Slow-slip events and small earthquake clustering - implications for the locked region of the shallow Hikurangi subduction zone

R. J. Beavan
R. McCaffrey
M. E. Reyners
L. M. Wallace

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R. J. Beavan, GNS Science, PO Box 30 368, Lower Hutt
R. McCaffrey, GNS Science, PO Box 30 368, Lower Hutt
M. E. Reyners, GNS Science, PO Box 30 368, Lower Hutt
L. M. Wallace, GNS Science, PO Box 30 368, Lower Hutt

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TECHNICAL ABSTRACT

Slow-slip events have been observed along the Hikurangi subduction margin since the continuous GPS network began to be established along the North Island east coast in 2002. These events appear at the surface as a change in direction and rate of motion at survey points monitored by continuously-operating GPS instruments. The change may last from days to months, after which the previous motion resumes. The signals are inferred to result from slip on a patch of the subduction interface fault. The slip rate between the two sides of the fault is on the order of 50 mm/day or less, which is orders of magnitude slower than the slip rate during a normal earthquake. Hence the name "slow-slip event" or "slow earthquake".

The slow-slip events release some of the strain energy that is being accumulated in the region of the plate interface as a result of the inexorable motion of the Pacific Plate beneath the North Island. The amount of energy that is released slowly is therefore unavailable for release in a future earthquake on this part of the subduction interface. If slow-slip events (SSE) release strain energy on parts of the plate interface that we have previously expected to be involved in great earthquakes, then either the moment or the frequency of such earthquakes would be less than we previously believed.

Slow-slip events appear to be located near the down-dip end of the well-coupled part of the subduction interface, in the transition between stick-slip behaviour above and continuous deformation below. By mapping the locations of the slow-slip events we can potentially map the well-coupled region of the interface. This provides information about the extent of the well-coupled region, at least as it appears at the present time. With proper interpretation, this can assist in estimating the maximum size of a future earthquake.

The occurrence of a slow-slip event causes stress changes at shallower depths on the subduction interface. From the data collected to date by ourselves and others, it appears that these stress changes may influence the occurrence of small earthquakes both close to the slip event and perhaps on a more regional basis. These effects are not yet well understood.

A slow-slip event causes an increase of the stress towards failure of the well-coupled part of the subduction interface just above. It is therefore possible that a slow-slip event could trigger a large earthquake at shallower depth on the plate interface. This phenomenon has not yet been confirmed, but its possibility means that the monitoring of slow-slip events as they occur may be of importance. For example, deep, slow-slip has been suggested by others to have occurred just prior to the 1960 Chilean (Mw 9.5), and 1944 and 1946 Nankai subduction earthquakes (each Mw 8.2). Stress triggering has been well demonstrated on crustal faults in various locations around the world, with the North Anatolian fault being a classic example.

For all these reasons we have undertaken a multifaceted investigation of slow-slip events in New Zealand over the past two years.

We have modelled the large 2004-05 Manawatu slow-slip event, and have detected and modelled three other events that occurred within the project period to define their location, depth, duration and magnitude. Two additional events started near the end of the project period.

The SSEs along the east coast occur at very shallow depths (10-15 km) and tend to be quite short (days to weeks). Those under Manawatu and along the Kapiti Coast are much deeper (30-50 km) and longer (many months). We have attempted to determine what causes the variation in depth. It has been assumed elsewhere that temperature is a primary control on the depth at which slow-slip occurs, but we show that this is not the case for the Hikurangi margin.

We have worked towards mapping the extent of the strongly-coupled zone of the subduction interface. We have used both the locations of SSEs and inferences from microearthquake locations together with results of seismic tomography, and we find that these give reasonably consistent results. A speculative model is proposed in which the permeability of the rocks in the overriding plate may affect the degree of coupling and hence the location of SSEs.

We have developed an automated method of detecting the onset of SSEs soon after they begin. The algorithm is running every day in test mode. It is based on the cross-correlation between the observed GPS time series and a large set of candidate SSEs.

It was suspected prior to this project that the 2003-04 Kapiti Coast SSE had triggered the occurrence of moderate (M 4-5) earthquakes beneath Upper Hutt in 2004 and 2005. We have not recognised similar triggering for any other New Zealand SSE to date. However, it has been discovered in a companion project that very small earthquakes (M ~2) do occur in conjunction with the Gisborne SSEs. Also, preliminary results from other projects suggest that the Gisborne earthquake of 20 December 2007, which occurred within the subducting Pacific Plate, may have triggered a slow-slip event on the plate interface above.

LAYMAN'S ABSTRACT

Slow-slip events, sometimes described as slow earthquakes, are a phenomenon that has been suspected for many years, but it is only in the past decade that scientists have found really good evidence for them. The evidence is from observations of ground displacement made by continuously-operating GPS instruments established on survey monuments that are well connected to the ground.

A typical scenario for a GPS station near Gisborne is that it will be moving steadily eastward at, say, 20 millimetres per year as a result of the push of the Pacific tectonic plate moving eastwards and downwards (or subducting) beneath the North Island. The station records this steady motion for perhaps two years, then over a 1-2 week period it moves back about 20 millimetres to the west. We believe this is due to slip between the two sides of the interface between the Pacific plate and the North Island. The motion of about 20 millimetres at the surface corresponds to about 200 mm of slip on the plate interface. The slip is something like the slip that occurs between the two sides of a geological fault during an earthquake, except that it takes weeks to occur, rather than the seconds it takes in an ordinary earthquake. In a slow-slip event, the slip is so slow that no damaging earthquake waves are produced. So slow-slip events are a "safe" way for the two sides of a geological fault to slip past each other. Further south in the North Island we have recorded slow-slip events that are even slower – they have durations of 6 months to more than a year.

