Implications for intraplate volcanism and back-arc deformation in northwestern New Zealand, from joint inversion of receiver functions and surface waves

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SUMMARY

We employ a joint inversion of teleseismic receiver functions and surface wave phase velocities to determine the shear wave velocity structure in the crust and upper mantle beneath northwestern New Zealand. Receiver functions primarily contain information on velocity contrasts, while surface waves are sensitive to the average shear velocity with depth. By performing a joint inversion we reduce the limitations of each method, resulting in a more robust shear wave model. Inversion results reveal regions of low shear wave velocity of \( \sim 2.8 \text{ km s}^{-1} \) in the mid-crust (10–19 km depth) and \( \sim 4.0 \text{ km s}^{-1} \) in the upper mantle (70–90 km depth) beneath Quaternary intraplate basalt fields. We infer that the mid-crustal low-velocity zones (LVZs) are bodies of partial melt, most likely rhyolite intrusions. We suggest that the upper mantle LVZ is caused by the melt-producing regions of the upper mantle and is a source for the basalts of the Auckland Volcanic Field. This is in agreement with models for a shallow upper mantle source rather than a deep-seated mantle plume for the Auckland volcanism.

Average shear velocities for the upper crust are 3.4–3.6 km s\(^{-1}\), increasing to 3.6–4.0 km s\(^{-1}\) in the lower crust. The Moho is interpreted to be 29 \( \pm \) 1 km deep in the southern end of the array, shallowing to 26 \( \pm \) 1 km towards the edge of the continental shelf of northern New Zealand. Low upper mantle shear velocities of 4.2 \( \pm \) 0.1 km s\(^{-1}\) are observed, and are thought to represent a small percentage of partial melt in the upper mantle beneath this back-arc region. These low velocities extend to greater depth (~90 km) beneath the southern stations, indicating that mantle deformation associated with the mantle wedge to the south may extend as far north as the southern end of our array. Near-surface low-velocity layers extend to 6 km depth beneath a station at the northern tip of New Zealand, likely representing the expression of a thick sequence of sediments from the Northland Allochthon and the more recent Northland Basin.

Key words: Auckland volcanic field, crustal structure, joint inversion, New Zealand, Northland, North Island, receiver functions, surface waves.

1 INTRODUCTION

Intraplate volcanism occurs throughout northwestern New Zealand yet its source has remained relatively unexplained. One third of New Zealand’s population lives in Auckland city, which is situated on the active Auckland Volcanic Field; the field poses a potential hazard to the Auckland region and a future eruption would cause major disruptions to the society. The region occupies a unique tectonic setting, straddling the transition from back-arc extension to a ‘normal’ back-arc setting on the continental landmass of New Zealand. Our knowledge about the volcanism and back-arc deformation processes occurring in northwestern New Zealand can be extended by using seismological methods to investigate the structure of the region.

Receiver function analysis of teleseismic earthquakes recorded on three-component seismic stations has become a standard tool for crust and upper mantle investigators. It has been successfully employed to invert for the 1-D shear wave velocity structure beneath stations (Langston 1979; Owen & Zandt 1985; Ammon et al. 1990; Ammon 1991; Cassidy 1992), to derive elaborate images of upper mantle discontinuities (Gurrola et al. 1994; Li et al. 2000) and to calculate Moho depth and \( V_p/V_s \) ratios for the crust (Zandt & Ammon 1995; Zhu & Kanamori 2000). Receiver function inversion is inherently non-unique, caused by a velocity-depth trade-off between the thickness and the shear wave velocity of a given layer. The non-uniqueness is apparent when a thick fast layer can cause a similar response (in time) to a thin slow layer (Ammon et al. 1990; Ozalaybey et al. 1997). Ammon et al. (1990) illustrated that...
caution must be taken when inverting receiver functions for velocity structure, as there is a high dependence on the initial starting model and the use of available a priori constraints should be mandatory. In recent work, there has been a major drive to combine receiver function data with additional data sets such as seismic reflection and refraction data (Eaton & Cassidy 1996; Baumont et al. 2001; Galve et al. 2002) and surface wave data (Ozalaybey et al. 1997; Julia et al. 2000; An & Assumpcao 2004; Chang et al. 2004; Lawrence & Wiens 2004). Surface waves are particularly useful when combined with receiver functions, as they are also dependent on the shear wave structure, but they are complementary, in that they are sensitive to the average shear wave velocity structure with depth rather than to velocity contrasts. In comparison, receiver functions cannot constrain absolute velocities but they provide a wealth of information on the fine-scale velocity contrasts (~1 km). A joint inversion involving both surface wave dispersion and receiver function data provides a better-constrained shear wave velocity model. This joint approach has long been used by active source seismic investigations, where the seismic refraction data constrain the absolute velocity model, while seismic reflection data provide the detailed structural image (Pavlis 2003).

1.1 Geologic and tectonic background

Northwestern New Zealand (Fig. 1) is a region that has been somewhat neglected as a target by crustal scale geophysical experiments, possibly because at present it is tectonically quiescent, situated in a behind-arc setting. Nonetheless, northwestern New Zealand has undergone a complex tectonic history in the past 80 Ma. The region lies behind the active arc of the Hikurangi Margin on a peninsula running perpendicular to the Hikurangi subduction zone (Fig. 1). The Hikurangi Margin is characterized by oblique subduction of the Pacific Plate beneath the Australian Plate (Walcott 1978). Relative plate motion vectors between the Australian and Pacific plates decrease from 47 mm yr\(^{-1}\) in the north to 37 mm yr\(^{-1}\) in the south, reflecting the oblique nature of the New Zealand plate boundary (DeMets et al. 1990, 1994). In the back-arc of the subduction zone, the Taupo Volcanic Zone defines the eastern boundary of the Central Volcanic Region, the zone of active extension in North Island (Stern 1985).

