inversion we use four-layers over a half-space [Table 4.2]. Group velocity dispersion curves have a larger bandwidth than those of phase velocity, going up to periods of 1.8 s compared to 1.0 s for phase velocity so we can parametrise deeper layers. The simple parametrisation used in the phase inversion was inadequate for the joint inversion as large variations in layer thickness and velocity resulted in a poorly constrained model with low misfit. Instead for every shear-wave layer, we linked a compressional, ν and density layer to the shear-wave layer. We found the

Table 4.2: Parametrisation of joint inversion. We link compressional-wave, Poisson's ratio (ν) and density to each shear-wave layer and allow them a range of 200-5000 m/s, 0.2-0.5 and 2000 kg/m respectively. By linking compression-wave velocity, ν and density to each shear-wave velocity layer, we ensure the change in all parameters occurs at the same depth.

Layer	Vs range (m/s)	Depth range (m)
1	150-700	1-100
2	300-1500	50-400
3	300-2000	150-600
4	500-2500	400-1000
Halfspace	500-3000	

four layer model gave the best results in terms of misfit values and was reproducible for multiple inversions.

We observe lower misfit in the phase inversion compared to the joint inversion [Figure 4.5]. The reason for this is only one dispersion curve has to be fitted at each point compared to two for the joint inversion. To achieve reasonable misfit values in the joint inversion we required 500 iterations, as opposed to 50 for the phase inversion. A more complicated parametrisation in the joint inversion means a larger parameter space has to be explored, thus taking longer to converge to the best model. A similar relationship in the relative change in misfit values for each cell is observed in phase and joint inversions [Figure 4.5]. In general the highest values are located in the north-east and the lowest in the central study area. The difference in relative misfit between the two inversions is in the southern area where low misfit is observed in the joint inversion whereas a higher misfit is observed in the phase inversion. Areas of high misfit correlate well with uncertainty in the observed dispersion curve [Figure 3.21a and e].

Figure 4.5: Comparison of misfit (decimal percentage of velocity) for the best model a) phase velocity inversion, and b) joint velocity inversion.

The larger uncertainty in the phase velocity dispersion curves than the group velocity curves leads us to adopt the joint inversion as our proposed model for interpretation. By only using phase velocity we expose the inversion to large uncertainty in the 0.5 to 0.6 s bandwidth [e.g. Figure 3.20a]. Using a joint inversion with group velocities that have lower uncertainty [e.g. Figure 3.21d-f] allows us to have more confidence in the inversion result as it can be better constrained with two dispersion curves. The longer periods of group velocities also allow us to model greater depths than when using phase velocities alone.

Employing the method of Behr et al. (2011) to calculate weighted mean shear-wave profiles we can compute the standard deviation of acceptable models for depth and shear wave error [see Section 2.3.4]. Standard deviation of depth is within 30 m for the upper 400 m of the pseudo-3D velocity model [Figure 4.6]. Below this there is an increase to 50 m in the central part of the study area. Values remain fairly constant with depth along the eastern and western edges of the model [Figure 4.6a]. The values around 50 m observed in the central part of the study area extend the full extent of the north-south profile [Figure 4.6b]. It is likely that the standard deviation

of depth along the eastern and western extent of the study area are underestimated when compared to the central area. A higher-velocity anomaly is observed in the central area which means the inversion process may have taken longer to converge on an best-fitting model, leading to a larger variability of acceptable models with the misfit threshold we defined. The anomaly does not appear along the margins of the velocity model so the inversion has probably converged faster on the best fitting models here.

Depth profiles of the standard deviation of shear-wave velocity shows three distinct zones: the upper 300 m, 300 to 800 m and below 800 m [Figure 4.7]. The upper 300 m of the velocity model have a standard deviation of 50 m/s, although there is a distinctive band of higher values around 100 m depth [Figure 4.7]. Between 300 and 800 m, values of 100 to 150 m/s are observed along the full extent of the south-north profile but are isolated to the central part of the east-west profile. Below 800 m, the standard deviation increases to above 200 m/s for most of the model, except along the eastern and western edges.

Figure 4.6: Standard deviation of depth from the pseudo-3D shear-wave velocity model. See Figure 4.8 for profile locations.

Figure 4.7: Standard deviation of shear-wave velocity from the pseudo-3D shear-wave velocity model. See Figure 4.8 for profile locations.

4.2 Interpretation of Velocity Models

In this section we interpret Rayleigh wave tomography models of group and phase velocity, and then interpret the pseudo-3D shear wave velocity model constructed from the tomography models.

