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Seismic Travel Time Tomography of the Auckland Volcanic Field

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Master of Science in Geophysics

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Abstract

The Auckland Volcanic Field (AVF) is a monogenetic intraplate volcanic field located right beneath New Zealand’s biggest city, Auckland. Despite the potential hazard it imposes, much of the structure and driving mechanism in the AVF are remain to be known. A high resolution three dimensional geophysical model is important to better understanding of the volcanism in the AVF. However, past geophysical studies, in particular regarding the model of deeper structures are generally lack in resolution. This research aims to infer three dimensional geophysical structure of the Auckland Volcanic Field, with focus on the seismic velocity structure deduced from P-wave travel time tomography. For this purpose, the Fast Marching Method, a grid based eikonal solver, is implemented. Capable of consistently finding the first arrival travel time for specific source-receiver paths, this method has low computing cost, and is stable, and robust.

The first part of this research is to perform synthetic checker-board tests to assess the potential resolution that can be achieved. The size and geometry of the checker-board pattern is the key in inferring the scale and coverage of the resolvable features. We found that subsurface features on the order of 60 km can be resolved to depth reaching 300 km while smaller features on the order of 30 km can be resolved at least to 80 km depth. The second part of this study is to use real earthquake data in order to probe into the subsurface structure of the Auckland Volcanic Field. The adaptive stacking method is implemented to obtain the travel time residuals of 131 teleseismic events. The outcome shows structural heterogeneity in the AVF, consistent with the Junction Magnetic Anomaly associated with the Dun Mountain Ophiolite Belt that extends to at least 80 km depth.
Acknowledgments

First and foremost I would like to thank my supervisor Dr. Kasper van Wijk for showing me the direction to go, keeping me in check with the bigger picture surrounding the research, and at the same time helping me get my head around the problem at hand. His constant help in all aspect is greatly appreciated. My special thanks is also reserved for Professor Nick Rawlinson. Being the author of the codes we used, without him we would not have the means to do this particular research. I also thank him for his constructive input and consultation throughout the year. Next I would like to thank Paul Freeman for helping us in the coding side of the research, making it so much easier for me to progress through. Thank you also to Josiah, Jonathan, and James. It is always good to share progress and ideas. The weekly meeting earlier in the year certainly served its purpose. And finally to my best friend Hui I thank you for keeping me composed and rational during the completion of this research.
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Chapter 1

Introduction

New Zealand straddles the Pacific and Australian tectonic plates. In the North Island of New Zealand, the Pacific plate is being obliquely subducted under the Australian Plate at a rate of about 41 mm/yr \cite{Henry5431}. The boundary is marked by the Hikurangi Margin \cite{Johnson} which is oriented in a Northeast-Southwest strike, and is located southeast from the central East coast of the North Island. The adjoining tectonism around New Zealand gave rise to volcanic activity across north to central North Island, spurring a series of magmatic arcs, beginning roughly 20 million years ago \cite{Ballance}. Figure 1.1 shows a diagram of volcanism around a subduction zone, taken and adjusted from \cite{Stern}. The subducting slab sinks into the asthenosphere, and the water within the slab is released upwards \cite{Frisch}. Water reduces rock melting points, which then induces partial melting of the rocks \cite{Frisch}. This whole process drives the magma at the magmatic front or the magmatic arc \cite{Frisch} (Figure 1.1).

Further back from the plate margin, a different kind of magmatism is observed, termed back-arc volcanism or the back-arc basin. The cause for the back-arc volcanic system is akin to the sea floor spreading mechanism \cite{Uyeda, Frisch}. The crust is weakened by the heat from rising magma and forces related to the slab rollback (see below) which creates a split in the area, and thus forms a new lithosphere from the partial melting of rising asthenospheric material \cite{Frisch}. A prime example of back-arc basin volcanism in New Zealand is the Taupo Volcanic Zone (TVZ) \cite{Sherburn, Cole, Behr}; located behind the magmatic arc of the Pacific-Australian plate boundary (Figure 1.1). \cite{Heuret} illustrate that back-arc deformation is related to several forces, including the slab pull force, the pushing force from the absolute motion of the upper plate, the slab anchoring force, and the pressuring force of the mantle flow on the side of the slab normal to the
trench. According to them, the cause of the extensional stress can be described with different models: (1) upper plate motion controlled model, where the absolute motion of the upper plate relative to the more or less fixed trench is in the landward direction; (2) slab rollback model, where the subducting slab makes a seaward bending or rollback movement which creates a spontaneous pull in the slab direction to the coupled upper plate and thus an extensional back-arc deformation; and (3) mantle flow induced model, where the mantle flow and the associated forces are pushing the slab from the upper plate slide, inducing an extensional stress on to the back-arc region. Heuret and Lallemand (2005) conclude that the upper plate absolute motion relative to the force from the partly anchored slab is mostly controlling the back-arc deformation regime.

With the previous descriptions of magmatism, the proximity of the observed volcanism is adjacent to plate margins. A different category of volcanic activity can be found further inland, described as intraplate volcanism. Generally the term is used to imply volcanism that is so far away from plate boundaries that magma generation cannot realistically be associated with mid ocean ridges, subduction zones, or transform faults (Johnson, 1989). As such, different kinds of models are needed to describe the sources that drive the magmatism in an intraplate system.
CHAPTER 1. INTRODUCTION

One of the most classic cases for the intraplate driving mechanism is the hot-spot or mantle plume theory, exemplified by the Yellowstone hot-spot in western United States (Puskas and Smith 2009; Smith et al. 2009). A volcanic hot-spot is a location on the earth where the tectonic plate is moving past a stagnant (or at least moving at a slower rate than the plate) hot source region underneath, and therefore resulting in a focused area of volcanism with a virtually fixed source of heat from deep within the Earth (Wilson 1963; Smith et al. 2009). A model from the Yellowstone study showing a mantle plume is presented in Figure 1.2. There are several ideas as to how a hot-spots originate. Morgan (1971) proposes that they are manifestations of lower mantle convection. Malamud and Turcotte (1999) argued instead that they could be formed at the 660 km seismic discontinuity; i.e., the transition zone between the upper and lower mantle. Malamud and Turcotte (1999) do not rule out a lower mantle origin, or more precisely the D" layer, as they consider that hot-spots are caused by basal thermal boundary layer instabilities in the mantle convection.

There are alternatives to the hot spot/mantle plume model. Volcanism in the Newer Volcanics Province (NVP) intraplate system of South-Eastern Australia is argued to not be necessarily related to a deep mantle plume, but instead can be explained by for example asthenospheric upwelling (Otterloo et al. 2014). An underlying hot-spot is deemed unnecessary in the study of Otterloo et al. (2014) to explain the volcanism, as geochemically metasomatic agents can modify a heterogeneous mantle source, which may induce melting. Other studies have also inferred that volcanism in the NVP can be caused by the Tasman Fracture Zone (TFZ) (Lesti et al. 2008) to shallower edge-driven convection (Demidjuk et al. 2007). Lesti et al. (2008) argue that the TFZ has a role in the reactivation of transtensional structures and triggers the magmatism in the NVP. The transmission of transform shear stress from the TFZ onto the continental crust causes the reactivation of structures by means of internal deformation of the continental plate, and these reactivated structures may trigger batch melting of the upper mantle as well as providing paths for magma ascension (Lesti et al. 2008; Hove et al. 2017).

The edge-driven convection theory was introduced by King and Anderson (1995) to explain the presence of large continental igneous provinces and high temperature magmas at the margin of cratonic lithosphere. Physical and geometric properties of the lithosphere or rather the asymmetry between the thick, older, cratonic lithosphere and the thinner, younger lithosphere translate to a difference in heat flux (King and Anderson 1995). This may drive a small scale convection that promotes mantle upwelling from beneath the thicker lithosphere to the shallow mantle (Figure 1.3) (King and Anderson 1995). Consequently, this process is capable of bringing continuous melts that would explain the
Figure 1.2: Classic example of a velocity model interpreted as showing a mantle plume model from the Yellowstone field (marked as Y) western United States. Figure is from Smith et al. (2009). The low P wave perturbation can be related to higher temperatures and melts.

Demidjuk et al. (2007) suggest that this edge-driven origin may be the case for the NVP, as it is consistent with their analysis of surface wave velocity anomalies where they infer an apparent step in the structure of the lithosphere. Demidjuk et al. (2007) also ruled out the presence of a mantle plume in the NVP and prefer the edge-driven effect to explain the upwelling. This is related to their finding that the U-Th disequilibria in the alkali basalts require dynamic melting in the upwelling mantle, but its rate of upwelling is less than what is required for mantle plumes (Demidjuk et al., 2007).

With regards to our study, intraplate volcanism has been regarded to be occurring in the Auckland Volcanic Field (AVF) (Huang et al., 1997; Horspool et al., 2006; Lindsay and
In addition, it has been suggested that a mantle plume is not the driving force for magmatism in the AVF. Some of the arguments to support this idea can be found in Huang et al. (1997); Horspool et al. (2006); and Mcgee et al. (2011). Horspool et al. (2006) did a joint inversion of teleseismic receiver functions and surface waves phase velocities. 1-D profiles are produced from each individual seismic receiver. By interpolating between the various 1-D profiles in addition to the shear wave dispersion consideration, they produced a 2-D shear wave velocity model (Figure 1.4). Their study finds a region of low shear wave velocity in the mid crust and upper mantle, at $\sim 80$ km deep below the AVF. This is considered to be an area of partial melting which feeds the magmatism in the AVF. A mantle plume model would have had an anomaly pattern with shallow manifestation extending from deep (e.g., Smith et al. (2009) shown in Figure 1.2). Huang et al. (1997) and Mcgee et al. (2011) used a more geochemical approach based on major and trace elements, and isotope studies including Sr-Nd and U-Th. In their model analysis, they found that the upwelling rates beneath Auckland Volcanic Field are too low for a hot-spot or mantle plume.