Here, we refer to northwestern New Zealand as the region to the northwest of the Central Volcanic Region. It is manifested by extensional faulting inherited during rifting from Gondawana 80 to 60 Ma (Sporli 1989). Basement geology consists of Mesozoic greywacke that is overlain by Cretaceous to Oligocene sediments, obducted during the Oligocene as the Northland Allochthon (Brothers & Delaloye 1982; Herzer & Isaac 1992; Isaac 1996; Rait 2000). Miocene-age sediments are present in the southern part of Northland (Edbrooke 2001). Along the west coast of Northland Tertiary andesite volcanoes are buried, defining the volcanic arc during the Miocene (Challis 1978; Brothers 1984, 1986; Smith et al. 1989). During the Miocene, a subduction zone was located off the east coast.
of Northland, oriented parallel to the peninsula with a southwest dipping slab (Hayward 1987; Smith et al. 1989; Herzer 1995; Hayward et al. 2001). The subduction zone is thought to have regressed and rotated to the southeast during the Pliocene until it reached its current position (Kear 1994, 2004). Intraplate basalt fields of Quaternary age are interspersed throughout northwestern New Zealand (Fig. 1), the positions of which have remained static during subduction zone migration, ruling out any ongoing relationship between the subduction zone and intraplate volcanism (Cole 1986; Smith et al. 1993; Kear 1994, 2004).

The origin of the intraplate volcanism was originally explained by a deep seated mantle plume system beneath northern New Zealand (Heming 1980; Smith et al. 1993). However, more recent geochemical work on the Auckland volcanic field, using U–Th isotopes to determine dynamic melt generation models, indicates that the field has a classic HIMU signature, suggesting a shallow mantle source. In addition, low upwelling rates of melt imply that the source region for the melt lies in the upper mantle at depths between 80 and 140 km rather than at greater depths (Huang et al. 1997). Helium isotope data collected at gas discharges throughout New Zealand have identified regions of recent mantle melting (Hoke & Sutherland 1999); ratios of \(^{3}He/^{4}He\) in the crust are three orders of magnitude less than that from the mantle, with high-\(^{3}He/^{4}He\) regions reflecting where mantle melting is occurring. Measurements of \(^{3}He/^{4}He\) from near the Kaikohe-Bay of Islands and Whangarei volcanic fields (Fig. 1) are the highest (>70 per cent mantle helium) in North Island outside of the Central Volcanic Region. No samples have been taken near the Auckland volcanic field. Thus, both very low upwelling rates and helium isotope data suggest that mantle melting is occurring beneath the Northland peninsula (Huang et al. 1997; Hoke & Sutherland 1999).

Geophysical evidence is in good agreement with the geochemical data for the source location of the intraplate basalts. Mooney (1970) used earthquake traveltime data from stations within the Northland peninsula (Heming 1980; Smith et al. 1993). There is some uncertainty regarding the origin of rhyolite centres within the Kaikohe-Bay of Islands volcanic field (Fig. 1). It is suggested that these rhyolites formed by partial melting of the lower crust as basaltic magma chambers located within the crust (Heming 1980; Smith et al. 1993). There is some uncertainty regarding the origin of rhyolite centres within the Kaikohe-Bay of Islands volcanic field (Fig. 1). It is suggested that these rhyolites formed by partial melting of the lower crust as basaltic magma chambers located within the crust (Heming 1980; Smith et al. 1993). There is some uncertainty regarding the origin of rhyolite centres within the Kaikohe-Bay of Islands volcanic field (Fig. 1). It is suggested that these rhyolites formed by partial melting of the lower crust as basaltic magma chambers located within the crust (Heming 1980; Smith et al. 1993).

During this study, the crust and upper mantle structure model of northwestern New Zealand have been solely based on a reconnaissance seismic refraction survey, which was designed to provide a first-order approximation of the crustal structure for the region (Stern et al. 1987). A crustal thickness of 25 ± 2 km was found. However, due to large receiver spacings and poor data recovery rates, the possibility of hidden layers and low-velocity zones (LVZs) could not be ruled out at the time. P-wave velocities in the lower crust were found to range from 5.3 to 5.9 km s\(^{-1}\), increasing to ~6.2 km s\(^{-1}\) in the lower crust. Low-P\(n\) wave speeds of 7.6 km s\(^{-1}\) were obtained just below the Moho, increasing to 7.9 km s\(^{-1}\) at 40 km depth (Stern et al. 1987).

Here, we investigate the shear wave velocity structure to a depth of 100 km beneath northwestern New Zealand using a joint inversion of receiver functions and surface wave phase velocity data. The 1-D structure is determined using data recorded by a linear array of three broad-band seismometers deployed between 2002 August and 2004 February (MATA, MKAZ, TIKO), referred to as the NORD array (NORTHland Deployment). Additional data recorded on three permanent broad-band stations (TOZ, WCZ, OUZ) during the same time period are also used (Fig. 1). Temporary stations used Guralp-40T sensors and Nanometrics Orion data recorders, while permanent stations are equipped with Guralp-70T sensors. They initially used Orion data recorders that were replaced by Quantera data loggers later in the deployment. We derive and present a model for the shear wave velocity structure of the crust and upper mantle in northwestern New Zealand.

### 2 SURFACE WAVE PHASE VELOCITIES

Surface waves have been extensively used to derive valuable information on the shear wave velocity structure with depth (e.g. Brune & Dorman 1963; Knopoff 1972; Kovach 1978). The dispersive characteristics (i.e. faster velocities for longer period waves) of surface waves provide an indication of the average shear wave velocity as a function of depth.

The linear, northwestern orientation of the NORD array offers a favourable opportunity to conduct a surface wave investigation of northwestern New Zealand. Calculation of interstation phase velocities requires the event to lie on a similar great circle arc to the station array. The NORD array benefits from seismicity along the subduction zones of the western Pacific.

#### 2.1 Data

The standard practice when applying the interstation method to measure surface wave phase velocities is to use earthquakes that lie within ~5° of the interstation great circle arc. Recent workers (e.g. Ritzwoller et al. 2002; Spetzler et al. 2001, 2002) have quantitatively expressed lateral sensitivity kernels for surface waves and have found that events can be included from a wider backazimuth range. The inclusion of more out-of-line events results in an average lateral sampling due to the wide lateral region that surface waves sample. West et al. (2004) determined that by selecting events within 15° of the interstation path a larger number of events are used in the analysis, resulting in an improvement in the rms velocity error. We search the IRIS-DMC database for earthquakes that lie within 15° of the interstation path and have \(M_w > 6\). A total of 11 events filled these criteria and produced dispersive properties (Fig. 2). All but one event is located to the northwest of the NORD array. The events that were selected are clustered together at similar epicentral distances, a direct consequence of the great circle path of the array crossing the seismicity belt of the western Pacific at a high angle (Fig. 2).