Slow-slip events are a "good thing" in that any slip that occurs slowly is no longer available to occur in a future damaging earthquake. They may also be a good thing in helping us to understand the maximum size of some future earthquakes. It's also possible that a slow-slip event on a deeper part of the plate interface could act as a trigger for a damaging earthquake on a shallower part of the interface. For this reason it is a good idea to monitor slow-slip events carefully, as well as small earthquakes that occur in the region of the event.

To understand the potential size of a future earthquake on the plate interface that lies under the eastern and southern North Island, it is important to map out those areas of the interface that are presently stuck together (or "coupled"), as it is these regions that can potentially slip in a major earthquake in the future. Progress has been made on mapping these areas in recent years. Slow-slip events may provide additional information because they seem to occur at the boundaries between (usually shallower) areas on the plate interface that are well coupled and (usually deeper) areas where the two sides are slipping freely past each other, and therefore not building up towards an earthquake.

During this project we have interpreted the ground motion recorded during the large 2004-05 Manawatu slow-slip event, as well as three other events that occurred near Hastings and Gisborne during 2006. This has allowed us to define their location, depth, duration and magnitude, as has helped to delineate the boundaries between coupled and freely-slipping parts of the plate interface.

The slow-slip events along the east coast occur at very shallow depths (10-15 km) and tend to be quite short (days to weeks). Those under Manawatu and along the Kapiti Coast are much deeper (30-50 km) and longer (many months). We have attempted to determine what causes the variation in depth. It has been assumed elsewhere that temperature is a primary control on the depth at which slow-slip occurs, but we have shown that this is not the case beneath the North Island.

We have worked towards mapping the extent of the strongly-coupled zone of the subduction interface. We have used both the locations of slow-slip events and inferences from very accurate micro-earthquake locations, and we find that these give reasonably consistent results.

We have developed an automated method of detecting the onset of slow-slip events soon after they begin. When an event is detected this allows us to monitor the region of the event carefully to see if it is having any effect on earthquakes in the vicinity.

We think that a slow-slip event off the Kapiti Coast in 2003-04 triggered the occurrence of moderate (M 4-5) earthquakes beneath Upper Hutt in 2004 and 2005. We have not recognised similar triggering for any of the more recent New Zealand slow-slip events.

KEYWORDS

Slow-slip events, slow earthquakes, interplate coupling, Hikurangi subduction zone.

1.0 INTRODUCTION

Great subduction interface earthquakes are often the largest and most damaging on Earth, as was demonstrated to terrible effect in the Indian Ocean region on 26 December 2004. Areas of the Hikurangi subduction interface that are currently well and less-well coupled[1] have been mapped out in recent years (Figures 1, 2; see also Reyners, 1998 and Wallace et al., 2004). Well-coupled regions are those on which we expect major or great earthquakes to occur. Less well-coupled regions are expected to have smaller earthquakes, or much rarer large earthquakes. Understanding the coupling distribution, and how it may change with time, is therefore critical for assessing seismic hazard from subduction interface earthquakes.

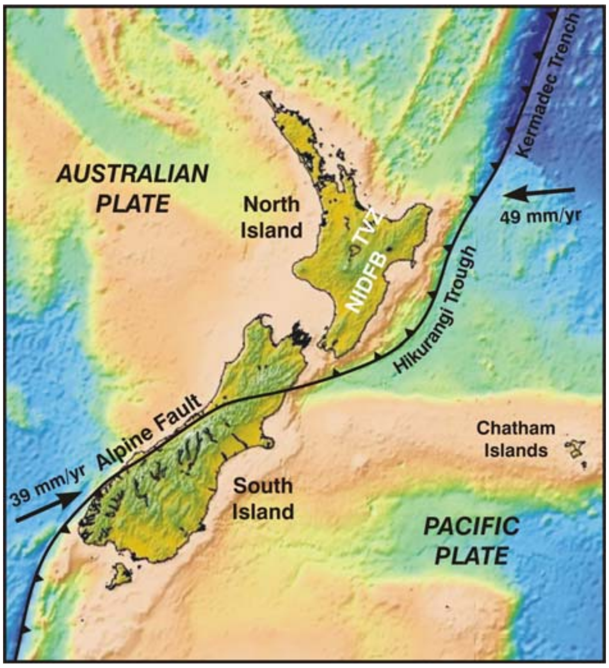

Figure 1 Tectonic setting of New Zealand. Orange areas are submerged continental crust. Subduction of the Pacific Plate beneath the North Island occurs at the Hikurangi Trough.

[1] A region of a subduction zone that is 100% seismically coupled means that 100% of the long-term plate motion on that section of the subduction zone is accommodated by earthquakes. Seismic coupling of 0% means that none of the plate motion is accommodated by earthquakes – the motion between the plates must be accommodated by some form of slow deformation process (aseismic slip or shearing), which may be continuous or episodic in nature. It is only possible to accurately measure seismic coupling if high quality earthquake-occurrence data are available over several major earthquake cycles. Only a few locations on Earth (e.g., parts of Japan) have such data, and these do not include New Zealand.
An alternative measure of interplate coupling is possible from interpretation of present-day surface deformation measurements, of which GPS is the most accurate and widely-used method at present. If the rocks of the subducting and over-riding plates are locked together (or 100% coupled), the inexorable ongoing motion of the descending plate causes elastic strain in the over-riding plate, which in turn causes deformation at the ground surface. If the two plates are slipping past each other freely (0% coupled), then there are no elastic strains transmitted and no deformation at the ground surface. By modelling the measured motions of a set of survey marks (using GPS, for example) it is possible to calculate the degree of coupling on the plate interface, and even to map out regions of the plate interface where the coupling is high and regions where it is low. It is tempting to think that the interplate coupling interpreted from surface deformation data may be directly equivalent to seismic coupling. But these interplate coupling estimates are assessed from deformation data collected over just the last 10 or so years, and we do not currently know if or how the degree or distribution of interplate coupling varies over an earthquake cycle.