Geophysical studies of the AVF are lacking in resolution in particular for the deeper structure (Horspool et al. 2006; Cassidy and Locke 2010). The Horspool et al. (2006) model of Figure 1.4 lacks the three dimensional perspective that might provide some influential information. And thus, the consensus of slow and local upwelling magmatism inferred from geochemical studies (Huang et al. 1997; Mcgee et al. 2011, 2013, 2015)
lacks the supporting evidence from geophysical studies. In order to take a step forward, we propose doing a seismic travel time tomography study using a Fast Marching Method (FMM) approach \cite{Sethian1996,SethianPopovici1999,Rawlinson2007,RawlinsonSambridge2003,RawlinsonSambridge2004,Kool2006,Rawlinsonetal2006}. Some questions remain whether the edge-driven convection is the mechanism that invokes the volcanism in the Auckland Volcanic Field. We wish to provide a three dimensional geophysical model of the AVF which hopefully will help in understanding the driving force behind the magmatism in the AVF, and provide the input model for geodynamic modeling \cite{Davies2011,DaviesRawlinson2014}.
Chapter 2

Geological and Methodological Backgrounds

2.1 Auckland Volcanic Field (AVF)

The evolution of the tectonic settings for the continental crust under Auckland can be explained as four major phases with episodes of subduction and rifting (Spörli and Eastwood, 1997): (1) It started 200 – 100 Ma with subduction along the margin of the Gondwana super-continent, which formed the Mesozoic basement. (2) Rifting then follows, separating New Zealand from the super-continent. (3) Renewed subduction proceeded 20 Ma, initiating a convergent setting to the east of Auckland. (4) Back-arc extensional rifting at the end of the Miocene took place as the plate boundary/subduction zone shifted to today’s location, putting Auckland in its current position.

This setting for Auckland, ~ 400 km west from an active subduction margin (Hikurangi trench) and ~ 200 km Northwest from Taupo Volcanic Zone (TVZ) an active volcanic arc (Kereszturi et al., 2013; Leonard et al., 2017), means that the Auckland Volcanic Field (AVF) is categorized as an intraplate volcanism. The AVF, located right underneath the city of Auckland is a monogenetic field with 50+ volcanic centers (Kereszturi et al., 2013; Hopkins et al., 2017; Leonard et al., 2017), and encompassing a total area of 360 km² (Molloy et al., 2009; Ashenden et al., 2011; Mcgee et al., 2013). A volcanic field is considered monogenetic when the eruptions are vented out at distinct locations (instead of from the same volcano) and are episodic with short periods (from days to years) of activity (Kereszturi et al., 2013).

The basement rock in Auckland (and in fact most of New Zealand) region is made up from Mesozoic meta-sedimentary terranes most commonly greywacke (Bradshaw, 1989).
Figure 2.1: Volcanic centers in the Auckland Volcanic Field. Colors show the age constraint of different centers. While the disparate patterns indicate the reliability of the inferred age. Figure taken from Leonard et al. (2017) having been modified from Kermode (1992a).
Auckland sedimentary rocks are predominantly composed of sequences of varying sandstones and mudstones of Waitemata group over Te Kuiti group, with ages ranging from Miocene to Eocene (Edbrooke et al., 1998; Shane et al., 2010). Drill log data taken from Mount Roskill in the Edbrooke et al. (1998) study also show the presence of siltstones, and organic materials along with volcanic deposits at the first 20 to 30 meters.

After establishing the fact that the AVF is an intraplate monogenetic field, an important consideration is to look at the products and overall history of eruption in the AVF. Basaltic intraplate volcanic fields developed in the Auckland area in the late Pliocene to Quaternary (Cook et al., 2005; Mcgee et al., 2013). The AVF being the youngest (Mcgee et al., 2013) and the closest to the Auckland CBD is differentiated to the older South Auckland Volcanic Field (Briggs et al., 1994; Cook et al., 2005). Eruption styles vary depending on the water to magma ratio, with two of the dominant types being magmatic and phreatomagmatic (Allen and Smith, 1994). The different types of eruptions are manifested physically in the form of varying volcanic centers. Under wet conditions, in the presence of sufficient ground or surface water for magma-water interaction, phreatomagmatic explosions occur, forming maar volcanoes and tuff rings; under dry conditions, the eruption style can be Strombolian to Hawaiian (magmatic), producing scoria cones, and once dry magmatic eruptions have been well established, effusive lava flows may also be present (Allen and Smith, 1994; Lindsay and Leonard, 2009; Hayward et al., 2011). These are of course exclusive cases of eruption styles and in the field, features from more than one eruption type may be observed in a single volcanic center (Allen and Smith, 1994; Lindsay and Leonard, 2009).

Recurrence rate or frequency of volcanic activity has not shown any patterns. A period of heightened activity or a flare-up period has been identified around 30 ka. During this time it is reckoned that there were at 6 volcanic centers erupting within a time window of only approximately 4 thousand years. Evidences for this flare-up period can be found from several case studies: Tephra-layer analysis indicates a simultaneous activity from multiple vents at 32 ± 2 ka (Molloy et al., 2009). $^{14}$C dating also shows a clustering of volcanic centers with age ranges between 28 to 33 ka (Lindsay and Leonard, 2009). More recently, high precision $^{40}$Ar/$^{39}$Ar dating point out that there are 6 eruptions in this flare-up period from 30 to 34 ka (Leonard et al., 2017). Comparatively, after this flare-up period we saw a long pause in activity and fewer eruptions but a trend towards larger eruptive volume (Ashenden et al., 2011; Kereszturi et al., 2013). This is illustrated by the Rangitoto eruptions, the youngest (Lindsay and Leonard, 2009; Leonard et al., 2017) and the largest in term of the erupted magma volume (Mcgee et al., 2011; Kereszturi et al., 2013).
Characteristics of the produced magma and volcanic rocks, mainly its geochemical signatures have also been studied. Volcanic rocks of Auckland are predominantly alkali basalts or basanites with some characteristics of HIMU i.e. high $\mu$ (where $\mu = \frac{^{238}U}{^{204}Pb}$) Ocean Island Basalts (Huang et al., 1997). This characteristics include high Ce/Pb, Nb/Ce, and U/Pb with low Nb/U and Sr and Nd isotope ratios. They also found that the initial Pb isotope ratios are inferred to be a young HIMU signature that is relatively unradiogenic compared to older more developed HIMU Ocean Island Basalts. Other types of magmatic rocks such as tholeiite, transitional basalt, and nephelinite are also observed less commonly (Allen and Smith, 1994; Cassidy and Locke, 2010).

Mcgee et al. (2013) expanded on Huang et al. (1997), concluded that three distinct mantle components have been identified in a single volcanic field (AVF). These are (1) fertile peridotite with similar Pb-isotope composition to the Pacific mid-ocean ridge basalt, (2) eclogite domains with HIMU-like isotope composition that are dispersed within fertile peridotite, and (3) slightly depleted subduction-metasomatized peridotitic lithospheric mantle.

Further conclusion has also been drawn related to the initial phase of the melting process. Melting starts in the garnet-bearing fertile asthenosphere greater than 80 km depth, with lithospheric sourced melts variably diluting the initial melts (Mcgee et al., 2013). Mcgee et al. (2013) also propose that there is a control in the size of the volcanic centers from the asthenospheric mantle dynamics, further signifying the need to study the deep asthenosphere-lithosphere interaction.

2.2 Methodology

The principal theory, mechanism, and the approaches applied behind the seismic travel time tomography study are to be presented in the following. The Fast Marching Method (FMM) approach (Sethian, 1996; Sethian and Popovici, 1999; Rawlinson, 2007; Rawlinson and Sambridge, 2003, 2004; Kool et al., 2006; Rawlinson et al., 2006c) is utilized for travel time prediction in the forward step. The inversion step meanwhile makes use of the subspace inversion scheme (Rawlinson, 2007; Rawlinson et al., 2006c).

FMM is a grid-based eikonal solver that is able to track the advancement of the wavefront in 3D media. There are several advantages of FMM in particular over the more conventional ray tracing method: (1) FMM is highly stable, robust, and "capable of computing travel times to all points of a velocity medium" (Kool et al., 2006, p. 254). (2) FMM will consistently find the first arrival (Vidale, 1988; Kool et al., 2006; Rawlinson et al., 2006c).
et al., 2006c). That is, the path from the source to receiver is guaranteed to be the path with minimum travel time. In ray tracing we cannot be sure whether the single arrival located is a first or in fact a later arrival (Rawlinson et al., 2006c). (3) With many points and many travel paths between them considered in the ray tracing method, computing cost may be high. FMM, on the other hand, can be very computationally efficient while maintaining the level of accuracy (Kool et al., 2006; Rawlinson et al., 2006c).