#### 2.2 Slant stacking method

We apply a slant stacking method to determine interstation phase velocities for multiple station triplets, following the method of Herrmann & Ammon (2002). This involves slant stacking a series of waveforms in the frequency domain for a range of slownesses.
We obtain a modulus of the stack, $F$, which is a function of frequency and slowness. The value of $F$ provides an indication of how well a given phase stacks for different velocities, with a maximum of $F$ indicating that significant energy is arriving at that velocity (Herrmann & Ammon 2002; West et al. 2004). Benefits of this method include removal of the source spectrum from the analysis, correction of geometric spreading and separation of the fundamental and higher modes.

Prior to the slant stacking method, we remove the instrument response from the waveforms to minimize distortion of the waveform due to varying corner frequencies of the two sensors used. Applying the slant stacking method, we limit the range of velocities to between 2.5 and 5 km s$^{-1}$ and display the results as phase velocity as a function of period (Fig. 3).

We follow the procedure of West et al. (2004), constructing a rolling bin to measure phase velocities; the slant stack is performed on a set of traces from three stations. The bin is then rolled along one station and the stack is performed again. The resulting phase velocities for each station triplet are representative of the centre station, since it is sampling the region between the first and third stations in the triplet. For the NORD array, this results in four station triplets (TIKO–OUZ–WCZ, OUZ–WCZ–MATA, WCZ–MATA–MKAZ, MATA–MKAZ–TOZ). Following calculation of the stack for each event for each station triplet, we perform a second stack of $F$ for multiple events for each station triplet. In the stack, each event is summed with its raw amplitude, so that larger events contribute more energy to the stack. This second stack removes effects of multipathing and spectral gaps that can affect individual stacks.

To determine the final dispersion curve for each station triplet, we search each second stack for the maximum value of $F$ for periods between 10 and 50 s, using a sampling interval of 1 s (Fig. 3, white line).

### 2.3 Observed phase velocities

Final stack functions for each station triplet show good dispersion between periods of 15 and 50 s (Fig. 3), indicated by the dark grey

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**Figure 2.** Hypocentres of events used to determine interstation phase velocities (stars) and in receiver function analysis (circles and stars). Note that most of the events are from the northwest.

**Figure 3.** Final stack functions, $F$, for each station triplet. The white line represents the resulting dispersion curve for each station triplet. We see good dispersion between 15 and 50 s for all of the stations. Poor signal at shorter periods could be a result of multipathing due to the wide station spacing.
colour which is the value of $F$, the stack function. There is little energy at periods lower than 15 s; this could be due to a lack of coherency across the array at short periods, due in part to the large aperture between stations (~100 km).

The lack of surface wave constraints at the near-surface (i.e. less than 15 s period) will not influence deeper parts in the model.

Some initial conclusions can be drawn from the raw phase velocities projected onto a 2-D profile (Fig. 4). Lateral variations are observed across the array, with phase velocities beneath the southern stations 5–10 per cent slower than beneath northern stations for periods shorter than 40 s. Slower velocities in the south are in agreement with regional surface wave studies in the southwest Pacific, which also see the same trend moving from oceanic crust to the continental crust of New Zealand (Pillet et al. 1999). There are two possible explanations for the change in phase velocities across the array, perhaps in combination. If the crust is the same thickness across the array, then upper mantle velocities must be lower in the south. However, if the crust is thinner in the north, faster velocities would be reached at shallower depths than in the south, where the crust is thicker. Phase velocities derived for the southern end of the NORD array (Fig. 3) are $\sim 0.2 \text{ km s}^{-1}$ lower at periods greater than 20 s than phase velocities obtained by Briscour & Stuart (1998) for interstation paths crossing eastern North Island, New Zealand, which may be affected by the high-velocity, subducting slab.

3 RECEIVER FUNCTIONS

3.1 Data

We searched the IRIS-DMC catalogue for earthquakes that lie within 25°–100° epicentral distance of the NORD array and have $M_w > 6$, and found 118 events which filled these criteria that were further used in the receiver function analysis (Fig. 2). The epicentral range was chosen to include events that arrive near vertically beneath the station but to avoid core phases for events at larger epicentral distances. We use the term backazimuth to describe the angle from north between the station and event. The majority of events had backazimuths ranging from 270°–020°, originating in subduction zones in the western Pacific (Fig. 2). Few events lie outside this backazimuth range; events that do are usually at the high end of the epicentral distance range and have low magnitudes, which results in poor-quality receiver functions. Events from the northwest quadrant, however, have good epicentral coverage. Not all events are recorded on every station due to noise levels on individual stations and technical difficulties.

3.2 Multiple-taper correlation method

The $P$-wave receiver function method has become a popular tool to investigate the shear wave structure beneath broadband stations using converted phases from interfaces (Langston 1979; Owen & Zandt 1985; Ammon et al. 1990). The method assumes that a seismic waveform is a product of three factors: source side, travel path and receiver side effects. It is assumed that source and travel path effects are removed from the waveform by rotating the components into the $Z$–$R$–$T$ coordinate system followed by deconvolution of the vertical component from the horizontal components. This leaves only the receiver response, characterized as a time-series that contains phases that result from $P$-to-$S$ conversions ($Ps$) at seismic discontinuities beneath the station. In addition to primary $Ps$ phases, we also record secondary reflections (multiples) that are generated by reverberations from the free surface or another interface. Multiples arrive later in the time-series and can mask deeper $Ps$ phases. In terms of primary $Ps$ phases, the time-delay between the initial $P$ wave and a given $Ps$ phase is a function of the depth of the interface and the difference of $P$- and $S$-wave slownesses. The amplitude of the phase is dependent on the velocity contrast across the interface and the incident angle of the incoming wave. Positive phases are generated by a velocity increase with depth (e.g. Moho) and a negative phase is caused by a velocity decrease with depth (i.e. the top of a low-velocity layer). Note that some multiples exhibit negative arrivals in spite of increasing velocity. We use the radial receiver function to determine the shear velocity structure beneath the station. Significant coherent energy on the transverse receiver functions is caused by out-of-plane energy, indicative of seismic anisotropy or dipping interfaces (Cassidy 1992; Savage 1998).
We compute teleseismic $P$-wave receiver functions using the multiple-taper correlation (MTC) method of Park & Levin (2000) as an alternative to popular methods such as time domain deconvolution (Gurrola et al. 1995; Abers 1998) and spectral division (Ammon 1991). The MTC method computes receiver functions in the frequency domain, avoiding numerical instabilities often associated with spectral division. The MTC method distinguishes between coherent and incoherent scattering to reduce unwanted seismic energy. Once the receiver functions are computed, they are binned as a function of backazimuth. Each bin is separated by 5° with a 10° overlap between bins so that each trace influences each adjacent bin. An advantage of using the MTC method is that each receiver function is weighted by the inverse of its variance during bin formation or stacking of multiple events. This positively biases towards the best receiver functions and results in an improvement in multiple event stacks, both for receiver function gathers and for formulating summed traces that are used to represent stations (Park & Levin 2000). We compute receiver functions for all events recorded on each station, applying a low-pass filter of 1 Hz during analysis.