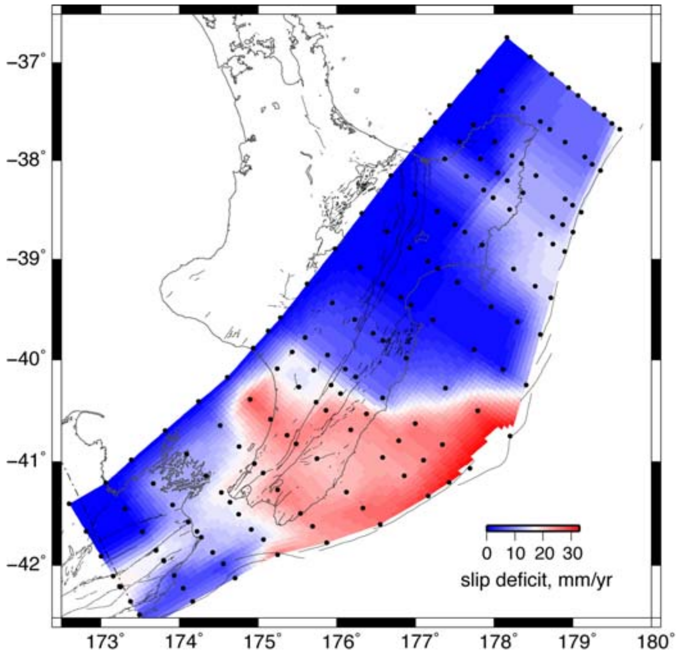

Figure 2 Distribution of interplate coupling on the Hikurangi subduction interface, estimated from ~12 years of campaign GPS site velocities using an interacting elastic crustal block model. The dots are the nodes at which slip deficit is estimated, with bilinear interpolation between nodes. Red areas show a slip deficit of 30 mm/yr, meaning this amount of slip is being stored as elastic energy for eventual release as slip on the subduction interface, probably in a large earthquake. Dark blue implies slip is occurring steadily with no elastic strain build up. The coupling distribution offshore of the eastern North Island is not well resolved. There is strong coupling on the strike-slip faults in the northern South Island and southern North Island, but this is not shown on this figure.

Ever since the first continuous GPS stations were installed above subduction interfaces in 1995, workers in Japan, Canada, the western U.S., Mexico, New Zealand and Costa Rica have observed episodes of non-linear deformation at the ground surface (e.g., Hirose et al., 1999; Dragert et al., 2001; Ozawa et al., 2003; Obara et al., 2004; Larson et al., 2004; Douglas et al., 2005). These have been interpreted to result from slow-slip events, or “slow earthquakes”, on the subduction interface. The slip is called “slow” because it occurs much more slowly than the slip rate in normal earthquakes. However, it can be substantially faster than normal tectonic plate motion. More than a dozen slow-slip events have been observed since mid 2002 in the North Island of New Zealand, at locations ranging from the Raukumara Peninsula to the Kapiti Coast.

These slow-slip events, which may be an important contributor to the seismic moment release budget at subduction zones, appear to occur in the transition zone between regions that are “coupled” (and possibly building up towards large earthquakes) and regions that are aseismically “creeping” (e.g., Dragert et al., 2001; Obara et al., 2004). It has also been suggested that redistribution of stress as a consequence of slow-slip can cause earthquakes to occur on other parts of the plate boundary (Dragert et al., 2001).

In Japan and western North America, the slow-slip events have been associated with a seismic noise signal that has some characteristics in common with the volcanic tremor that is often observed around active volcanoes (Obara, 2002). This is referred to as subduction zone episodic tremor. It appears to be spatially and temporally correlated with slow-slip events (Rogers and Dragert, 2003; Obara et al., 2004; Kao et al., 2005). By analogy with volcanic tremor, it indicates that fluid flow may be one of the physical processes involved in the overall slow-slip phenomenon, though subduction-zone episodic tremor is not harmonic like volcanic tremor.

At the same time, seismologists studying Japanese subduction zones have demonstrated the usefulness of clusters of small earthquakes near the plate interface in defining parts of the interface which are undergoing episodic slip. Often these small earthquakes have identical waveforms, and they have been interpreted as repeated rupture of small asperities (rough spots) on the plate interface resulting from slow-slip in the surrounding region (Uchida et al., 2003). Thus small earthquake clusters provide a useful addition to GPS measurements in defining which parts of the interface may undergo episodic slip.

1.1 Objectives

With this background, the original objectives of our project were:

1. Model any slow-slip events that are detected by GeoNet continuous GPS stations to determine their location, slip magnitude and direction, and time history.
2. Start to develop ways to detect slow-slip events in real time using continuous GPS and seismic tremor observations. This objective will be addressed in close collaboration with John Townend's Marsden-funded study of seismic tremor beneath the East Coast.
3. Investigate the spatial and temporal correlation of small earthquake clustering with episodes of slow-slip, and identify the faults on which earthquakes are clustering using double-difference location techniques.
4. Investigate whether the earthquake clustering can be explained simply by stress triggering as a result of slow-slip events, or requires another driver such as fluid flow.
5. Investigate the implications of slow-slip events and small earthquakes for the occurrence of major or great earthquakes on the subduction interface.
6. Elucidate, in conjunction with John Townend's Marsden-funded study at Victoria University, the association between slow-slip events, clusters of small earthquakes, and subduction zone tremor.