2.2.1 Fast Marching Method (FMM)

The Fast Marching Method (FMM) is first described in detail by Sethian (1996) as a scheme for solving the eikonal equation. The eikonal equation is a mathematical model derived from the wave equation which describes the wavefronts propagation within a given velocity model (Sethian, 1996; Fomel, 1997; Sethian and Popovici, 1999; Rawlinson and Sambridge, 2003, 2004; Kool et al., 2006). The eikonal equation can be expressed as:

\[ |\nabla T| = s(x), \]  

(2.1)

where \( T \) is the travel time, \( s \) is the slowness or inverse of velocity, \( x \) describes the position, and \( \nabla \) is the gradient operator. The expression describes that the magnitude of travel time gradient at any point along the wavefront is proportional to the slowness at that front (Fomel, 1997; Kool et al., 2006). Therefore, the link between seismic travel time tomography and the eikonal equation can be illustrated by taking the integral of slowness along the ray path \( l \) so that:

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

(2.2)

Which gives us the travel time from the source to a particular point \( x \). However, the obstacle in terms of solving this eikonal equation is that in heterogeneous media, we see discontinuities in the gradient caused by wavefront self-intersection, which means that the equation itself needs the gradient of travel time \( \nabla T \) to be defined (Rawlinson and Sambridge, 2003; Kool et al., 2006). Specifically, a complex velocity model may lead to multiple paths in terms of the first arriving wavefronts meaning self-intersecting between the wavefronts (Rawlinson and Sambridge, 2003; Kool et al., 2006). This complication can be overcome by introducing the "weak solutions" by means of enforcing an entropy condition (Sethian, 1996; Sethian and Popovici, 1999; Rawlinson and Sambridge, 2003). One analogy to an entropy condition is that of propagating flame (Sethian and Popovici, 1999). If the front is symbolized as the extent of a propagating flame with the region
at one side is burned and unburned at the other side, as such that at time $T$ the front is signified by the set of all points positioned at distance at $T$ from the source. Then the entropy condition weak solution may be expressed in a simple sense as "once a point burns, it stays burned" (Sethian and Popovici, 1999). This means we need to consider the direction of flow of information while trying to evaluate the gradient in Equation 2.1 such that information should propagates from smaller to larger values of $T$ (Sethian and Popovici, 1999; Kool et al., 2006). Therefore, to compute the arrival time of the front at a given node, we will only use nodes that are already passed by the wavefront. This can be called the upwind finite differences scheme (Kool et al., 2006).

The upwind schemes that are expressed in Sethian and Popovici (1999); Kool et al. (2006); Rawlinson et al. (2006c) are very much analogous to each other and can be expressed as:

$$ S_{ijk} = \frac{1}{2} \left[ \max(D_{i-j}^{+T}, -D_{i-j}^{-T}, 0)^2 + \max(D_{i-j}^{+T}, -D_{i-j}^{-T}, 0)^2 + \max(D_{i-j}^{+T}, -D_{i-j}^{-T}, 0)^2 \right]^{1/2} $$

(2.3)

Again $T$ is the travel time, and $(i, j, k)$ are the spherical grid increment variables in $(r, \theta, \phi)$. $S_{ijk}$ would be the slowness at $(i, j, k)$, while $D^-$ and $D^+$ are the forward and backward operators. Equation 2.3 defines how new travel time values can be calculated using adjacent grid points with known travel times (Kool et al., 2006). Where the only slight variation is $(x, y, z)$ would replace $(r, \theta, \phi)$ if a cartesian coordinate is used instead of spherical coordinate system.

Therefore essentially, the modus operandi of the Fast Marching Method is to solve Equation 2.3 by constructing the solution outward, beginning from the smallest value of $T(r, \theta, \phi)$ (Sethian and Popovici, 1999). In particular, this is achieved by implementing the narrow band approach (Sethian and Popovici, 1999; Kool et al., 2006; Rawlinson et al., 2006c). The concept of the narrow band is to have a set of points within a narrow band around the existing front (Figure 2.2). Approximately, this narrow band front will represent the wavefront of the first seismic arrival. We then propagate the front forward in an upwind-downwind manner, suspending the travel time values of those existing points, before including new points into the narrow band, until all points in the grid are included. In accordance to all of this, we shall distinguish grid points into three separate descriptions (Sethian and Popovici, 1999; Kool et al., 2006; Rawlinson et al., 2006c): alive points which have correct travel time values calculated using Equation 2.3 and lie upwind from the narrow band; close points which lie within the narrow band and have trial travel time values calculated using Equation 2.3 while considering alive points only;
and far away points which have no value of travel time calculated, and lie downwind from the narrow band. Specifically, the action is done by looping through the grid points (see Figure 2.2) (Sethian and Popovici, 1999; Kool et al., 2006; Rawlinson et al., 2006c):

1. Let’s say that a trial is a point in close with the minimum value of $T$.

2. Add point trial to alive and remove it from close.

3. Let all points in the upwind direction adjacent to trial that are not alive be tagged as close. Similarly in the downwind direction, if the point is in far, then remove from far and add to close. This creates the narrow band.

4. Using Equation 2.3, recalculate corresponding values for $T$ at all new points in close that are adjacent to the new alive point.

Figure 2.2 gives the illustrations to the approach. Mind that for simplification purposes this is shown in two dimensions, the actual progression shall be three dimensional. If a teleseismic source is used for the computation i.e. originating far and outside of the model, then the starting narrow band will be all points at the bottom of the local model. Whereas if we start from a local point source (sourced within the model), then the first alive point shall be the hypocentre (Figure 2.2b). Furthermore for a teleseismic source, we also need to consider the computation of travel time from the source to the base of the model, by assuming a spherically symmetric earth outside the model (Rawlinson et al., 2006c,a; Rawlinson, 2008b). The recalculation of $T$ in an upwind fashion guarantees the new values will not be smaller than any of the alive points, therefore we can always move outward by redefining the narrow band and reassigning adjacent points (Sethian and Popovici, 1999).

### 2.2.2 Subspace Inversion Routine

The inverse problem of finding the model parameters (e.g. seismic velocity) from the data observations (i.e. travel times) utilizes the subspace inversion method presented in Kennett et al. (1988). The approach is developed for tackling inversion problems with multiple parameter classes and large non-linear problems. It involves solving the linearized problem associated with the projection of the objective function approximation onto the subspace iteratively, where at each current model, the gradient vector of the misfit function is evaluated and separated into different components (Kennett et al., 1988; Rawlinson and Sambridge, 2003; Rawlinson et al., 2006c). Each component is then related to a single parameter type, and a new model is found by performing minimization
simultaneously over several search directions that span a subspace of the model space (Kennett et al., 1988; Rawlinson and Sambridge, 2003).

The inverse problem has the following objective function (Rawlinson et al., 2006c; Rawlinson, 2007):

$$S(m) = \frac{1}{2} [\Psi(m) + \epsilon \Phi(m) + \eta \Omega(m)]$$ (2.4)

Where $m$ is the vector representing the model parameters, while $\epsilon$ and $\eta$ are the damping and smoothing factor respectively. $\epsilon$ and $\eta$ will compromise between the data misfit, how much divergence from the reference model, and the model smoothness. Each is described by the terms $\Psi(m)$, $\Phi(m)$, and $\Omega(m)$ respectively. The data misfit $\Psi(m)$ is at minimum when the model travel times become closest to the observed travel times, and is defined as (Rawlinson, 2007):

$$\Psi(m) = (g(m) - d_{obs})^T C_d^{-1} (g(m) - d_{obs})$$ (2.5)

Where $g(m)$ is the travel times predicted by model $m$, and $d_{obs}$ is the observed travel time data. $C_d$ is the data covariance matrix. Since the assumption is that data errors are uncorrelated (no direct consequence in terms of errors between parameters), we can then define that $C_d = [\delta_{ij} (\sigma_d^j)^2]$ with $\sigma_d^j$ as the uncertainty at the $j^{th}$ data point.

The $\Phi(m)$ and $\Omega(m)$ denotations are regularization terms, introduced so that under-determined or mixed-determined condition can be overcome. In other words, regularization terms help tackle the problem where not all model parameters can be constrained sufficiently by the data alone (Rawlinson, 2007). The first out of these two terms, $\Phi(m)$ is defined as (Rawlinson, 2007):

$$\Phi(m) = (m - m_0)^T C_m^{-1} (m - m_0)$$ (2.6)

With $m$ as the solution model and the reference/starting model $m_0$. $\Phi(m)$ constrains $m$ to not differ too much from $m_0$. $C_m$ is the covariance matrix for a priori model whose values normally determined from a prior information. Again, data errors are assumed to be uncorrelated therefore $C_m = [\delta_{ij} (\sigma_m^j)^2]$ with $\sigma_m^j$ as the uncertainty of the $j^{th}$ model parameter of the initial model.

The second regularization term $\Omega(m)$ controls the adjustment between satisfying the data, and getting a smooth model with minimum structural variation (Rawlinson, 2007). It is defined as (Rawlinson, 2007):

$$\Omega(m) = m^T D^T D m$$ (2.7)
Where $\mathbf{D}$ is the second derivative operator making $\mathbf{Dm}$ the finite difference estimate of a specified spatial derivative. $\Omega(\mathbf{m})$ thus smooths the model and will reduce in size as the model becomes smoother.

The iterative approach to the inversion means that the subspace method provides the model perturbation $\delta \mathbf{m}_i$ for the new model $\mathbf{m}_{i+1} = \mathbf{m}_i + \delta \mathbf{m}_i$ at each iteration step (Rawlinson et al., 2006c). The appropriate number of iterations is when the change in $S(\mathbf{m})$ is deemed to be significantly small (converge) or in the less likely situation, the data is completely satisfied. The perturbation $\delta \mathbf{m}$ for the objective function expressed in Equation 2.4 is given in Rawlinson and Sambridge (2003); Rawlinson et al. (2006c) as:

$$
\delta \mathbf{m} = -\mathbf{A}[\mathbf{A}^T(\mathbf{G}^T \mathbf{C}_d^{-1} \mathbf{G} + \epsilon \mathbf{C}_m^{-1} + \eta \mathbf{D}^T \mathbf{D})\mathbf{A}]^{-1}\mathbf{A}^T \hat{\gamma} 
$$

(2.8)

Where $\hat{\gamma}$ is the gradient vector ($\gamma = \frac{\partial S}{\partial \mathbf{m}}$), and $\mathbf{G}$ is the Fréchet derivatives matrix. The Fréchet derivative simply describes the rate of change of travel times with respect to the model parameters ($\mathbf{G} = \frac{\partial \mathbf{g}}{\partial \mathbf{m}}$) (Rawlinson and Sambridge, 2003).