4 JOINT INVERSION FOR SHEAR WAVE STRUCTURE

We determine the shear wave velocity as a function of depth using a linearized joint inversion of radial receiver functions and Rayleigh wave phase velocities for northwestern New Zealand, following the method of Julia et al. (2000, 2003). A solution is obtained by a damped least-squares method that smoothes velocities in adjacent layers based on a priori constraints. An influence factor, $p$ (0 ≤ $p$ ≤ 1) is used to specify the weighting of receiver function and surface wave data, where $p = 0$ is receiver function data only and $p = 1$ is surface wave dispersion data only. We tested the influence of $p$ and determined that a value of $p = 0.5$ was robust and provided the best fit to both data sets.

We base our starting velocity model on the velocity model obtained by Stern et al. (1987) for northwestern New Zealand, which consisted of a two-layer crust and a two-layer mantle. We invert for the shear wave structure down to 100 km, as the vertical sensitivity kernels of the surface waves sample to a depth of ~200 km and our receiver functions are sensitive to 100 km depth. Layer thicknesses are set to 1 km in the upper 6 km and 2 km for the rest of the model. We solve for $V_s$ and determine $V_p$ by keeping $V_p/V_s$ fixed at 1.75 for the crust and at 1.8 for the mantle. Parts of the region have low $Pn$ wave speeds, indicating partial melt in the uppermost mantle (Stern et al. 1987; Stratford & Stern 2005). This will subsequently increase the true value of $V_p/V_s$ and result in underestimation of the $V_p$ values.

Input for the joint inversion uses a stacked receiver function obtained by summing all events from the northwest quadrant. We lack a range of backazimuths to complete a thorough investigation of the presence of anisotropy and dipping layers, thus we take a first-order approach by determining the average velocity structure beneath each station by stacking good quality events with backazimuths in the northwest quadrant. This approach will tend to smooth the interpreted structure, since the receiver functions will be smoothed by arrivals at slightly different times across the backazimuth range. The corresponding surface wave data consist of dispersion curves from each station triplet, with a sampling interval of 1 s between points on the dispersion curve. The two edge stations are represented by using the dispersion curves of the adjacent stations. We assume that the structure is 1-D and isotropic, which is a reasonable assumption, given that there is minimal shear wave splitting observed across the array from SKS phases (Duclos 2005) and that there is only minor energy observed on the transverse receiver functions (Figs 5–10). The strength in the joint inversion lies in that fact that surface wave data at the near-surface will not influence the rest of the model which typically happens during forward modelling where the user takes a down-model approach to fitting the waveform.

It is difficult to estimate errors during the inversion procedure using a linear damped least-squares method (see Julia et al. 2000). Therefore, we estimate errors by examining errors in surface wave phase velocities and timing of receiver function phases. We first invert for the shear structure by using three dispersion curves: the observed dispersion and two obtained by adding and subtracting the phase velocity error of 0.09 km s$^{-1}$. The shear velocities from the resulting three models vary on average by 0.12 km s$^{-1}$. An error representing the misfit in timing of receiver function phases is estimated to be ±0.05 s. We investigate the error in the depth of an interface by using the relationship between the timing of a $Ps$ phase and the depth to the interface (Gurrola et al. 1994):

$$H = \frac{TPs}{(V_s s^{-2} - P^2)^{0.5} - (V_p s^{-2} - P^2)^{0.5}},$$

where $H$ is the depth to the interface, $TPs$ is the delay time of the $Ps$, $V_s$ and $V_p$ are the shear and compressional velocities of the layer. The ray parameter $p$, takes into account variations from vertical ray paths. We calculate the depth to the interface, using extreme values of $V_s$ (±0.12 km s$^{-1}$) and $TPs$ (±0.05 s) based on errors mentioned previously. From these crude error calculations, shear velocities in a given layer vary by ±0.12 km s$^{-1}$ and the depth to an interface varies by ±1 km.

4.1 Station TOZ

Station TOZ is located just west of the Central Volcanic Region, the area of active extension in the central North Island. TOZ has the largest data set of all of the stations, though the majority of events still come from the northwest quadrant (Fig. 5). A strong $Ps$ phase arrives at 3.5 s delay time, which is interpreted to be from the Moho. The timing indicates a shallow Moho, around 25–30 km depth. There is also some coherent energy on the transverse component for backazimuths 300°–340° at 2–3 s delay time.

Fig. 5 shows the inversion results for station TOZ. We plot the observed receiver functions and surface wave dispersion data as grey lines. The predicted data are shown as a black line. In general, we see a good fit between the two data sets; we note that the predicted dispersion is smoothed compared to our observed data. Receiver functions fit well but diverge slightly at delay times between 14 and 16 s. Low shear velocities of 3.2 km s$^{-1}$ are observed to 4 km depth, but sharply increase in the mid-crust to faster velocities of 3.5–3.7 km s$^{-1}$, which extend to 15 km depth. Lower velocities occur from here to the Moho. A sharp Moho is seen at ~29 km, where shear wave velocities of 4.2 km s$^{-1}$ are reached. This matches the $Ps$ phase at 3.5 s delay time well. Low upper mantle velocities extend down to 100 km depth. The velocity model is remarkably similar to shear velocity models found ~50 km to the southeast, for stations CAMV and MAMV, just inside the western boundary of the Central Volcanic Region (Bannister et al. 2004).