By its nature, this was an open-ended project. We have made excellent progress on objectives 1-4 and good progress on objectives 5-6. Three papers have been published in high-quality international journals that rely on work funded entirely or in part by this project, and a fourth manuscript has been submitted. A fifth paper was published in an International Association of Geodesy compilation.

2.0 TIME HISTORY AND MODELLING OF SLOW-SLIP EVENTS

Our work under this heading comprised two general areas. One was the regional filtering of the daily coordinate time series of the GPS stations, so that slow-slip signals could be more easily recognised and modelled above the noise in the time series. This aspect was funded largely by a FRST contract for GeoNet development and in part by this project. The second was the actual modelling of the time series.

We completed the modelling of the large Manawatu slow-slip event that had occurred over an 18 month period in 2004-2005, as is fully reported by Wallace and Beavan (2006). The Manawatu slow-slip event caused up to 30 mm of horizontal and vertical displacement of several CGPS sites in the North Island. We inverted these displacements to estimate slip on the subduction interface, and up to 350 mm of slip over an area ~75 km x 100 km was required to fit the data. If all of the slip occurring in the Manawatu event had occurred in a single earthquake, it would have resulted in a Mw 7.0 earthquake. Thus, events like the Manawatu SSE could be major contributors to the overall moment release budget at the Hikurangi subduction margin. To further examine the time evolution of the event, we separated it into three subevents, and inverted each of them separately. We found that most of the slip occurred during the last six months of the event, and that in general the location of slip migrated up-dip along the subduction interface with time. We also modelled the three other slow-slip events that occurred during the project period using similar methods; these results are described below.

2.1 Data cleaning

GPS time series contain signals from a variety of sources, as well as the ground deformation signals of interest for geophysical interpretation. One source that has been recognised since the work of Zhang et al. (1997) is referred to as “regional common-mode noise”. This is a signal that appears more-or-less the same at GPS stations over a region whose dimension may be hundreds of km to a thousand km or so. The sources of this type of signal may be varied, such as: errors in the satellite orbits; regional or global-scale mass movements within the ocean, atmosphere or hydrosphere; inaccurate models within the GPS modelling software; and probably others. The common-mode signal may be removed by averaging, or in more sophisticated implementations by principal component analysis. We have followed a simple method similar to that described by Beavan (2005). Some care is needed not to include stations that have large non-linear signals (e.g., GISB, DNVK, WANG) when calculating the average. Because the network is expanding, the number of stations contributing to the average also changes with time. The funding for this work has come largely from the FRST contract for GeoNet development, and in part from this EQC contract. (As an aside, these “regionally-filtered” time series have been made available publicly at the GeoNet web site since May 2008, in addition to the “raw” time series that have been available for some years).

2.2 Modelling of slow-slip events

Figure 3 shows the locations and estimated slip distributions of the slow-slip events that have been recognised between 2002 and early December 2007. Table 1 gives additional details of the events that occurred during the project period – one near Gisborne and two near and south of Hastings.