$\mathbf{A} = [\mathbf{a}^j]$ is the $M \times n$ projection matrix for $M$ number of unknowns and $n$-dimensional subspace. The basis vector $\mathbf{a}^j$ spans the $n$-dimensional subspace in which $S(\mathbf{m})$ is minimized (Rawlinson and Sambridge, 2003; Rawlinson et al., 2006c). It is based on the gradient vector of the model space $\mathbf{\gamma} = \mathbf{C}_m \hat{\mathbf{\gamma}}$, plus the model space Hessian $\mathbf{H} = \mathbf{C}_m \mathbf{\hat{H}}$, where $\mathbf{\hat{H}} = \frac{\partial^2 S}{\partial \mathbf{m}^2}$ (Rawlinson et al., 2006c).

As mentioned previously, the method shall perform minimization over several search directions. The first search direction would be the direction of the steepest ascent $\mathbf{a}^1 = \mathbf{\gamma}$. The remaining directions would all be defined by $\mathbf{a}^{j+1} = \mathbf{H} \mathbf{a}^j$ for ($j = 2, \ldots$). As such, the rate of change of $\mathbf{a}^j$ will govern $\mathbf{a}^{j+1}$ (Rawlinson et al., 2006c). Matrix $\mathbf{A}$ is then to be orthonormalized through singular value decomposition so that there will not be any linear dependence between disparate $\mathbf{a}^j$ (Rawlinson et al., 2006c). Once we have computed the orthonormalized projection matrix $\mathbf{A}$, we can get the model update $\delta \mathbf{m}$ rapidly through inversion of a comparably small $n \times n$ matrix (Rawlinson et al., 2006c).
Figure 2.2: Schematic layout of how travel time values are calculated using the upwind narrow band scheme of Fast Marching Method (FMM). (a) Equation 2.3 is used to solve for travel time $T$ until all grid points became alive (have their values correctly calculated). Close points in the narrow band have trials calculated travel time values, and far points have no value of travel time calculated yet. (b) Shows in particular how the narrow band is evolving from a local point source. The black dots are alive points, in the case of the left picture, it is the point source/hypocentre. Grey dots are close points constituting the narrow band, and white dots are far away points. The top diagram is extracted from Rawlinson et al. (2006c) and the bottom is taken from Kool et al. (2006).
Chapter 3

AVF Resolution: Checker-board Synthetic Test

In this chapter we will describe the procedure of performing inversion of synthetic seismic data tests on to the Auckland Volcanic Field. The main purpose for this type of test is to assess the resolution that we can expect when seismic tomography using real earthquake data is performed. The scale of the structure that we might resolve will become apparent in the wake of the completion of this test. The synthetic test applied here will be the checker-board test [Lévêque et al. 1993; Rawlinson and Spakman 2016].

The idea behind the checker-board test is to model the earth structure under the study area by a checker-board pattern of alternating high and low velocity fields. Then we compute the synthetic seismic wave arrival times at the stations from this model. The computed travel times will be our synthetic observed data. By inverting these synthetic data we aim to recover the checker-board pattern of the "true" model. Regions where the inversion results show a clear checker-board pattern suggest a good resolution for that particular location.

A vital component of information to be captured regarding the sources and receivers data for this synthetic test is their locations. And in the anticipation of seismic inversion study using real data, we shall consider hypocenters of genuine earthquakes catalog and the positions of existing seismic stations in the region. The earthquakes have been divided into local and teleseismic events, and illustrated comprehensively in Section 3.2. The earthquake data information is downloaded from the IRIS catalog using the FDSN web service through ObsPy (see Section 3.2). Station information is obtained in a similar way from the GeoNet database (see Section 3.1). The stations used here are the 18 stations that are active or have been active in the past ten years. We mainly focus on their particular coordinates and assume that data exist from the sources to all the receivers.
The FMTOMO (Fast Marching Tomography) package (Rawlinson, 2007) will be utilized to carry out the technical aspect of the tomographic inversion. The code works by computing the travel time from the earthquakes to the base of a predetermined 3D velocity model, assuming a spherically symmetric earth outside the boundary of the model as described by the ak135 earth velocity model (Rawlinson, 2007). The fast marching method is then executed in the forward step through the model to calculate the arrival time at the specified station locations. Subspace inversion is then implemented to obtain the velocity perturbation. The forward and inverse step are repeated for a number of iterations before a final fast marching implementation which calculates the travel times through the final model. The final velocity model which comes out of these steps will indicate the achievable resolution.

A more detailed step by step account of the code and the technicalities involved are well presented in Rawlinson (2007). In order to give better details regarding the construction of the tests, we will explain some relevant input parameters. Many of the more important ones are regarding the dimension and geometry of the initial checker-board model. Others include the parameters for the propagation grids, size of the subspace dimension, subspace inversion damping and smoothing factor, and whether noise is added or not when synthesizing the data. Most of the parameters are kept consistent between different trials, with the exception of those that are related to the initial model. And unless we use a different set of events then there should not be any change in relation to the source parameters. This is done so we do not unknowingly introduce any variable that might influence the final results. Nonetheless, it is noteworthy to describe what the various significant parameters are and what each value represents.

To start, the damping ($\epsilon$) and smoothing ($\eta$) factors are regulating parameters as described by Equation (2.4) in Section 2.2.2. These parameters can be tuned depending on the preference, whether to obtain a smooth solution model or for the solution model to not stray too much from the initial model (Rawlinson, 2008b). For all of our subsequent resolution tests, we specify these factors to be $\epsilon = 5$ and $\eta = 10$. These values mean that the multiplication of the factor to its respective term leads to more emphasis on finding a smooth model for our results based on the objective function (2.4). We can define the size of the $n$-dimensional subspace in which the objective function $S(m)$ is minimized (Equation 2.4 refer to Section 2.2.2). We kept a maximum number of 20, however in practice this may decrease during the run of the code. This reduction is determined from the orthonormalizing procedure using singular value decomposition. To replicate picking errors, Gaussian noise can be added to the synthesized dataset. However we did not add any noise to any of our trials of synthetic test in the attempt to simulate an optimal
scenario.

Probably the more significant parameters are those for the initial model. They will govern the extent of the 3D model but perhaps more importantly, the scale of subsurface features which the method can resolve. The model will be compartmented into grid points each with assigned velocity values. This velocity grid will be the main aspect of our model. In addition to the velocity grids, there are the propagation grids. The propagation grids are needed by the method as a means to sample the velocity fields for use by the eikonal solver, and the two grids need not to be related to each other (Rawlinson 2007). A good practice though is to have significantly more number of propagation grids than the velocity grids, so that it can regularly resample the velocity fields.

Later sections will explain the two approaches for the checker-board resolution test. One is to use a coarse grid spacing while the other takes more closely spaced velocity grids. We are to compare the results and analyze any advantage and disadvantage of using one over another. Further descriptions regarding the specifics of the test, including any important values that we use as the parameters will be addressed in each of the corresponding sections on the synthetic checker-board test.

### 3.1 Earthquake Inventory

Earthquake activity in New Zealand is monitored by a countrywide network of seismograph stations called GeoNet that is operated by GNS science. The station (meta) data are available through the International Federation of Digital Seismograph Networks (FDSN) web service. We access this information with ObsPy, a python based seismic processing and analyzing tool. In order to provide sufficient information for data download, we need to determine several criteria. One criterion is to specify a time window. We made it so that the time frame includes all seismograph stations that were active in the 10 years commencing from the 1st of January 2006 to the 31st of December 2016. The next content we need to designate is in relation with the space or position of the stations. There are two ways to establish this criterion. We can appoint the four corners corresponded to the geographical restriction of the desired area (i.e. the maximum-minimum latitude and longitude), and obtain all the listed stations within the limit of the region. Or, if we already know the name or ID of the particular stations that we think might provide useful informations to our study (for example through materials in the GeoNet website), then we can straightforwardly input those stations to be downloaded. In our case, we use the combination of these two approaches to download the informations for
the receivers. We initially specify a geographical restriction to obtain informations on all stations within the Auckland area, but then we specifically add complementary stations whose data are known to be available from the GeoNet website.

The procedure allowed us to obtain stations information from 18 different seismic receivers distributed around the North Island. Table 3.1 presents the detailed information of these stations in regards to their location coordinates and their associated networks. The locations of these stations are showed in Figure 3.1. Note that there are different symbols used to mark the stations in Figure 3.1. The particular reason to show them like this is to signify the contrast in terms of the station locations and array distribution, as well as to mark the type of different receiver instrumentations.
Figure 3.1: Location of the 18 stations considered in this study. (a) Magenta triangles mark the Auckland seismograph stations surrounding the Auckland center. Red and white circles mark the short period and broadband seismometers respectively from the New Zealand National Seismograph Network sites that form the outer ring around the densely spaced Auckland stations. (b) Zoomed-in view of the Auckland network array distribution given by the red box.
The receiver array can be categorized as those stations distributed more densely surrounding the Auckland center (magenta triangles, Figure 3.1a), and the seven outer stations (red and white circles, Figure 3.1a) forming a ring around the formerly mentioned central array. The outer ring stations are made up the New Zealand National Seismograph Network, and include four broadband seismometer stations (white circles, Figure 3.1a). The Auckland stations consists of receivers from the Auckland Volcano Seismic Networks. The majority of these stations are short period seismometers, except for MKAZ which is a broadband seismometer. All 18 stations as an inclusive network are considered in the resolution analysis of using synthetic checker-board test.
Table 3.1: Supplementary informations on the 18 stations considered in this study.