The steep change in velocity contours from low to high in the north-western section in both phase and group velocity maps is difficult to interpret as it occurs in a poorly sampled area [see Section 4.1.3]. The rapid change in velocity may be a transition in lithology from sedimentary to basement rock, as observed in DFDP-2B (Sutherland et al., 2015). As period increases, the observed change in velocity moves to the east which suggests the basement-sediment interface is deepening in this direction [Figure 4.8]. We suggest fundamental Rayleigh waves become sensitive to basement in the north-east at 0.7s period for phase velocity and at 1.1s period for group velocity [Figure 4.8]. At these periods the velocities at increasing periods appear to stabilise in the 0-250meter distance section of the profiles in Figure 4.8. The discrepancy in group velocity between the 1.1 and 1.3s period contour and the 1.5 and 1.7s period contour is likely an artefact [e.g. Figure 3.12].

Lay et al. (2016) observe a low-velocity zone in the upper basement within this area, but we are not confident enough in the tomography of this area (or depth) to suggest a link from the current results. The low-velocity zone in the central and eastern edge of phase and group models is interpreted to have deeper sedimentary cover than the north-eastern sector. The circular nature of the anomaly suggests it could be the thickest part of the overlying sedimentary sequence in the study area.

Figure 4.8: Velocity tracks at discrete periods from phase and group velocity maps. The line markings signify the period of each line in the left panel plots. The right panel shows a map of the track profile marked as NW-SE and the E-W and N-S markings show the track of the error profiles in Figure 4.6 and 4.7. The p-wave tomography model transect from (Lay et al., 2016) is marked by the red line and the seismic reflection line 2 transect from Lepine (2016) is marked by the blue line.

From the fence diagram of 2D profiles extracted from the pseudo-3D model we are able to identify structures within the study area [Figure 3.22]. We identify four main layers within the velocity model; three sedimentary layers and a basement layer. The first sedimentary layer was identified in Figure 3.21 as having a depth of

100 m and velocity range of 500 to 700 m/s. The equivalent layer is identified in the pseudo-3D velocity model as having a upper velocity of 720 m/s [Figure 4.12c]. We extract the isosurface of this velocity from the psuedo-3D velocity model and find the layer deepens from 100 m to nearly 120 m in the centre of the valley [Figure 4.9]. Sedimentary logging of the DFDP-2B borehole by Sutherland et al. (2015) found an alluvial layer of river gravels and sands deposited from 60 m depth to the surface. Below this layer lacustrine silty sand sediments are deposited down to the basement contact at 240 m. We interpret our 720 m/s depth isocontour as the contact between the silty sand and the gravels.

Figure 4.9: Smooth velocity contours of the 720 m/s isosurface, extracted from the psuedo-3D velocity model. The red contour lines are the isosurface in meters and the black contours are 50m topographic contours. The Whataroa River, road and DFDP-2B (yellow circle) are marked for reference. The isosurface is smoothed using a 200 m wide filter to remove short wave-length features.

The second sedimentary layer is interpreted to be the lacustrine silty sand sediments identified by Sutherland et al. (2015). The velocity range of this layer is 720 to 1100 m/s and its lower depth isocontour is plotted using a velocity of 1100 m/s [Figure 4.10a]. The isocontour shows a fairly constant gradient dipping to the west

across our study area. The shallowest point is to the south of DFDP-2B at 300 m and the deepest point occurs along the eastern edge at 400 m. We identify a third sedimentary layer that has velocity range of 1100 to 1250 m/s and a maximum depth of 700 m. The lateral distribution of the middle sedimentary layer is confined to the central part of the Whataroa Valley. Lepine (2016) observed a similar structure in a seismic reflection line where a sedimentary sequence similar to our third sedimentary layer thins to 25 m within the western flank of our study area [Figure 4.13a].

Figure 4.10: Smooth velocity contours of the a) 1100 and b) 1250 m/s isosurface, extracted from the pseudo-3D velocity model. The red contour lines are the isosurface in m and the black contours are 50 m topographic contours. The Whataroa River, road and DFDP-2B (yellow circle) are marked for reference. The isosurface is smoothed using a 200 m wide filter to remove short wave-length features.

Basement velocities of 1250-1650 m/s are observed down to 1000 m in most of our study area, except for the high velocity anomaly in the central area which reaches velocities of 2200 m/s. This anomaly is associated with an error of ± 150 m/s so is likely an over-estimate. Due to the high error, we choose to interpret the basement as a single layer.