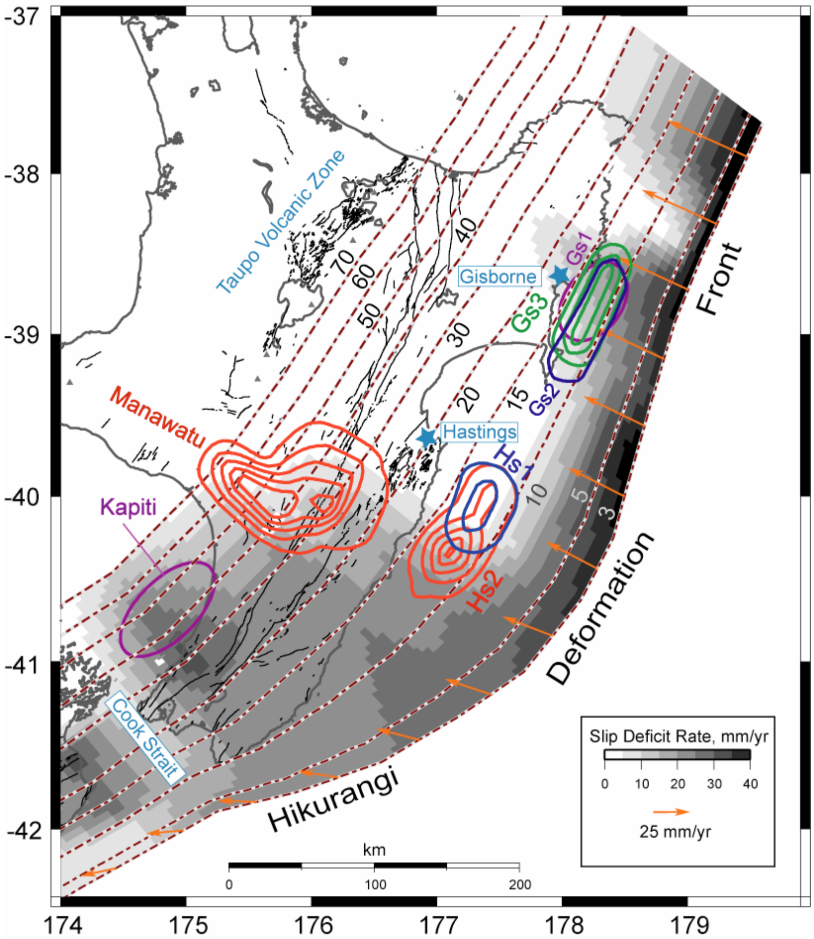

Figure 3 Estimated slip distributions of major NZ slow-slip events through November 2007; two other events (not shown) started in December 2007. Slab contours (in km, dashed lines), interplate coupling distribution (shown as slip rate deficit), convergence vectors (red arrows) and locations of slow-slip events (coloured contours). Locking distribution is updated from Wallace et al. (2004). Slow-slip events are shown as 50 mm contours of slip on the plate interface, and as ellipses when only the general location is known (Gs1 and Kapiti). The times of the events are given below.

Kapiti	2003-04
Gisborne, Gs1	Oct 2002
Manawatu	2004-05
Gisborne, Gs2	Nov 2004
Hastings, Hs1	Jun 2006
Gisborne, Gs3	Jul 2006
Hastings, Hs2	Aug 2006

Table 1 Parameters of slow-slip events that occurred during 2006-2007

Date[a]	Region	Depth[b], km	Duration[c], d	Max slip, mm	M_w[d]
20060610	off Hastings (Hs1)	15	7	120	6.6
20060707	off Gisborne (Gs3)	15	12	160	6.7
20060824	south of Hastings (Hs2)	15	8	250	6.8

[a] Date is the approximate date at which the event becomes noticeable in the GPS position time series.
[b] Depth is the approximate deeper limit of slip; the shallower limit is not well constrained.
[c] Duration is the approximate interval over which motion is noticeable in the time series.
[d] Approximate moment release, using $\mu = 4\times10^{10}$ Nm.

3.0 AUTOMATED DETECTION OF SLOW-SLIP EVENTS

We have developed and implemented an algorithm to detect the onset of slow-slip events on the Hikurangi subduction interface. The procedure is run daily on the updated regionally-filtered GPS time series with the goal of detecting slow-slip as soon as possible after onset. This procedure is undertaken internally at GNS, but there is the potential to include the detection algorithm as part of GeoNet's ongoing operations.

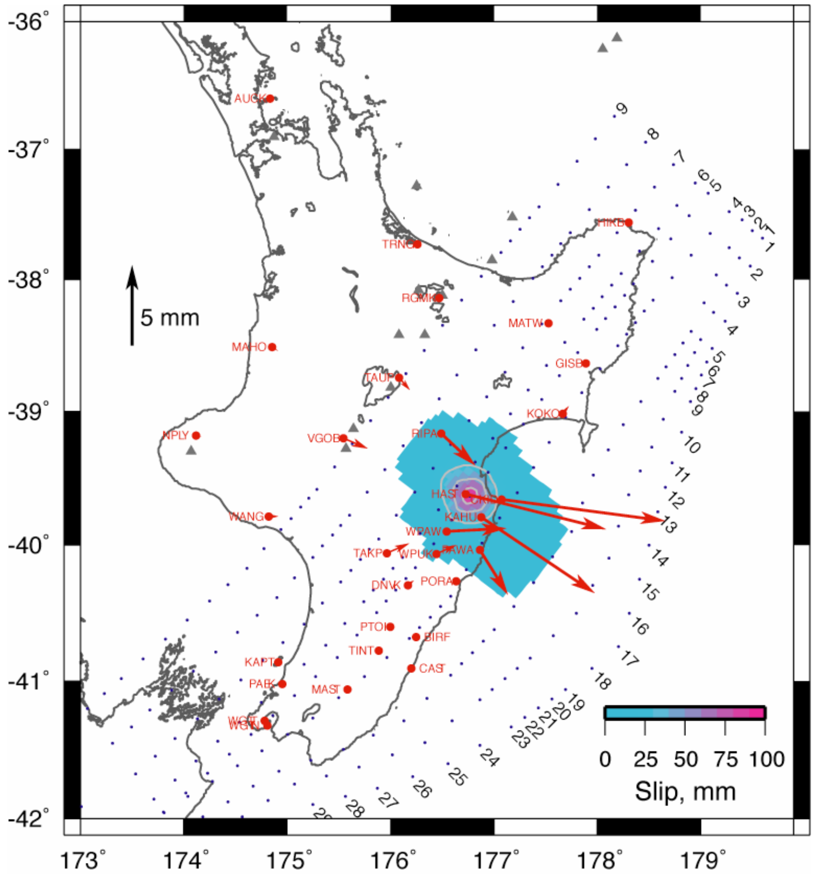

Figure 4 Example of a synthetic slip event on the Hikurangi margin. The peak slip on the plate boundary is 100 mm and decreases away from the central point (slip contours shown by shading). Red arrows show pattern of surface displacements at GPS sites expected from the hypothetical slip event.

The basis of the method is cross-correlating the geodetic time series with synthetic Green's functions that represent slow-slip on some part of the plate interface during some specified time interval. The synthetic slow-slip events are temporally characterised by a ramp function of variable duration, in order to detect events of short (days) or long (months) duration. At times or places where the GPS time series comprise signals or random noise that do not have the spatial and temporal characteristics of the synthetic events, the cross-correlation factor (CCF) will be small or negative. Where the GPS time series contains deflections that are consistent with the slow-slip event the CCF will become larger and will be largest at the time and place of the greatest slip (Fig. 4).