<table>
<thead>
<tr>
<th>GeoNet Stations Network</th>
<th>Auckland Volcano Seismic Network</th>
<th>NZ National Seismograph Network</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Station Name/ID</strong></td>
<td><strong>Description</strong></td>
<td><strong>Latitude</strong></td>
</tr>
<tr>
<td>ABAZ</td>
<td>Army Bay</td>
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</tr>
<tr>
<td>AWAZ</td>
<td>Awhitu Peninsula</td>
<td>$-37.06378^\circ$</td>
</tr>
<tr>
<td>EPAZ</td>
<td>Eden Park BICEP</td>
<td>$-36.875472^\circ$</td>
</tr>
<tr>
<td>HBAZ</td>
<td>Herne Bay Borehole</td>
<td>$-36.85013^\circ$</td>
</tr>
<tr>
<td>MBAZ</td>
<td>Motutapu Borehole</td>
<td>$-36.768787^\circ$</td>
</tr>
<tr>
<td>MKAZ</td>
<td>Moumoukai*</td>
<td>$-37.10413^\circ$</td>
</tr>
<tr>
<td>RVAZ</td>
<td>Riverhead Borehole</td>
<td>$-36.769996^\circ$</td>
</tr>
<tr>
<td>WIAZ</td>
<td>Waiheke Island</td>
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<td>Waiatarua</td>
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</tr>
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<td>KBAZ</td>
<td>Karaka Road Borehole</td>
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<tr>
<td>OUZ</td>
<td>Omahuta</td>
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</tr>
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<td>WCZ</td>
<td>Waipu Caves</td>
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<tr>
<td>GRZ</td>
<td>Great Barrier Island*</td>
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<td>Kuaotunu</td>
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<td>Tolley Road*</td>
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</tr>
<tr>
<td>HIZ</td>
<td>Hauiti*</td>
<td>$-38.51292897^\circ$</td>
</tr>
</tbody>
</table>

*broadband seismometers
3.2 Catalog

The earthquakes we use in this study can be differentiated into local and teleseismic events. Technically speaking in terms of the code specifications and requirements, local events are sourced within the model. Otherwise they shall be categorized as teleseismic events. However such is our construction of the model and event selection, we can simplify this definition into those events that are originated within and outside New Zealand. In this synthetic study we would like to gauge the maximum potential resolution. In addition to this, the fact that the most significant parameters in this test are those in relation to the earthquake and station locations means that we can afford to include a high number of earthquakes in the synthetic data test.

In order to retrieve the necessary earthquake information, we use the ObsPy FDSN web service to download the relevant data from the Incorporated Research Institutions for Seismology (IRIS) catalog (http://service.iris.edu). Similar to establishing the inventory, we need to define some restriction parameters. This includes the time window in which earthquakes are considered, magnitude, and geographical restriction, including depth of origin. Epicentral distance may also be considered as an indicator whether some events are to be discarded or not. The procedure of the test shall take the earthquake epicenters in order to generate synthetic arrival data.

3.2.1 Local Events

To decide what earthquakes are to be considered, for local events we limit our range of earthquake search to latitude between 34.6° S to 39.4° S, longitude between 172.6° E to 177.4° E, and depth to be less than 278 km. Those values shall also become the three dimensional extent of our synthetic model (Section 3.3 and Section 3.4). Eventually with real data tomography we would like to consider only earthquakes with good data quality. As such for global magnitude of completeness we would like to implement in the synthetic test a minimum magnitude of 4.0, knowing that any lower value is likely to include poor signal to noise ratio. On top of this, too many earthquakes can take a toll in computing time. To be consistent with the time range specified for the stations in Section 3.1 we define the earliest and the latest date for an earthquake to be considered to be also between the 1st of January 2006 and the 31st of December 2016. The constrain above allow us to obtain information for 681 earthquakes shown in Figure 3.2. The 681 local events made up of sources coming in from the Southeast general direction, more particularly, from the Hikurangi subduction zone.
3.2.2 Teleseismic Events

In terms of the global teleseismic events, instead of using a latitude-longitude-depth limit like we did for local events, we distinguish the sources by calculating their epicentral distance from a reference point. This is to avoid including events that are inside the P wave shadow zone from the location of the receivers. At the same time this would also exclude any local event while implementing deep earthquakes deeper than 600 km. Prior to calculating this distance, we collect global earthquake occurrences limiting the minimum magnitude to 6.0 and time window from the 1st of January 2011 to the 31st of December 2016. The magnitude is more restrictive than what we used for the local sources, to obtain data that we expect to have a high signal to noise ratio. Since events further away are attenuated more, it is reasonable to increase the minimum magnitude constraint to 6. Epicentral distances are then calculated to a reference point (37° S/175° E central to AVF). Those with epicenter between 19 to 93 degrees from the reference coordinate are selected as the teleseismic sources for the synthetic resolution test. The final total of teleseismic events included for this section of the research tallied up to 649 earthquakes (Figure 3.2).

The teleseismic sources complement the illumination from local sources by including incoming rays mostly from the Northwest of the receivers array (Figure 3.2). These sets of events total 1330 earthquakes surrounding New Zealand to form the coverage in illumination for a tomography study.

3.3 Synthetic Resolution Test with Coarse Checkerboard

We will first assess the potential resolution of AVF tomography by defining a checkerboard model that is coarse in spacing. This implies a single checker pattern that is rather large in size. To obtain the desired characteristic, we need to know how to define a model with FMTOMO. In FMTOMO, the model is defined by specifying the East-West, North-South, and top-bottom extremes. Then we need to input the number of grid points to fill between the limits of the model. This is done for each direction by indicating how many grid points should be spread over the two ends in that dimension. And thus, we make it so that the model would span from 34.5° S to 39.5° S in latitude and 172.5° E to 177.5° E longitudinally. The top of the model is set at 1.5 km elevation. The top needs to be above the surface because all the stations are required to be inside the model. The bottom of the model is at 278.5 km deep. Notice that the constraints that are set to
Figure 3.2: All the events selected for use in the synthetic resolution test. Color schemes scale for depth. (top) 649 teleseismic events with epicentral distance between 19° to 93° from the center of the receivers array. (bottom right) The same set of teleseismic events viewed with New Zealand at the center. (bottom left) 681 local events. Also shown here the locations of the receiver stations in magenta triangles and red squares.
categorized the earthquakes is in relation to the model specifics. We need to make sure that any event that is defined as local originated within the extent of our model.

We shall have 14 grid points from top to bottom. Laterally we have 25 grid points latitudinally and longitudinally each. The propagation grids are also appropriated with twice as many grid points i.e. 28 in depth, and 50 by 50 laterally. The values to be assigned into the velocity grids will follow a checker-board pattern of contrasting high and low velocities. These settings yield a three dimensional checker-board model where the size of each cube is roughly 60 km in depth, 62 km to the EW and 67 km in the NS, i.e. on the order of 60 km. The model encapsulates a significant area of the North Island, extending to a depth of 278.5 km. The 60 km scale also implies the size of the smallest feature resolvable in the inversion. Figure 3.3 depicts the initial coarse checker-board model that is used in the inversion process.
Figure 3.3: Checker-board resolution test using coarse grid spacing. The three cross-sections on the left hand side are the initial true model of the synthetic earth. The three cross-sections on the right hand side are recovered patterns from the resolution test. Cross-sections are taken at 90 km, 36.85° S, and 174.75° E. The red box signify the extend where we think 60 km features are resolvable.
Table 3.2: Parameters used in the two differing checker-board models. Model depth, NS, and EW refer to the extent of the model. Resolution refer to the smallest scale of features resolvable using that model. The dimension of the red boxes that indicate the extent of the region where resolutions are most exemplary are also noted.

<table>
<thead>
<tr>
<th></th>
<th>Coarse Checker-board</th>
<th>Fine Checker-board</th>
</tr>
</thead>
<tbody>
<tr>
<td>Model depth</td>
<td>278.5 km</td>
<td>280 km</td>
</tr>
<tr>
<td>Model NS</td>
<td>34.5° – 39.5° S</td>
<td>34.5° – 39.5° S</td>
</tr>
<tr>
<td>Model EW</td>
<td>172.5° – 177.5° E</td>
<td>172.5° – 177.5° E</td>
</tr>
<tr>
<td>Grid points in depth</td>
<td>14</td>
<td>28</td>
</tr>
<tr>
<td>Grid points in NS</td>
<td>25</td>
<td>50</td>
</tr>
<tr>
<td>Grid points in EW</td>
<td>25</td>
<td>50</td>
</tr>
<tr>
<td>Resolution</td>
<td>60 km</td>
<td>30 km</td>
</tr>
<tr>
<td>Recovery depth</td>
<td>278.5 km</td>
<td>80 km</td>
</tr>
<tr>
<td>Recovery NS</td>
<td>35.8° – 38.6° S</td>
<td>36.0° – 37.6° S</td>
</tr>
<tr>
<td>Recovery EW</td>
<td>174.0° – 176.5° E</td>
<td>174.2° – 176.2° E</td>
</tr>
</tbody>
</table>

The tomographic inversion took six iterations of travel time and velocity computation. Figure 3.3 shows the final result of checker-board recovery. Cross-sections are made at 90 km deep, 36.85° S latitude, and 174.75° E longitude. We see in Figure 3.3 that in the region indicated by the red box, the checker-board pattern can still be identified. The box covers much of our model, extending to the full depth of the model at 278.5 km, from 35.8° S to 38.6° S latitude, and from 174.0° E to 176.5° E longitude. This spans from Northland down to the Waikato region. Outside this box smearing artefacts become prominent. Within the box, we can resolve features with dimension correspond to the size of our checker-board pattern or larger. In other words, subsurface features on the order of 60 km can potentially be resolved in a seismic tomography experiment with real data.

3.4 Synthetic Resolution Test with Finer Sized Checker-board

Similar to the test using a coarse checker-board, we perform a synthetic resolution test using a finer sized checker-board. Also similar to the coarser version of the test, model extent are still uniform. However the number of grid points, as well as the propagation
grid nodes are doubled (Table 3.2). Increasing the number of grid points within the same extent of the model creates a more compact and densely spaced velocity grid that is smaller in size. This results in a test to see if smaller structures are recoverable.