4.2 Station MKAZ

MKAZ is located in the Hunua Ranges, a region where greywacke basement outcrops. The Auckland Volcanic Field lies 30 km to the
The receiver function gather (Fig. 6) shows a coherent $Ps$ phase arriving at $\sim 3$ s delay time, interpreted to be from a shallow Moho. Some variability is noted for phases arriving later in the receiver functions but a positive–negative phase with a large amplitude arriving at 9 s delay time is seen on multiple events. When plotted against epicentral distance (not shown here), this phase exhibits normal moveout, indicating that it is a primary conversion and not a multiple. There are some coherent arrivals on the transverse component at 0–2 and 6–8 s delay time.
Inversion results show low velocities of 2.5–3.0 km s\(^{-1}\) in the top 1–2 km (Fig. 6). Below this, a steady positive gradient in shear velocities extends to 20 km depth. Upper mantle velocities are reached at 27–29 km depth. On the receiver function, the negative–positive arrival at 9 s delay time (Fig. 6) is modelled by an LVZ in the upper mantle at 70–90 km depth with shear velocities as low as 4.0 km s\(^{-1}\). These velocities are 15 per cent lower than typical continental upper mantle velocities (Christensen & Mooney 1995). The positive phase is generated from the bottom of the LVZ, while the negative phase at 9 s derives from the top of the LVZ.

4.3 Station MATA

MATA is located in a sedimentary basin of Miocene age consisting of alternating sandstones and mudstones. There is some variation in the radial receiver functions with backazimuth for station MATA (Fig. 7), with \(P_s\) phases arriving at 3–4 s delay time and some earlier phases arriving at 0–2 s time over a varying range of backazimuths. The transverse component is quite noisy with a semicoherent arrival at 5–6 s time; this varies in amplitude, period and timing as a function of backazimuth. The presence of anisotropy or dipping structures could be causing the heterogeneity in the receiver functions.

Inversion results indicate a model with velocities decreasing from 3.6 to 3.2 km s\(^{-1}\) in the near-surface to 8 km depth (Fig. 7). Beneath this a steady increase in velocities from 3.6 to 3.9 km s\(^{-1}\) is observed, with a jump from crust to upper mantle velocities at 26–29 km depth. Another step in upper mantle velocities is seen from 4.3 to 4.5 km s\(^{-1}\) at 45 km depth, with velocities subsequently gradually increasing to 4.7 km s\(^{-1}\) at the base of the model. The observed and predicted data fit well and again we see smoothing of the predicted dispersion curve.

4.4 Stations WCZ and OUZ

Figs 8 and 9 show the receiver function gathers for stations WCZ and OUZ. These two permanent stations are located near the Whangarei
and Kaikohe-Bay of Islands Volcanic Fields, which consist of intraplate basaltic vents and rhyolite centres of Quaternary age. The radial receiver functions for the stations are remarkably similar. They both have a large $P_s$ phase at $\sim 3.5$ s delay time, interpreted to be from the Moho. This phase is preceded by positive phase at 2–3 s delay time and a large amplitude negative phase at 1–2 s delay time. The initial $P$ wave exhibits no delay, indicating that near-surface layers are not causing this strong negative arrival. It is thought that the two arrivals before the Moho $P_s$ originate from a mid-crustal discontinuity; relative arrival times are consistent with conversions from the top and bottom of an LVZ. For station WCZ, there is also some energy arriving at 2 s delay time on the transverse component.

The inversion results for stations WCZ and OUZ (Figs 8 and 9) are quite similar. WCZ is located near the Whangarei Volcanic Field. The best-fitting velocity model for WCZ shows low velocities of 3.0 km s$^{-1}$ in the upper 1–2 km, which increase to 3.5 km s$^{-1}$ at $\sim 11$ km depth. Beneath this is an LVZ between 12 and 18 km depth, with shear velocities as low as 3.0 km s$^{-1}$. The LVZ is required to be present to fit the large-amplitude negative arrival in the receiver functions at 1–2 s delay time. The predicted receiver function underestimates the magnitude of the negative arrival at 1–2 s time. As a consequence, the velocities in the LVZ are probably overestimated and are perhaps lower than 3.0 km s$^{-1}$. Beneath the LVZ, velocities increase and at 26–28 km depth jump to upper mantle velocities. A gentle gradient is observed in the upper mantle, with velocities reaching 4.5–4.6 km s$^{-1}$ at 100 km depth.

The inversion results for OUZ show velocities of $\sim 3.0$ km s$^{-1}$ for the near-surface increasing to 3.5–3.8 km s$^{-1}$ down to 9 km depth (Fig. 9). A large negative arrival on the receiver function at 1–2 s time requires the model to have an LVZ between 10 and 19 km depth with velocities of 2.9–3.0 km s$^{-1}$, a 20–25 per cent reduction from normal mid-crustal velocities. As with the WCZ model, the best-fitting model for OUZ overestimates the velocities in the LVZ, not quite fitting the full amplitude of the negative phase. Beneath the LVZ, a steady increase in shear velocities is observed down to 24–25 km depth. A jump to upper mantle velocities occurs at 26 km depth. Upper mantle shear velocities then increase to a depth of...
Figure 8. Receiver function gathers and joint inversion analysis for station WCZ. Details as in Fig. 5. An LVZ in the mid-crust is needed to fit the negative arrival at 1–2 s time on the receiver function.

55 km where a slight decrease of 0.3 km s\(^{-1}\) occurs at 55–70 km depth. We see good fit between both data sets for both stations.

4.5 Station TIKO

Station TIKO is located at the northern tip of New Zealand. The station is situated on a thick sequence (~1000 m) of obducted oceanic sediments that form the Northland Allochthon (Herzer & Isaac 1992) and also more recent sediments that make up the Northland Basin (Hayward 1993). We observe a strong coherent \(P_s\) arrival at 3 s delay time (Fig. 10). An earlier negative arrival at 2 s time is also seen. However, unlike stations OUZ and WCZ, the initial \(P\) wave exhibits interference with a near-surface conversion that causes it to broaden and be slightly delayed. This indicates that the negative conversion is most likely from a multiple from an interface between basement and overlying sedimentary rocks.

The initial \(P\) delay and the negative phase arriving at 2 s delay time require the model to have low velocities of 2.6–2.9 km s\(^{-1}\) in the upper 7 km (Fig. 10). Below this, velocities strongly increase to 10 km depth, to mid-crustal wave speeds of 3.5 km s\(^{-1}\). A large step to upper mantle velocities occurs at 25–26 km depth, below which velocities increase to 4.5–4.7 km s\(^{-1}\). A good fit is seen between both data sets.