The cross-correlation function (CCF) is:

$$CCF(t) = \sum_{i=1,s} \sum_{k=1,3} \sum_{n=1,r} R(n) D_{ik}(t+n) G_{ik}$$

where the sums are over the numbers of sites (s), number of components (3) and the duration (in days) of the ramp (r). R, the ramp function, is a linear ramp from zero to 1 over r days, D_k is the kth component of the GPS time series, t is time in days, and G_{ik} is the expected displacement at station i and component k due to the synthetic slip event. Synthetic slip events are generated at a grid of points covering the Hikurangi subduction interface (Fig. 4) and the CCF is calculated separately for each source.

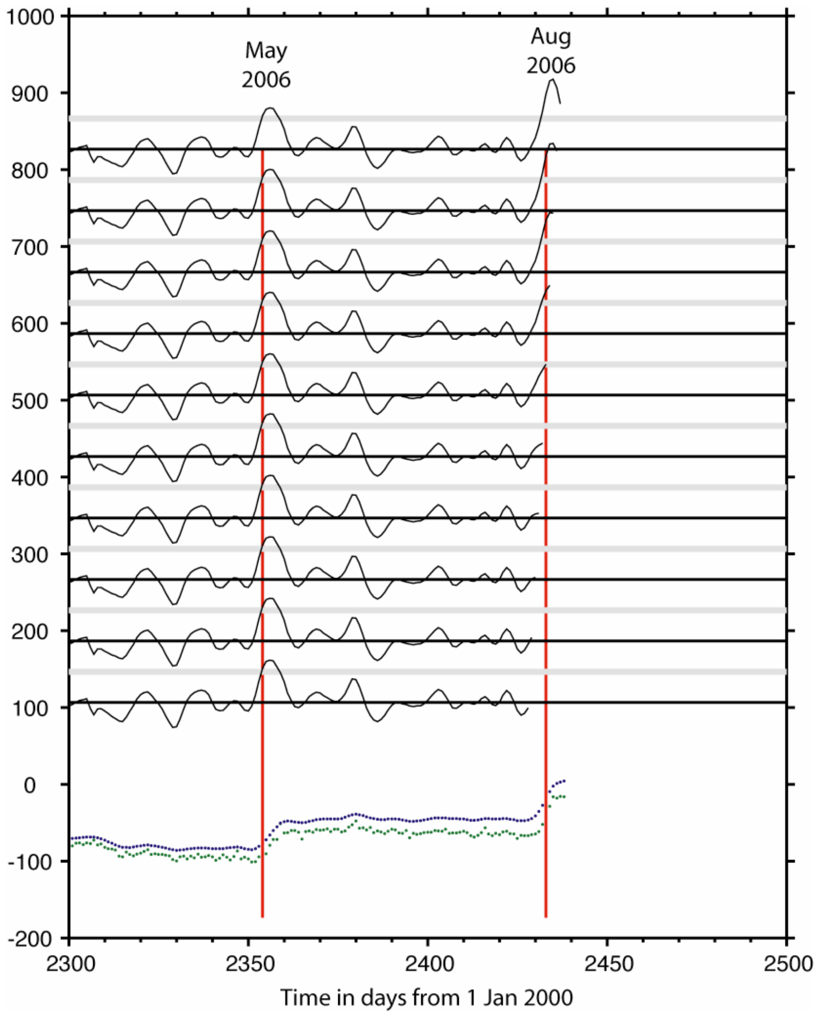

Figure 5 Cross-correlation functions (CCF) for two slip events offshore of Hastings (May and August 2006) using the 3-day ramp function. The time series at bottom is the east component of site PAWA (blue is low-pass filtered, green is unfiltered). The black curves are the CCF calculated using all available GPS data (not just PAWA) for time periods ending at days 2436 (bottom curve) through 2445 (top curve). In each case, the gray horizontal bar is 1 unit above zero (black horizontal line). The vertical red lines show the day at which the CCF reaches a value of 1, our initial detection threshold. For the first event, the detection threshold is reached about 3 days into the event; for the second it is about 4 days.

We tested the algorithm against slow-slip events that occurred in 2006 (Fig. 5) by running the data stream through it to simulate real-time analysis. For these short-duration events the detection was triggered within a few days. These and other tests suggest the short-duration events are detected when the 3-day ramp function reaches a CCF of about 1 and the 7-day ramp CCF is about 0.5. A long duration, deep event began beneath the Kapiti region around the end of November 2007. The peaks in the CCF were seen on December 3rd to be about 0.57 (3-day ramp) and 0.30 (7-day ramp), which are lower than the threshold set earlier. The CCF is sensitive to the number of nearby sites and the emergence of the slip time function. Further calibration is required to optimally set the thresholds.

3.1 Slow-slip events and episodic tremor

Our original hope had been to use automated recognition of episodic tremor, in addition to modelling of GPS time series, to detect the occurrence of slow-slip events. Episodic tremor has been associated with slow-slip events in a number of overseas cases, particularly in Japan, Canada and the Pacific Northwest of the USA. We had planned to take tremor results from the Marsden-funded project on episodic tremor being led by John Townend at Victoria University. However, in careful examination of seismic recordings from the 2004 and 2006 Gisborne slow-slip events, and a representative month during the 2004-2005 Manawatu event, no tremor signal was detected. Instead of tremor, variations in microseismicity (typical magnitude M_L 2) were observed in some cases, and these have been related to the slow-slip events through a rate- and state-dependent friction model (Delahaye et al., 2006; Delahaye, 2008; Delahaye et al., 2008). Until greater understanding has been developed on the seismic signals that may or may not be associated with slow-slip events in New Zealand, we cannot make use of these signals for slow-slip detection purposes.