The resulting checker-board model has an individual size of the rectangular cube of roughly 30 km in depth, 31 km in EW, and 33 km in NS. The scale of the resolvable feature is halved from the coarse checker-board model, to 30 km. Another six iterations of the travel time tomographic inversion are performed. Figure 3.4 shows the initial checker-board model and also the recovered final result. Cross-sections are made at 40 km depth, 36.85° S, and 174.75° E. Similarly we confine the region where checker-board pattern is recovered well in a red box (Figure 3.4), which reflects the extent where a sizable features can be resolved. The box spans the entire Auckland region from 36.0° S to 37.6° S latitude, and from 174.2° E to 176.2° E longitude, to 80 km in depth.
Figure 3.4: Checker-board resolution test using finer grid spacing. Similarly the cross-sections on the left are the initial true model of the synthetic earth, and the recovered patterns are shown by the three cross-sections on the right hand side. Cross-sections are taken at 40 km, 36.85° S, and 174.75° E. The red box signifies the extent where we think resolution is reasonable.
3.5 Data Misfit

We analyze the data misfit calculated by the inversion code. The misfit is expressed as the sum of the difference between the observed data and the predicted data. The observed data are those calculated by the synthetic model while the predicted data are calculated by model produced by the inversion. A small misfit means that the data are well accounted by the inverted model. Typically, a misfit analysis would use the $\chi^2$ measure that compare the data misfit in relation to the standard deviation or the uncertainty of the data. However, in this synthetic setting we did not introduce any noise, and therefore an analysis on the misfit rather than $\chi^2$ seems more appropriate.

We shall present the misfit as a progression after multiple iterations. We show this in Figure 3.5 with the misfit presented in milliseconds. The number of different panels came from the permutation based on the sources type used (local, teleseismic, or both) and the checker-board model size (coarse or fine). The numbers on the vertical axes show that if we use only the 681 local sources or only the 649 teleseismic sources, the misfit gets small enough to reach 20 ms in three out of the four cases (Figure 3.5). In the last case where we implement fine checker-board with teleseismic sources, the misfit is reduced to 50 ms.

An abnormality occurs as we jointly invert the local and teleseismic data. Data misfit deteriorate to the 1.7 s range even though we have shown that independently local and teleseismic sources produce low values of misfit. Ray paths for local and teleseismic sources also came from generally different direction so presumably it should not affect the data misfit much as they pass through different parts of our model. As an examination, we use a small sample of 10 events each from local and teleseismic catalog in the inversion. The misfit came out to pass the 100 ms mark using a fine checker-board model. The catch is that although this is reasonably small, compared to the returns of 600 plus events this is still larger, even though it uses considerably less number of sources.

This observation certainly poses a lingering question which becomes a section of the study where we can further dive into in future work. A possibility is that the misfit adds up from the sheer number of earthquakes used in the inversion. Added to this the various approximations that exist in the tomography which might not effectively create the perfect data fit. These are certainly very compelling to be investigated. However from our results and experience the checker-board recovery improves by involving joint inversion of local and teleseismic sources due to the sheer observational data and the encapsulating ray coverage. We also think that the recovery pattern of the joint inversion reflects the trends highlighted from independent local and teleseismic inversion. This is
overtly illustrated by the examples in Figure 3.6 where to the East, local events primarily contribute to the resolution, while similarly to the West and slightly in depth, teleseismic events provide clearer checker-board recovery. Therefore, we think that our results have enough of a solid foundation for us to have confidence in the inferred resolution.

3.6 Discussion

There are three points that are worth pointing out in this synthetic checker-board result. The first has already been mentioned, which is the smearing prominence outside the inferred red boxes. The second is the apparent trade-off between the resolvability and the region in which good recovery can be observed. Lastly, it is worth pondering about the general arrangement of the receivers and the circumstances that it brings.

We see in Figure 3.3 and Figure 3.4 that outside the range where we believe checker-board pattern has been recovered, the velocity anomalies may be elongated or stretched. The circumstance creates a merging of diagonally adjacent velocity perturbations with the same sign, forging an apparent dipping form. This is what we call a smearing artefact which can be a common complication in a checker-board resolution test. This peculiarity is an effect caused by insufficient data coverage leading to prevalent ray paths with similar orientation as represented by the smearing direction (Rawlinson et al., 2006a,c; Rawlinson and Spakman, 2016). With this there are not enough rays crossing each other in the grid cells associated to the smearing location since they have not converged enough yet. In accordance we did try to minimize the smearing effect by using plenty of events from all directions (Figure 3.2). However it seems that the assumption where teleseismic events come through the bottom interface of the model contribute to the situation where the ray paths are more uniform than we thought.

Fortunately in our situation smearing artefact is limited. In other tests not strictly aiming for resolution assessment, smearing can be a source of ambiguity. For example we can wonder whether a dipping feature is of a physical structure or is it a result from the smearing. There are however different implementation of synthetic tests developed to identify smearing instances. For example Bezada et al. (2014) suggested a 'synthetic squeezing test' to check whether a smear like anomalous feature observed in Atlas Mountains of Morocco is authentic or not. The test performs two inversions. The first inversion deliberately exclude the feature of interest. The resulting residuals are then used for the second inversion. If the data require the feature to exist, then it should be reproduced in the second inversion. Another synthetic test is that applied by Bijwaard et al. (1998) called 'layer cake' synthetic model. This test is especially good for assessing suspected
Figure 3.5: Progression of data misfit in millisecond at every iteration. The panels are separated by the checker-board model cube size and the type of sources used in the inversion. Separately, inversion of local and teleseismic sources produce misfits well into 50 millisecond range for both type of model. Joint inversion however deteriorate the data fit even though pattern recovery increase.
vertical smearing. By introducing separate blocks of distinct anomaly at a set depth intervals virtually mimicking the orientation of a smeared feature, synthetic data are calculated. If the inversion output shows that each block can be recovered individually, it indicates that smearing is not present in the inversion.

Considering the recovered patterns in the coarse and fine checker-boards (Figure 3.3 and Figure 3.4), it is clear to surmise that with increase in spatial resolution the model extent which encapsulate those recoverable resolution is diminished. We found that in our situation, the ratios between the size of individual grid cube and the commencing volume of the red box signifying the recovered area are similar for coarse and fine checker-board model. Following up from this, with the volume of the coarse checker-board cube being roughly eight times bigger than the more detailed fine checker-board model, the volume of the red box is also around eight time as big in the coarse model as in the fine checker-board model. Although for this relationship to be more empirically convincing we need more tests and in depth study. Qualitatively, the cause for this trade-off is because with
the decrease in the velocity grid size, there will be more occurrence where a grid would not have a sufficient number of rays passing through it. And still related to the smearing effect, this means only at shallow depths and confined enough location that ray paths are not unidirectional anymore and converging enough to recover the checker-board pattern.

A higher resolution project which uses even smaller checker-board size maybe feasible. However as pattern recovery coverage diminished with details, we predict that the result would be even more restricted to a smaller region. Furthermore the fast marching method will take some strains in computing time if we implement higher detail. In our case for the whole six iterations, computing time took a matter of hours to finish. In particular the fine checker-board model inversion took 9-10 hours. Therefore we think that due to the recoverability and computing time, the current approach might not be the most efficient strategy.

The arrangement of our receivers array plus the shape of New Zealand landmass means that the stations are not placed systematically. From Northland down to the Waikato region that is the coverage of our study, New Zealand is quite elongated in shape. Therefore it is only natural that seismic stations allocation lacks the width in array span and has a bias towards the north-south direction. Added to this is the fact that the stations are not evenly spaced, with denser array in the middle. An example seismic study where receivers array are wide and more or less evenly spaced, covering plentiful area of the study region is [Rawlinson et al. (2006a)] in south east Australia. The study is just generally better covered, over larger area, and uses more of the seismic stations in a compact configuration. However, having a systematic array configuration is not a prerequisite. Other seismic studies can have different approaches with commendable results, such as [Tkalčić et al. (2008); Jakovlev et al. (2011); Schmedes et al. (2012)].

We know for sure that better station coverage in numbers and range will improve the study. This includes the consideration of implementing some offshore stations. But for a start, introducing a station around 38° S/175° E seems like a justifiable suggestion. Looking at the checker-board result in Figure 3.4 the fact that there is no station around the coordinate creates a gap in the recovered pattern that could have meant an increase in the volume of decent coverage in checker-board recovery. We think that on top of improving the southern extent of the recovered volume, it might even improve depth extent to the 120 km depth as we can diminish the smearing effect in that region.

There seems to be little advantage in introducing stations closer to the center of the dense array as the Auckland Volcano Seismic Network already cover significant part of this area. As an overstatement for the purpose of reproducing the maximum resolution potential, we involve additional stations located within the extent of the Auckland
Volcano Seismic Network. We realize that realistically, for studies related to seismic tomography which uses New Zealand stations array, our synthetic result may not be a direct implication of the inversion resolutions due to this disparity in the utilized stations. However since the likeliest stations that would be omitted are clustered well inside the densely spaced Auckland Volcano Seismic Network, we understand that the disparity is small and can be tolerated.

Subsequently, it would be interesting to see what kind of results we might get by using wider coverage and more evenly spaced receivers that include offshore stations also. We do not think that our current situation is a major detriment in solving questions with regards to the AVF. Though we always want to know how we can improve on the resolution, and maybe this is one part where we have the potential to do so.
Chapter 4

Real Earthquake Data Travel Time Tomography

Next, we perform travel time tomography using real seismic data. The objective of our work in this chapter is to make the first attempt on performing tomographic inversion in the Auckland Volcanic Field, showing that the Fast Marching Method tomography has the potential to be applied there. We shall include an adaptive stacking method from Rawlinson and Kennett (2004) as an approach to help us obtain the travel time residuals. The extent of our study involves sampling three events from similar hypocentres. Using these events, by comparing the residuals of adaptive stacking results and manual pickings, we inspect whether the adaptive stacking can be used to automate arrival pickings. Once convinced by the robustness of the method, we perform the inversion for a number of events much less than the 1330 used in the synthetic test with high signal quality. The result shows a boundary between fast and slow velocity fields, aligned with the Junction Magnetic Anomaly.