5 DISCUSSION

5.1 Crust and upper mantle structure

We create a 2-D shear wave velocity model by interpolating between individual 1-D profiles for each station (Fig. 11). A surface is created using splines in tension with a tension factor of 1. We follow the method of Bannister et al. (2004) by investigating the derivative of the shear wave velocity with depth. Receiver functions are primarily sensitive to velocity contrasts, so by observing the areas where the shear velocity changes the most rapidly we can identify where the receiver functions are most sensitive (Fig. 11). In general, our crustal
structure model varies with similar trends to the P-wave model of Stern et al. (1987). Although their model involved a simple two-layer crust, we observe similar lateral variations in the upper crust. In the central region of our array (between OUZ and MATA), shear velocities in the upper crust are 3.4–3.6 km s\(^{-1}\), while at the northern and southern end velocities are lower at <3.2 km s\(^{-1}\) (Fig. 11). Stern et al. (1987) modelled P-wave velocities of 5.9 km s\(^{-1}\) for the central portion of our array and 5.3 km s\(^{-1}\) and 5.4 km s\(^{-1}\) for the north and south end, respectively. The lower crust is remarkably consistent across the profile with average shear wave velocities of 3.6–4.0 km s\(^{-1}\). This equates to a \(\frac{V_p}{V_s}\) of 1.72–1.76 for the upper crust and 1.70–1.75 for the lower crust, using the P-wave velocities of Stern et al. (1987). Low shear wave velocities of 4.2 km s\(^{-1}\) are present in the uppermost mantle across the profile but extend to greater depths (∼90 km) beneath TOZ. Upper mantle velocities of 4.2 km s\(^{-1}\) are ∼10 per cent less than the global average (Christensen & Mooney 1995) and indicate a disturbed upper mantle. Velocities are also lower than the 4.45 km s\(^{-1}\) obtained for the northwestern North Island by Haines (1979) for Sn phases using earthquake travel-times. It must be noted that the value Haines (1979) found was based on very few measurements and that the ray paths most likely sample slightly deeper mantle. A \(\frac{V_p}{V_s}\) for the uppermost mantle is 1.81 ± 0.03 using the \(Pn\) velocities of Stern et al. (1987).

Localized features are present along the array. Beneath TIKO, low-velocity layers of 2.5–2.9 km s\(^{-1}\) in the upper 6–8 km most likely reflect a thick sequence of sediments, comprised of 1–2 km of material from the Northland Allochthon (Herzer & Isaac 1992) and the remainder from sediments in the Northland Basin. Tertiary sediments associated with the Northland Basin, located 30 km southwest of Cape Reinga, have been inferred to be up to 5 km thick from gravity modelling (Woodward 1993). An LVZ is present beneath OUZ and WCZ in the mid-crust at depths of 10–19 km, with shear velocities as low as 2.8 km s\(^{-1}\). The top and bottom of the LVZ are apparent in the velocity gradient plot (Fig. 11). The seismic refraction line of Stern et al. (1987) did not identify any LVZ in Northland; however, their receiver spacing was very large (20–50 km) and it is doubtful that they could identify such layers if present. In the upper mantle beneath MKAZ, another LVZ is present at 70–90 km depth and velocities drop to ∼4.0 km s\(^{-1}\), ∼15 per cent lower than normal upper mantle velocities (Christensen & Mooney 1995).
5.2 Crustal thickness

There is often heated debate as to the depth of the Moho in tectonically active regions of the Earth (Christensen & Mooney 1995). Investigations using the same data set have even resulted in different depths to the Moho in the Central Volcanic Region, New Zealand (Harrison & White 2004; Stratford & Stern 2005) and the Sierra Nevada, California (Jones et al. 1994; Savage et al. 1994). The underlying question fuelling the debate is ‘what velocities signify that the upper mantle has been reached?’ This seemingly simple question is complicated by the fact that the upper mantle in regions that have undergone extension or are adjacent to plate boundaries is usually hot with small amounts of partial melt (Stern 2002; Wiens & Smith 2003). The presence of high temperatures and partial melt decrease seismic velocities, with shear velocities being affected to a larger degree than compressional velocities (Sato & Sacks 1989; Hammond & Humphreys 2000). P and S velocities in the upper mantle beneath northwestern New Zealand have been shown to be lower than that would be expected for typical continental settings (Stern et al. 1987; Bannister et al. 2004; Harrison & White 2004; Stratford & Stern 2005).

Instead of assuming an upper mantle velocity and inferring that the Moho lies where this velocity is reached, we determine the Moho by examining the shear velocity gradient (Fig. 11), and interpret the Moho to lie just below where the shear velocity changes the most. This occurs at $29 \pm 1$ km beneath TOZ at the southern end of the array and at $26 \pm 1$ km beneath TIKO at the northern end (Fig. 11). The Moho depth ranges between these two values along the profile, although the vertical gradient is not as strong beneath MKAZ. Projecting the Moho from the gradient plot onto the velocity profile, we obtain a velocity of $4.2 \pm 0.1$ km s$^{-1}$ for the upper mantle (Fig. 11, dotted line). The only other crustal scale investigations that correspond directly to our array obtained a depth of $25 \pm 2$ km for the Moho (Stern et al. 1987) for a seismic refraction line that ran along the Northland Peninsula. However, their crustal thickness estimate is a minimum because the presence of hidden high-velocity layers may...
Figure 11. Top panel: shear wave velocity profile across the array formed by interpolation between each 1-D profile using splines in tension (tension factor = 1). Note that horizontal sampling is sparse. Locations of intraplate basalt fields (black bars) are projected onto the profile at top: Kaikohe-Bay of Islands (KBIVF), Whangarei (WVF) and Auckland (AVF). The triangles represent the locations of the broad-band stations. Key features include an LVZ in the mid-crust beneath OUZ and WCZ and a broader LVZ in the upper mantle beneath MKAZ. Low upper mantle velocities extend to greater depths beneath TOZ. There is a sequence of low-velocity materials beneath TIKO. The blacked dotted line indicates the position of the Moho based on the velocity gradient plot. Bottom panel: shear wave velocity gradient with depth, down to 50 km depth. Blue colours represent an increase in shear velocity with depth, as would be expected at the Moho. We can identify a Moho where the gradient is the highest along the array. This ranges from 29 km beneath TOZ to 26 km beneath TIKO. The LVZs beneath OUZ and WCZ are also apparent by the presence of a negative velocity gradient at the top of the LVZ.

increase this value. Other studies located 30–50 km north of TIKO on the Reinga Ridge, a continuation of Cape Reinga, have determined a crustal thickness of 25 km based on modelling of marine gravity data (Zhu & Symonds 1994; Herzer et al. 1997). A number of studies have been conducted across the Central Volcanic Region, which has its western margin only 30–40 km to the southeast of TOZ. There is debate over the crustal thickness in the Taupo Volcanic Zone, the active part of the Central Volcanic Region, with estimates varying from 15–21 km (Stratford & Stern 2004, 2005) to 25–30 km (Bannister et al. 2004; Harrison & White 2004). In the western part of the Central Volcanic Region, values for the Moho range from 25 km from wide-angle seismic data and earthquake traveltimes (Stern et al. 1987; Harrison & White 2004; Stratford & Stern 2005) to 26–30 km from receiver function inversion (Bannister et al. 2004). All of these estimates are similar to the 26–29 km obtained from our joint inversion model.