4.0 MECHANISM OF SLOW-SLIP EVENTS – THE EFFECT OF TEMPERATURE

Slow-slip events are thought in general to occur within the transition zone from unstable sliding (stick-slip, where earthquakes occur) to stable sliding (where creep occurs). Most conceptual models and laboratory tests cite temperature as one of the more important parameters among those that determine the frictional stability field. A commonly accepted temperature range for the stability transition is 350ºC to 450ºC in granitic rocks and possibly higher in gabbro.

The first well-documented slow-slip event at the Hikurangi margin was offshore of Gisborne in October 2002, where the plate interface is less than 15 km depth. Since then, several more shallow events have been recorded, offshore of both Gisborne and Hastings. Given our expectation (based on previous studies; Hyndman et al., 1997; Oleskevich et al., 1999) that the temperatures at which such slip occurs should be at least 350°C, we estimated temperatures of the Hikurangi plate interface where slow-slip is occurring. Adopting an advection model of England and Molnar (REF), we take the age of the incoming lithosphere as a proxy for its thermal structure and use the rate of convergence and the specific geometry to estimate temperature. Using heatflow estimates from the forearc to guide the calculations, we also added a reasonable amount of heating from shearing on the plate interface (Fig. 6). From these calculations, we estimate that the shallow Hikurangi slow-slip events occur at temperatures of no more than 150ºC. As far as we know, this observation is unique globally and requires some rethinking of the role of temperature in stabilizing frictional regimes on subduction faults.

The work summarised in this section is described in greater detail in McCaffrey et al. (2008).

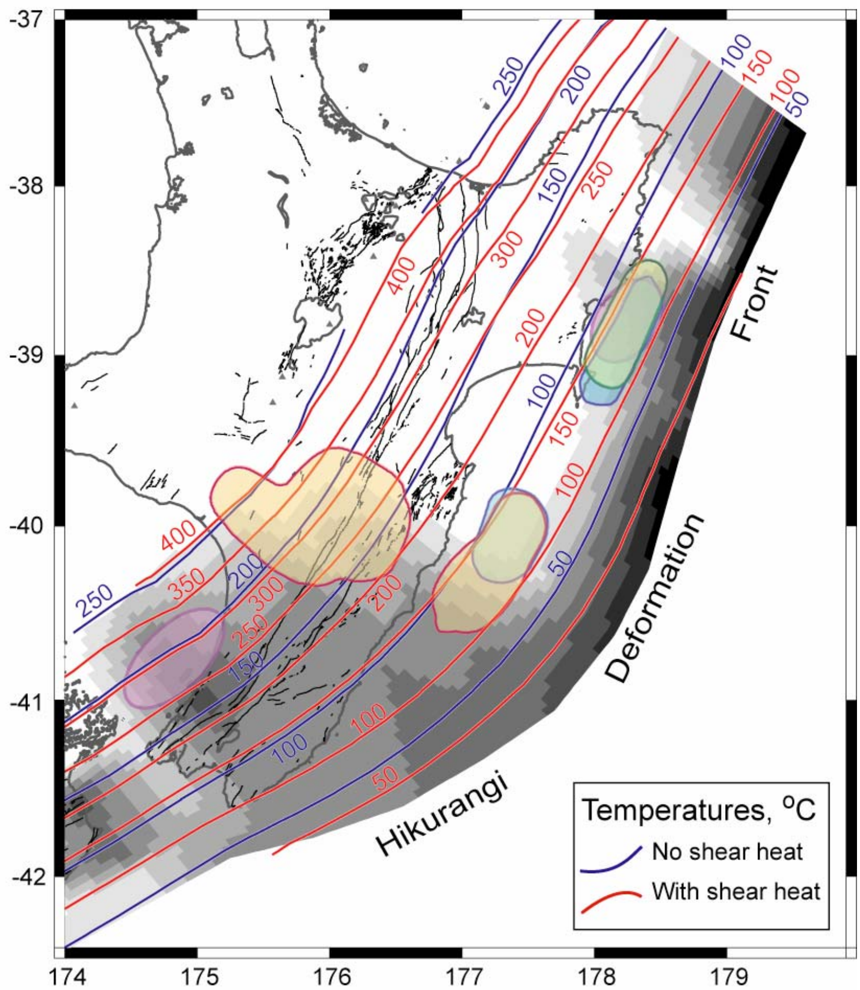

Figure 6 Estimates of temperature (red and blue contours) on the Hikurangi thrust interface and the general locations of slow-slip events (coloured regions). The gray areas are the locking distribution from Wallace et al. (2004). Temperatures were estimated by taking into account the age and rate of the incoming subducting lithosphere and the geometry of the margin. Blue temperature contours do not include shear heating while red contours include heating from 20 MPa of shear stress on the plate interface. In either case, the slow-slip events off the coast of the north and central North Island occur at temperatures lower than about 150ºC.

5.0 WHAT DO SMALL EARTHQUAKES TELL US ABOUT INTERPLATE COUPLING?

In parallel with the discovery of slow-slip events, there has been a realization that small earthquakes near the plate interface are useful in defining parts of the interface which are undergoing episodic slip (e.g., Igarashi et al., 2003; Uchida et al., 2003). These small earthquakes are often known as “repeaters”, as they are interpreted as repeated rupture of small asperities on the plate interface resulting from slow-slip in the surrounding region. The usefulness of such repeaters was recently demonstrated after the M_W ~8.0 Tokachi-oki earthquake, which ruptured the plate interface off Hokkaido. Afterslip in the region surrounding this rupture was associated with a rich sequence of small repeating earthquakes (Matsubara et al., 2005). Indeed, repeating earthquakes can provide episodic slip data in offshore areas where resolution from land-based GPS is poor.