4.1 Stations Configuration and Earthquake Waveforms Specification

Similar to our synthetic analysis, we obtain stations data through GeoNet FDSN web service using ObsPy. Once again we consider stations from the Auckland Volcano Seismic Network and the New Zealand National Seismograph Network (Table 3.1). This means the same total of 18 stations mentioned in the checker-board study are included (Figure 3.1). Only the waveform trace data to these stations from chosen earthquakes are to be downloaded. Bear in mind that it is possible that data is unavailable for a given source-
receiver pair. In addition, even if data are available, there is no guarantee that they have good signal quality. Therefore, not all earthquakes may have a trace corresponding to every station identified here (18).

The earthquake data themselves are retrieved in the same way through ObsPy from the IRIS catalog. On top of retrieving event locations, we also concerned ourselves with the associated waveforms recorded by GeoNet at each station. Waveform data are processed to improve the signal. This includes removing the instrument response and frequency filtering. Another step is to trim the data to make sure that all traces have the same number of samples over an equal time window. This is necessary for the adaptive stacking application described in Section 4.2. We make it so that the time window is ±20 s from the expected arrival time to the center of our stations array (37° S/175° E) calculated from the ak135 model. This is enough to include arrivals even for the stations at the edge of the array.

4.2 Adaptive Stacking

To help us determine the residual values from our arrivals, we utilize tcas, a code written by Nick Rawlinson to apply adaptive stacking (Rawlinson and Kennett, 2004). The adaptive stacking approach consists of two steps of trace alignment. The first initial alignment takes the predicted time from an earth model of an earthquake to a chosen reference station from the array. For every other stations, the predicted time to that station are then subtracted from the reference station time, producing the initial time shift for that particular station. Therefore, we get zero shift for the reference station, and a station further away than the reference station from the event should have a negative shift and vice versa. The aligned traces resulting from this step should already formed a close alignment. The stacked trace from this alignment then formed the reference trace. The second step is then to further improve the alignment through cross-correlation of the reference trace and the trace of each station. This is done iteratively, and the output should contain a measure of residual from the earth model that we used (Rawlinson and Kennett, 2004, 2017). Figure 4.1 shows an example of how tcas works. Starting from the raw data on the left panel, we first shift each traces based on their predicted time from ak135 1D-Earth model, the outcome of which is shown in the middle panel. The final panel of Figure 4.1 shows the final alignment after adaptive stacking approach has been performed. The stacked trace of the final alignment shown by the trace second from the top, should be a good representative waveforms across all stations for the "true" noise free trace (Rawlinson and Kennett, 2017).
4.3 Comparing Adaptive Stacking and Manual Pickings Using Events from Similar Hypocentres

Manually picking seismic arrivals can be a strenuous work, especially if it concerns a great number of events. Therefore the adaptive stacking approach can be a great help by removing the need to perform manual pickings (Rawlinson and Kennett [2004]). To evaluate the performance of the adaptive stacking code \textit{tcas} in comparison to work by manual picks, we consider three events with similar hypocentre. These events all should have similar ray paths and therefore residual patterns. So by doing this experiment we can inspect two behaviors. We address whether consistent output can be reached for sources that are expected to produce such output. This will confirm the robustness of the method. Secondly, an agreement in residual pattern between \textit{tcas} and manual picks would suggest that we can use the former to automate our pickings on many events.

We consider for this experiment three events from the Celebes Sea south of Philippine...
Figure 4.2: Three events with similar hypocentres used in the experiment. Specifically, event with ID 2876688 is located at 6.7403° N/123.327° E and 633.7 km deep; event 2876678 is at 6.7113° N/123.4876° E and 610.2 km deep; and event 2876682 is located at 6.4232° N/123.5796° E with 584.7 km depth.

(Figure 4.2). Figure 4.3 shows that the ray paths projected from these sources are indeed of uniform trajectory where we have basically a single path to each station from all three events. If you observe the lines in each panel of Figure 4.3, they represent a source-receiver path. And for all three events, the ray path to one station are overlapping with each other suggesting that they all travel through the same portion of the Earth. Furthermore, the waveforms for all three earthquakes show good quality signal with a lowpass filtering of 4 Hz applied. Therefore we think that these are good sample events for us to perform our experiment.

4.3.1 Manual Pickings Residuals

Results of the manual picks are shown in Figure 4.4 for all three events. The gap between the red and blue line in Figure 4.4 signify the differential time of predicted ak135 model (red) and observed arrival (blue). Since they all travel the same path, we expected the residuals to be similar at individual stations. The calculated residuals are compiled in Table 4.1 which also present the computed travel time expected from the ak135 model.
CHAPTER 4. REAL EARTHQUAKE DATA TRAVEL TIME TOMOGRAPHY

Figure 4.3: Ray paths projection indicating that the three events of Figure 4.2 all travel through the same path to reach the stations. (a) Top-down view of the ray paths. (b) Side on view looking on an East-West cross-section. (c) Side on view looking on a North-South cross-section. To reach one station, the ray path from all three events overlap with each other.

We recognize that the depth to our sources are not exactly equal therefore a two seconds difference between events might be observed at the travel time to a specific station (Table 4.1). However we think that this is a reasonable range in consideration, thus overall in terms of the values, the differences in residuals between events are fairly consistent.

4.3.2 Adaptive Stacking tcas Residuals

We input the same set of events (Figure 4.2) into tcas, aiming to replicate similar residual output compared to those found in Section 4.3.1. The final tcas alignment can be seen in Figure 4.5. The output associated with the residuals are compiled in Table 4.2 and is presented along with observed residual of Table 4.1 only with the mean subtracted for individual event.

Residuals from the reference 1D-Earth model computed by adaptive stacking approach are shown to be comparable to manual pickings. Figure 4.6 depicts this comparison using the values presented in Table 4.2. Each circle in the manual series represents
Figure 4.4: Plot of manual picks for all three events. Clock-wise from the top left panel are event ID 2876688, 2876678, and 2876682. Red lines represent the predicted arrivals using ak135 model whilst blue lines represent the observed arrivals at those stations.
Table 4.1: Expected travel time from ak135 model for all source-receiver pair regarding the three events south of Philippine. Also the residual calculated by subtracting predicted with observed arrival time.

<table>
<thead>
<tr>
<th></th>
<th>Predicted Travel Time (s)</th>
<th>Observed Residuals (s)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>2876688</td>
<td>2876678</td>
</tr>
<tr>
<td>ABAZ</td>
<td>625.217</td>
<td>624.345</td>
</tr>
<tr>
<td>ETAZ</td>
<td>626.813</td>
<td>625.947</td>
</tr>
<tr>
<td>GRZ</td>
<td>626.877</td>
<td>626.002</td>
</tr>
<tr>
<td>HBAZ</td>
<td>625.593</td>
<td>624.725</td>
</tr>
<tr>
<td>HIZ</td>
<td>631.696</td>
<td>630.853</td>
</tr>
<tr>
<td>KBAZ</td>
<td>627.108</td>
<td>626.244</td>
</tr>
<tr>
<td>KUZ</td>
<td>629.659</td>
<td>628.791</td>
</tr>
<tr>
<td>MKAZ</td>
<td>628.344</td>
<td>627.481</td>
</tr>
<tr>
<td>OUZ</td>
<td>614.956</td>
<td>614.062</td>
</tr>
<tr>
<td>RVAZ</td>
<td>624.653</td>
<td>623.783</td>
</tr>
<tr>
<td>TLZ</td>
<td>634.013</td>
<td>633.169</td>
</tr>
<tr>
<td>TOZ</td>
<td>631.897</td>
<td>631.044</td>
</tr>
<tr>
<td>WCZ</td>
<td>620.814</td>
<td>619.932</td>
</tr>
<tr>
<td>WTAZ</td>
<td>625.170</td>
<td>624.304</td>
</tr>
</tbody>
</table>

The mean removed observed residual at that station. These are close in relation to the residuals obtained from adaptive stacking which are shown by the orange circle. The fact that the orange circles show consistent output between three events of equal path indicates that the adaptive stacking is robust. On top of that, the comparison with our manual picks shows similar perturbation patterns in the stations. This suggest that the adaptive stacking has the capability to automate our pickings.
Figure 4.5: Final tcas alignment for the three events south of Philippine. Event id from left to right are 2876688, 2876678, and 2876682.

Table 4.2: Residual output from adaptive stacking tcas and manual pick observed residual with the mean subtracted.