5.3 Low-velocity zones

Joint inversion results reveal LVZs in the mid-crust and upper mantle beneath northern New Zealand.

The mid-crustal LVZ is located beneath WCZ and OUZ, two stations that lie on or adjacent to intraplate basalt fields of Quaternary age. A source region for the basalts is thought to lie in the mantle (Heming 1980; Smith et al. 1993). However, the presence of rhyolite centres and an active geothermal system in the Kaikohe-Bay of Islands Volcanic Field indicate that melting of the crust is occurring and that there is the possibility of a magma body heating
the geothermal field (Heming 1980). The LVZ is located at 10–19 km depth, with shear velocities as low as 2.9 km s\(^{-1}\) (Fig. 11), a ~23 per cent reduction on normal crustal velocities. Large reductions in shear velocities are usually caused by the presence of fluids or melt (Sato & Sacks 1989; Hammond & Humphreys 2000). LVZs have been identified in the mid-crust by teleseismic receiver functions and been interpreted as melt bodies in Socorro (Sheetz & Shlue 1992), the Andes (Chmielowski et al. 1999) and in the Taupo Volcanic Zone (Bannister et al. 2004). Therefore, our LVZ is interpreted to represent a body of partial melt in the crust. The partial melt could be generated by a batch of basaltic magmas that have migrated from the mantle into the crust, melting the lower crust and forming alkalic rhyolite magma, thus explaining the presence of rhyolite centres in the Kaikohe-Bay of Islands Volcanic Field. A rhyolite intrusion could also be the heat source for the Ngawha geothermal system near OUZ, as it is thought that the basalt magmas do not reside in the crust for long enough to create a heat source. The interpretation of a zone of partial melt in the mid-crust is consistent with patterns of volcanism at the surface. Regional Bouguer gravity data do not have any significant anomalies that would indicate partial melt in the crust; however, the resolution of the data set is too low to identify such bodies.

Beneath station MKAZ, an LVZ is present in the final model at a depth of 70–90 km in the upper mantle (Figs 6 and 11), with velocities of 4.0 km s\(^{-1}\). This LVZ is inferred from a large negative P\(s\) arrival at 9 s delay time on the receiver functions (Figs 6 and 12). This time-delay is near where we would expect multiples from the Moho to arrive. Therefore, we test whether this arrival is a P\(s\) arrival or instead just a multiple by using reflectivity forward modelling of synthetic receiver functions (Fig. 12). We begin by removing the LVZ from the final velocity model for MKAZ. By doing this, we will observe when the Moho multiples should arrive (Fig. 12a). The P\(s\) phase from the Moho arrives at ~3 s time but the multiple (P\(s\)P\(p\)ms) from the Moho arrives at 11–12 s delay time, indicating that the 9-s phase is not a multiple of the Moho in the inversion model. Secondly, we determine the thickness of the crust required to generate a multiple arriving at the same time as our 9-s P\(s\) phase. We use a simple two-layer model with a shear velocity in the crust of 3.6 km s\(^{-1}\), and 4.3 km s\(^{-1}\) in the mantle (Fig. 12b). We then modify the crustal thickness until we fit the Moho P\(s\) phase at 3 s time and the 9-s phase. We need a 21-km-thick crust in order to obtain a Moho multiple arriving at 9 s time for the simple velocity model used. Based on these models, we can determine that this LVZ is real, unless the crust has a thickness of 21 km which is much thinner than that previously observed. It would also mean that the crust would thin from 29 km beneath TOZ and 28 km beneath MATA to 21 km beneath MKAZ, an unlikely scenario.

As a consequence of the incoming waves arriving at an angle beneath the station, the receiver functions actually sample an area surrounding the station. We backproject all of the receiver functions recorded from the northwest of MKAZ to observe their piercing points at 90 km depth, the base of the LVZ (Fig. 13). We plot the piercing points for receiver functions as crosses if they have the 9-s phase, or as circles if they do not (Fig. 6). The majority of the receiver functions with the 9-s P\(s\) phase pierce in or around the boundary of the Auckland Volcanic Field (Sporli & Eastwood 1997).
indicates that the LVZ is located not beneath MKAZ, but beneath the Auckland Volcanic Field. Recent work suggests that temperature variations and melt content can cause velocity reductions in the upper mantle; thus, the LVZ is thought to represent a melt-producing region in the upper mantle (Hammond & Humphreys 2000).

In light of this work, we propose that our LVZ at 70–90 km depth represents the source region for the Auckland Volcanic Field. The location of the melt-producing zone has implications for the location of future volcanism. Receiver functions that contain the $P_s$ phase from the LVZ mostly pierce directly beneath the Auckland Volcanic Field (Fig. 13). However, a small number of receiver functions pierce the base of the LVZ ∼10 km north of the present extent of volcanism (Sporli & Eastwood 1997). If we assume that volcanism occurs directly above the source region, then this implies that future volcanism could occur further north than the present boundary of the field.

5.4 Limits on the lateral extent of mantle wedge deformation below the Central Volcanic Region

We observe a vertical boundary between station TOZ and MKAZ that marks a change in upper mantle velocities (Fig. 11). This is consistent with several lines of evidence that indicate that a boundary marking the extent of mantle deformation in the mantle wedge lies around or near the western boundary of the Central Volcanic Region. Pulford & Stern (2004) calculated buoyancy forces in the mantle beneath North Island, based on rock-uplift histories. A buoyancy force per unit area of 65 MPa is found beneath the Central Volcanic Region and 35 MPa beneath northwestern North Island. Stratford & Stern (2005) expressed the buoyancy force as density and showed that there is a density anomaly of −70 km m$^{-3}$ to a depth of 80–100 km below the Central Volcanic Region and a −40 kg m$^{-3}$ anomaly below northwestern North Island. They attributed half of the anomaly to temperature differences and the remainder to partial melt. Using $P_s$ velocities for northwestern North Island (Stern et al. 1987) and the Central Volcanic Region (Stratford & Stern 2005), they estimated 1 and 2 per cent melt, respectively, based on estimating melt percentage from $P$-wave speed reductions (Hammond & Humphreys 2000). Similarly relating our upper mantle shear wave velocities to partial melt, we obtain 1 per cent beneath the array and 1–2 per cent beneath TOZ. However, the relation between percent melt and geophysical properties is an area of active investigation, and these values will change if new values are determined.