We provide two interpretations of the depth to basement. The first uses the 1100 m/s velocity contour which separates the second sedimentary layer from the basement

layer, we call this interpretation one [Figure 4.10a]. The second uses the 1250 m/s contour which includes the third sedimentary layer, we call this interpretation two [Figure 4.10b]. These interpretation models are also presented in a schematic showing the interpreted basement depths [Figure 4.11]. Interpretation one shows a fairly constant gradient dipping to the west across our study area. The shallowest point is to the south of DFDP-2B at 275 m and the deepest point occurs along the western edge at 400 m. Interpretation two has a far steeper gradient than interpretation one, reaching 700 m depth along the south-western edge of the study area. The geometry of the surface appears to be bowl-like with steep sides [Figure 4.10b]. Both of our basement depth interpretations suggest a large amount of glacial erosion has taken place. Similar depths have been observed in other glacial valleys where a deep U shaped valley have been carved by the glaciation (e.g. Benn et al., 2014). Similar thickness has been observed in Tasman Glacier, where historical seismic reflection profiles suggest the glacier may have been up to 700 m thick (Kirkbride and Warren, 1999).

For each interpreted layer we are able to estimate depth and shear wave velocity error based on standard deviations calculated in Figure 4.6 and 4.7. The velocities of the first and second sedimentary layers are estimated to have an uncertainty of ± 50 m/s and depth uncertainty of ± 15 m. The third sedimentary layer has a higher uncertainty of ± 75 m/s and ± 25 m. The basement has large variability in the potential uncertainty as it includes the high velocity anomaly which has a high uncertainty. We chose to provide two uncertainty estimates for the basement, the first for the high velocity anomaly with velocities above 1600 m/s and the second the basement with velocity between 1250 and 1600 m/s. For the high velocity anomaly the uncertainty is ± 150 m/s and ± 30 m. The remainder of the basement has an uncertainty of ± 75 m/s and ± 20 m.

Figure 4.11: Schematic geological interpretation of the pseudo-3D shear-wave velocity model in Figure 3.22. The dashed blue line refers to basement depth interpretation one and the dashed red line refers to basement depth interpretation two. We do not show the first sedimentary layer as it is not visible in Figure 3.22.

4.3 Comparison to Previous Studies

Here we compare our velocity model and interpretations with previous studies of Jenkins (2017), Lay et al. (2016), Lepine (2016), and Sutherland et al. (2015). Overall, we find our shear-wave velocity model is in somewhat of an agreement with past studies.

As the only direct measurement of the subsurface within our study area, we first compare our results to DFDP-2B (Sutherland et al., 2015). The true basement depth at the borehole is 240 m, which is shallower than our velocity interpretations, which estimate it to be between 290 and 310 m based on the 1100 and 1250 m/s velocity contours [Figure 4.12a and b]. The borehole does not help us to determine the more reliable model of our two interpretations as both estimate the depth to be deeper than seen at the borehole. An interesting feature of the velocity profile through the borehole location is to trace the velocity contour at 240 m (at the borehole) in the velocity model across the profile. In Figure 4.12 this contour is approximately represented by the transition from the darkest red to the next darkest red. Tracing this contour along the profile shows a rapid decrease in depth just west of the borehole. To explain a steep Bouguer anomaly gradient between the borehole and the western hillside, Jenkins (2017) modelled a steeply dipping basement or a reverse fault at this location. As the rapid change in velocity occurs along a different contour we are not confident making this interpretation in our model. The uncertainty of our measurements in the area to the west of DFDP-2B is large so it is likely that we do not have the resolution to make such an observation.

Figure 4.12: Comparison of the shear-wave velocity model to the DFDP-2B core log. a) shear-wave velocity profiles from west to east and transect the DFDP-2B borehole, marked as a yellow line on the profile. b) A simplified DFDP-2B core log with the major units encountered. The transition from yellow to grey marks the sediment to basement transition.

As outlined in section 4.2, our second interpretation using the 1250 m/s contour is consistent with observations made by Lepine (2016). His interpretations suggests a two layered sedimentary model above the basement, with a steeply dipping basement down to 550 m. Unlike our model, the second layer is present across the entire profile. We believe the equivalent layer in our model, the third sedimentary layer, is below the 25 m resolution so it is unlikely that we can resolve it across our entire velocity model. Basement interpretation one is more consistent with the observations made by Jenkins (2017). The 1100 m/s contour suggests the basement is thickest in the centre of the valley and along the western edge of our study area. The Bouguer anomaly low calculated by Jenkins (2017) occurs in a similar location, although it extends to the west of our study area.