<table>
<thead>
<tr>
<th></th>
<th>2876688</th>
<th>2876678</th>
<th>2876682</th>
<th>2876688</th>
<th>2876678</th>
<th>2876682</th>
</tr>
</thead>
<tbody>
<tr>
<td>ABAZ</td>
<td>0.096</td>
<td>0.195</td>
<td>0.369</td>
<td>0.350</td>
<td>0.350</td>
<td>0.300</td>
</tr>
<tr>
<td>ETAZ</td>
<td>0.048</td>
<td>0.067</td>
<td>0.068</td>
<td>0.000</td>
<td>-0.050</td>
<td>-0.100</td>
</tr>
<tr>
<td>GRZ</td>
<td>0.330</td>
<td>0.473</td>
<td>0.355</td>
<td>0.400</td>
<td>0.350</td>
<td>0.300</td>
</tr>
<tr>
<td>HBAZ</td>
<td>-0.027</td>
<td>-0.408</td>
<td>-1.208</td>
<td>-0.150</td>
<td>-0.150</td>
<td>-0.200</td>
</tr>
<tr>
<td>HIZ</td>
<td>-0.933</td>
<td>-0.969</td>
<td>-0.524</td>
<td>-0.950</td>
<td>-1.000</td>
<td>-1.100</td>
</tr>
<tr>
<td>KBAZ</td>
<td>0.023</td>
<td>-0.175</td>
<td>0.177</td>
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<td>-0.350</td>
<td>-0.400</td>
</tr>
<tr>
<td>KUZ</td>
<td>0.499</td>
<td>0.523</td>
<td>0.501</td>
<td>0.550</td>
<td>0.450</td>
<td>0.450</td>
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<tr>
<td>MKAZ</td>
<td>-0.108</td>
<td>0.006</td>
<td>-0.075</td>
<td>0.000</td>
<td>-0.050</td>
<td>-0.050</td>
</tr>
<tr>
<td>OUZ</td>
<td>-0.005</td>
<td>0.052</td>
<td>0.053</td>
<td>-0.050</td>
<td>-0.100</td>
<td>-0.150</td>
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<tr>
<td>RVAZ</td>
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<td>-0.005</td>
<td>0.000</td>
<td>0.000</td>
<td>-0.050</td>
</tr>
<tr>
<td>TLZ</td>
<td>0.156</td>
<td>0.123</td>
<td>0.277</td>
<td>0.300</td>
<td>0.200</td>
<td>0.100</td>
</tr>
<tr>
<td>TOZ</td>
<td>0.232</td>
<td>0.274</td>
<td>0.205</td>
<td>0.300</td>
<td>0.200</td>
<td>0.200</td>
</tr>
<tr>
<td>WCZ</td>
<td>0.041</td>
<td>0.034</td>
<td>0.059</td>
<td>0.050</td>
<td>0.000</td>
<td>-0.100</td>
</tr>
<tr>
<td>WTAZ</td>
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<td>-0.066</td>
<td>-0.252</td>
<td>-0.200</td>
<td>-0.300</td>
<td>-0.300</td>
</tr>
</tbody>
</table>
Figure 4.6: Plot showing the agreement between residuals from manual picks and tcas. The pattern of residual from different event of similar path shows consistency which indicates that the method is robust. The events used are those from Figure 4.2 with values from Table 4.2.
4.4 Inversion of Teleseismic Data

Confident on the adaptive stacking approach, we proceed with adding more events to our source catalog. The plan is to use this extended catalog in the travel time tomographic inversion with Fast Marching Method. The adaptive stacking code tcas is used to gain the input travel time residual information for the inversion, once we are convinced that the final traces are aligned correctly. Residual values as perceived at every stations are then plotted on a map, superimposed for all events. This will provide a preliminary output to suggest prevalent fast and slow velocity trends. Inversion results can then improve on this and give more detailed and precise information on the subsurface velocity pattern.

The utilized earthquakes catalog consist of 131 events, ranging from the Sumatra subduction zone to the Solomon Islands (Figure 4.7). Data are downloaded from the IRIS catalog using ObsPy FDSN web service. Initially, the earthquakes are confined between $15^\circ$ N to $60^\circ$ S and $90^\circ$ E to $180^\circ$ E. But to avoid triplication of the P-wave arrival due to the variation in the take off angle from the point source, any event closer than 27 degrees epicentrally from our array is discarded. Source depths are all deeper than 10 km, and the minimum magnitude is 6.5. The time window of the earliest to the latest event is between the 1st of January 2006 to the 31st of December 2016. Earthquake waveforms are lowpass filtered below 2 Hz.

All 131 events are then fed to the adaptive stacking code tcas. We found that the final alignment for all events are satisfactory. To gauge the underlying velocity trend, the output residuals from all events are plotted on a map with respect to the values at every stations (Figure 4.8). The map shows that there is a prevalent slow region to the east of the AVF.

For inversion, we assume a data uncertainty of 0.1 s. This is considerably larger than the error estimate computed by tcas. However based on our estimation from the manual picks and the different instrumentation of receivers, we think this is a reasonable approach. The model is specified to have velocity grid cells spread roughly every 10 km in depth, and every $0.1^\circ$ in both latitude and longitude, creating a similar checker-board pattern with 30 km resolution as in Section 3.4. Regularization terms are set with $\epsilon$ equals to 5 and $\eta$ equals to 10. These values correspond to the damping and smoothing factor respectively, presented in Equation 2.4.

The inversion result using input residuals from tcas is shown in Figure 4.9. The output velocity perturbation follows the preliminary suggestion of slow to the East, showing with more precision the split from positive to negative anomaly. Laterally, the boundary between fast and slow anomaly is oriented with a Northwest-Southeast strike. This
Figure 4.7: One hundred plus teleseismic events considered in the tomographic inversion.

feature extends to the depth of 80 km as shown by the East-West and North-South cross-sections. In the East-West cross-section, we can still see the split pass 100 km. But this is beyond the confinement of the box we defined from the resolution test and the smearing effect may well contribute to the output at this depth, which makes us less confident of its accuracy.

This observation is in line with result from Ensing et al. (2017) who use ambient seismic noise generated by the ocean to construct crustal shear wave velocity profiles. Their result points to the existence of a slow (shear) velocity field to the East which they associated to the Junction Magnetic Anomaly signifying the Dun Mountain Ophiolite Belt striking Northwest-Southeast, also perceived in Eccles et al. (2005) from aeromagnetic data. Ensing et al. (2017) see this boundary down to the 22 km depth which is the extent of their model depth. The result we presented suggests that this pattern of velocity anomaly may extend to greater depth than 80 km.
Figure 4.8: Residual map showing preliminary velocity field trend. Red signify a positive arrival that is larger than an average value thus a slow field (and vice versa for blue). Size of the circle indicates how big is the calculated residual. We show in green dashed line the estimate of an apparent boundary between fast and slow velocity.
Figure 4.9: Inversion result using 131 teleseismic events. The red box confine the region where we are the most confident in the result as inferred from checker-board resolution test. Cross-sections are made at 35 km, 36.75° S, and 174.75° E. Velocity field pattern shows slowness to the east of the Auckland Volcanic Field parallel to a magnetic feature from the Dun Mountain Ophiolite Belt.
Chapter 5

Conclusions and Future Works

In this study we advance toward providing the first high resolution three dimensional geophysical model of the Auckland Volcanic Field. Such a model will help us understand more about the AVF and other intraplate volcanic fields around the world. We do so by implementing the Fast Marching Method in a P-wave travel time seismic tomography study.

Our first step towards the goal is to understand the feasible resolution that can be achieved by implementing this approach. Our coarse checker-board synthetic test result implies that for a $60 \times 60 \times 60$ km$^3$ checker size, we can resolve features of the same volume to more than 200 km depth, from Northland to Waikato. And for a trade off in penetration depth and coverage of resolvability, our fine checker-board synthetic test shows that even smaller features 30 km in scale can be resolved. To improve from this inferred recoverability we can start by introducing a station near the $38^\circ$ S/$175^\circ$ E coordinate. Then we might consider adding even more stations to have better overall coverage, possibly include even offshore stations.

With our new grasp on the resolution that we can expect from a travel time tomography implementation, we proceed to perform tomography experiment using real seismic data. The adaptive stacking approach is used to help us find the travel time residuals automatically. Our real data tomography study shows a jump near the AVF from a fast to slow velocity anomaly. The strike of the boundary is in the Northwest-Southeast general direction, and it appears aligned with the Dun Mountain Ophiolite Belt, identified as the Junction Magnetic Anomaly. This finding is in agreement with ambient noise study of Ensing et al. (2017).

We offer the groundwork for seismic tomography studies in the Auckland Volcanic Field. There is still potential for future studies to build on this work. In terms of improving on resolution capability especially in range and penetration depth, we can look at the
station configuration. We can expand this for example by experimenting with station configuration, assessing the resulting checker-board recovery for different arrangements. Then we can start to assess the possibility of reducing the smallest resolvable features. In terms of the real data tomography study, the ensuing development should be to build upon seismic tomography result adding many local events to the teleseismic sources. The first step towards this is to proceed with obtaining solid inversion results for many local events to incorporate illumination from the direction not covered by ray paths from teleseismic events in our results. One proposal is to replace the ak135 1D-Earth model as the source for reference time information, opting instead for a more sophisticated multiple steps inversion. The idea is to develop our own reference model by using a coarse model spacing, and then use the predicted time from the final model as a substitute for ak135 prediction. The latter should use a model with finer grid spacing, and the reference time should then be more representative of our field.

Ultimately, we would like to strive toward performing a geodynamic modeling to explain the volcanism in the Auckland Volcanic Field. The thinking is to analyze the flow and melting system, assessing whether edge-driven convection is a possible mechanism to drive the magmatism in the AVF. But before that, we need a clear understanding of the crust and upper mantle interaction. A good resolution structural model will certainly provide important information for this purpose.


Appendices
Appendix A

Station Polarity Reversal

During the progression of our research we found that several stations have a reversal in polarity. We imply to the direction of the seismograph commencing the first arrival break. This is seen on the trace whether first break is a peak or a trough. A consistent array should produce traces where all first break point to the same direction. It is likely that this reversal has been addressed on the field which lead to the shift to a consistent polarity after a point in time. However this has not been transfered to the meta data information available as they still show a disparity. At station HIZ, we notice that at some time in 2014 polarity reversal has been reversed back and fixed. Similarly at station OUZ, we see a fix to normal polarity at some time in 2009. We also know that station KBAZ and WTAZ have polarity reversed throughout the time window of our study.

This polarity consistency is significant in the adaptive stacking step. As an example, we show in Figure A.1 the stacking alignment using traces of consistent polarity records and another alignment with polarity reversed for station OUZ. It can be observed that the method could not find the right alignment across the array when a trace shows an inconsistent polarity. This then leads to a big deficiency for that station in terms of residuals time. This may results in a fast arrival being inferred as slow, which then of course influence the inversion results.
Figure A.1: Influence of polarity reversal to the adaptive stacking alignment. In the example on the left panel, station OUZ display a polarity reversal which makes the corresponding trace do not aligned correctly with other stations. Once this issue is addressed, we see the correct final alignment displayed on the right panel.