A swath of seismometers deployed across much of the central North Island have been used to image the bulk structure of the subduction zone beneath North Island by body wave tomography (Reyners et al. 2005). The mantle wedge is identified as a region with low $V_p$ (7.5–7.9 km s$^{-1}$) and high $V_p/V_s$ (1.87) that extends to a depth of ∼100 km. It is unclear how far north this zone extends as the model has lower resolution north of the Central Volcanic Region (Reyners et al. 2005). Converting the $V_p$ found in the mantle back-arc wedge to $V_s$, we obtain a value of 4.0–4.25 km s$^{-1}$ for $V_s$ in the mantle wedge. This is similar to the upper mantle shear velocities we obtain beneath TOZ down to a depth of 90 km, indicating that the processes causing the low velocities in the mantle back-arc wedge are still occurring beneath TOZ. Since there is a change in upper mantle velocities between TOZ and MKAZ, we propose that the change in velocities at 40–90 km depth marks a boundary to the extent of mantle deformation in the mantle wedge, approximately 100 km northwest of the active volcanic front. Lateral velocity boundaries have been observed from body wave traveltime tomography in other subduction zone settings, such as Japan (Zhao et al. 1992) and Tonga (Zhao et al. 1997). It is difficult to compare the extent of velocity anomalies in the mantle wedge between different subduction systems. Variations in dip of the subducting slab and age of the subduction zone contribute to the size of the mantle back-arc...
wedge, making it difficult to suggest a standard model for the extent of horizontal deformation.

Seismic anisotropy has been successfully used to infer in situ deformation in the mantle and crust by means of shear wave splitting (Savage 1999). Fast orientations from SKS splitting are used to infer the direction of mantle flow. Shear wave splitting from SKS phases has been carried out on a number of stations in North Island (Marson-Pidgeon et al. 1999; Hoffman 2002; Audoine et al. 2004; Duclos 2005). In North Island, nearly all stations exhibit trench-parallel fast orientations (Marson-Pidgeon et al. 1999; Hoffman 2002; Audoine et al. 2004), indicating trench-parallel flow above and below the subducting slab (Savage 1999). Recent SKS splitting results on the same stations used in this study show somewhat different results (Duclos 2005). TOZ is similar to other stations in North Island, with a NE–SW fast orientation, suggesting similar mantle deformation as in the mantle back-arc wedge. However, the other stations exhibit E–W fast orientations, indicating that the trench-parallel flow is not occurring beneath them. No anisotropy is observed beneath MKAZ. This change in fast orientation implies that a boundary to the trench-parallel mantle flow occurs near TOZ (Duclos 2005).

The change in the SKS fast orientation (Duclos 2005), the presence of low shear velocities that extend to greater depth beneath TOZ than the rest of northwestern North Island (Fig. 11) and a density difference between the Central Volcanic Region and northwestern North Island (Pulford & Stern 2004; Stratford & Stern 2005) all indicate that a boundary exists near station TOZ, which marks the extent of deformation associated with the mantle back-arc wedge to the south.

6 CONCLUSIONS

We successfully constrain the shear wave velocity structure to 100 km depth beneath northwestern New Zealand by performing a joint inversion of receiver functions and surface waves. A joint inversion removes the limitations of each method and results in a more robust velocity model.

Crustal thickness ranges from 26 ± 1 to 29 ± 1 km, thinning towards the northern tip of New Zealand, near the boundary between oceanic and continental crust. The inferred thicknesses are in agreement with previous seismic and marine gravity investigations.

Two stations located on or adjacent to Quaternary intraplate basaltic fields exhibit large-amplitude negative arrivals on the receiver functions at 1–2 s delay time. Inversion results attribute these phases to an LVZ at 10–19 km depth, with a 23 per cent reduction in shear velocities. We infer that the LVZ represents partial melt in the mid-crust, consistent with surface volcanic and geothermal features.

An LVZ is also present in the upper mantle at 70–90 km depth beneath the Auckland Volcanic Field. Shear velocity reductions indicate the LVZ may represent a melt-producing region in the upper mantle. Such a zone is interpreted to represent the source region for the intraplate basalts of the Auckland Field. This is the first time such a source region has been directly imaged beneath the field but is in agreement with theoretical source zones based on geochemical and other geophysical data, and is in agreement with models for a shallow upper mantle source rather than a deep seated plume.

Lateral changes in upper mantle shear velocities are observed beneath the array. A boundary that marks a change from low to normal upper mantle velocities ~100 km northwest of the active volcanic front may represent the northwestern extent of mantle deformation which occurs in the mantle wedge beneath the Central Volcanic Region. This would imply that the deformation extends 50 km northwest of the western Boundary of the Central Volcanic Region. A change in the mantle properties further to the northwest is also supported by changes in seismic anisotropy fast orientations.

This first-order determination of the shear wave velocity structure beneath northwestern New Zealand could be tested by deploying a more dense network of receivers around the stations. This would also allow the magnitude of anisotropy and dipping layers that are affecting our models to be determined.

ACKNOWLEDGMENTS

We are grateful to Chuck Ammon and Robert Herrmann for the use of their computer programs in seismology suite and Jeff Park and Vadim Levin for their MTC program. Figures were generated using Generic Mapping Tools (GMT) of Wessel & Smith (1991). Mathieu Duclos and Ken Gledhill helped with field work and initial data processing. GeoNet is acknowledged for providing data from their permanent broad-band seismometer stations. Helpful discussions with Tim Stern, Mathieu Duclos, Rick Herzer and Audrey Galve improved this study. Funding for the instruments was provided by the Planet Earth Fund, New Zealand Lotteries Board and the Royal Society’s Marsden Grant. Additional financial support was provided by the Foundation for Research, Science and Technology (FRST).

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