In comparing our results to previous studies we can evaluate the efficiency of ambient noise tomography in imaging a target geologic structure. For surveys using a similar sized array to this study, we recommend that ambient noise tomography be used

as a secondary method to model surface and shear wave velocities. For the target frequencies (>0.5 Hz for this arrays aperture), there can be no guarantee that there will be a reliable noise source and as such deploying an array with the aim of using ambient noise tomography as primary means of surveying should be considered high risk. Through our results we have shown that deploying a continuously recording array when performing a reflection seismology experiment should be considered as ambient noise tomography may be able to provide a reliable shear wave velocity model.

Figure 4.13: Figures from previous studies by Lepine (2016)(a), Jenkins (2017)(b) and Lay et al. (2016)(c). Sub-figure (a) is a seismic reflection and interpretation of line 2 from Lepine (2016), showing the basement (pink), overlain by fluvial (green) and marine (blue) sediments. The profile location is marked on Figure 4.8. Sub-figure (b) is a residual Bouguer gravity anomaly of the Whataroa Valley from Jenkins (2017). The residual gravity values are marked on the map. Sub-figure (c) is the final compressional-wave velocity model from Lay et al. (2016). The location of the estimated Alpine Fault surface trace is marked by a AF, the dashed lines mark the Alpine Fault damage zone, and the black line marks the DFDP-2B well trajectory. The black arrows mark the approximate location of our study area. The profile location is marked on Figure 4.8

5 Conclusions

Tomographic analysis of continuous recordings of ambient seismic noise obtained during the 2016 Whataroa Active Source Seismic Experiment (WASSE) allows us to draw the following conclusions:

- It is possible to construct a robust shear-wave velocity model using short (four days) recordings of ambient noise;
- We identify four layers in the shear-wave velocity model. Three corresponding to sedimentary layers and the deepest to a basement layer;
- The basement depth is interpreted from the shear-wave velocity models to be between 400 and 600 m deep in the centre of the Whataroa Valley.

The use of seismic ambient noise recordings to construct velocity models is typically associated with continental and regional studies. We show it is possible to use several days of continuous ambient noise data, recorded on 4.5 Hz active-source geophones, to perform similar analysis at a much smaller scale. An important component of the processing flow used to construct the cross-correlation functions was the use of the phase-weighted stack. Higher signal-to-noise ratio are achieved through use of this method compared to the common method of linear stacking. During this study, an alternate cross-correlation method known as phase cross-correlation has been tested. Despite not using phase cross-correlation extensively in this study, we show through comparison to regular cross-correlation that it shows promise, particularly in removing a peak around zero lag-time. Our testing of different methods has highlighted the importance of testing processing techniques for use with short-duration high frequency ambient noise datasets.

The preferred shear-wave velocity model contains three layers in the Quaternary sedimentary deposits overlying the basement. The first is constrained to the upper 100 m, with a gradual deepening towards the centre of the Whataroa Valley. It has an average velocity of 600 m/s and is inferred to correspond to alluvial deposits identified in the DFDP-2B borehole. A second sedimentary layer has a depth profile of 300 m near the DFDP-2B borehole, deepening to 400 m in the centre of the valley with an average velocity of 750 m/s. It is inferred to correspond to fluviolacustrine deposits identified in the DFDP-2B borehole. The third sedimentary layer is laterally isolated to the central part of the valley and has a velocity 1250 m/s, reaching a maximum depth of 600 m. The basement shows a range of velocities increasing with depth. In the central valley, the basement velocity exceeds 1600 m/s while along the valley edge it is between 1300 and 1600 m/s.

We have constructed interpreted basement isosurfaces based on the lower depth isosurface of the second and third sedimentary layers. We give a maximum depth to basement within the study area a depth range of 400 to 600 m, based on the two isosurfaces. Near the DFDP-2B borehole and western edge of our study area, we have a low degree of confidence in our basement depth estimate of 300 m. We link this to straight ray tomography method and suggest bent-ray tomography may resolve the surface-wave velocity of the area better.

5.1 Future Work

The dataset used in this thesis could be used for further analysis. We suggest:

- Calculating the orientations of the horizontal components to allow all nine components of the Green's function to be estimated. This would allow Love waves to be included in the analysis, providing further constraints on the shear-wave velocity inversion. It would also impose a stronger constraint on the excitation of higher modes.

- The cross-correlation functions provide some indication of attenuation across the Alpine Fault trace. To investigate this further, cross-correlation functions could be calculated with distant seismometers outside of the array. This would allow the wave-front to be tracked across the array (including the Alpine Fault) using eikonal tomography.
- To improve cross-correlation functions, Julian day 16 and 20 should be included by shifting the start time of each day by four hours. This would allow four full-days of data to be processed using the MSNoise package at its current capacity. The other option would be to edit MSNoise codes and allow for sub-daily cross-correlation functions to be computed.

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