To better understand the nature of interplate coupling in the southern North Island of New Zealand, we have compared the distribution of relocated small earthquakes with both (1) interseismic geodetic locking of the plate interface determined from GPS, and (2) structure within the overlying plate from tomographic inversions. Relocated seismicity shows a good correlation with both the distribution of strong geodetic locking and geological terranes in the overlying plate. It appears that the strong Permian/Triassic Rakaia geological terrane may control frictional coupling at the plate interface. It also modulates the small earthquake distribution within the underlying slab, promoting clustered seismicity in the slab crust and suppressing lower plane seismicity in the mantle. The seismicity and structural data suggest a model where plate coupling is controlled by the ability of fluid to cross the plate interface. When an impermeable terrane in the overlying plate prevents such fluid flow, plate coupling appears to be strong. This work is described in more detail in Appendix B.

6.0 IS THERE A RELATIONSHIP BETWEEN SLOW-SLIP AND EARTHQUAKES?

The Kapiti Coast slow-slip event of 2003-2004 (Beavan et al., 2007) appears to have triggered a swarm of earthquakes beneath Upper Hutt in April-May 2004 (largest event M_L 4.6). A nearby M_L 5.5 earthquake several months later in January 2005 appears to have been triggered by stress changes due to the swarm. High precision double-difference relocation of these events has revealed that they occurred on normal faults within the top of the subducted plate, rather than on the plate interface itself. Further, the swarm appears to have involved incremental slip on adjacent patches of the fault, with slip propagating to shallower depths with time. The slightly greater depth of the January 2005 mainshock relative to the swarm suggests a rapid decrease in mechanical damage with depth in the subducted plate. It is suggested that the monitoring and mapping of this type of swarm activity might be another tool in mapping the distribution of the strongly coupled portion of the subduction interface. For a fuller development of these arguments the reader is referred to Reyners and Bannister (2007). (This work was started by Martin Reyners under a previous EQC contract, but was completed and published under this contract.)

7.0 PUBLICATIONS AND PRESENTATIONS RESULTING FROM THIS PROJECT

7.1 Published papers

Beavan, J., Wallace, L., Douglas, A., Fletcher, H., and Townend, J., 2007. Slow-slip events on the Hikurangi subduction interface, New Zealand, Dynamic Planet, Monitoring and Understanding a Dynamic Planet with Geodetic and Oceanographic Tools, IAG Symposium, Cairns, Australia, 22-26 August, 2005, Series: International Association of Geodesy Symposia, P. Tregoning & C. Rizos (Eds.), Vol. 130, 438-444.

McCaffrey, R., L. M. Wallace and J. Beavan, 2008. Slow-slip and frictional transition at low temperature at the Hikurangi subduction zone, Nature Geoscience, 1, doi:10.1038/ngeo178.

Reyners, M., and S. Bannister, 2007. Earthquakes triggered by slow-slip at the plate interface in the Hikurangi subduction zone, New Zealand, Geophys. Res. Lett., 34, L14305, doi:10.1029/2007GL030511.

Wallace, L. M., and J. Beavan, 2006. A large slow-slip event on the central Hikurangi subduction interface beneath the Manawatu region, North Island, New Zealand, Geophys. Res. Lett., 33, L11301, doi:10.1029/2006GL026009.

7.2 Manuscripts submitted

Reyners, M., 2008. What do small earthquakes tell us about the nature of interplate coupling in the southern North Island, New Zealand?, to be submitted to Geophys. Res. Lett.

7.3 Abstracts of presentations

McCaffrey, R., L. Wallace, J. Beavan, and A. Douglas, 2007. Slow-slip events, temperature, and interseismic coupling at the Hikurangi subduction zone, New Zealand, Eos Trans AGU, May 2007.

McCaffrey, R., J. Beavan, and L. Wallace, 2007. Automated detection of slow-slip events at the Hikurangi subduction zone with continuous GPS time series, GSNZ/NZGS joint conference, Tauranga, November 2007, Geol. Soc. NZ Misc. Publ., 123A, p 96.

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Reyners, M., 2007. What do small earthquakes tell us about plate coupling in the southern North Island?, GSNZ/NZGS joint conference, Tauranga, November 2007, Geol. Soc. NZ Misc. Publ., 123A, p 139.

8.0 ACKNOWLEDGEMENTS

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Beavan, J., Wallace, L., Douglas, A., Fletcher, H., and Townend, J., 2007. Slow-slip events on the Hikurangi subduction interface, New Zealand, Dynamic Planet, Monitoring and Understanding a Dynamic Planet with Geodetic and Oceanographic Tools, IAG Symposium, Cairns, Australia, 22-26 August, 2005, Series: International Association of Geodesy Symposia , Tregoning, Paul; Rizos, Chris (Eds.), Vol. 130, 438-444.

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APPENDICES

APPENDIX A NEW ZEALAND SLOW-SLIP EVENTS THROUGH 2005

This appendix contains a preprint version of the published paper:

Beavan, J., Wallace, L., Douglas, A., Fletcher, H., and Townend, J., 2007. Slow-slip events on the Hikurangi subduction interface, New Zealand, Dynamic Planet, Monitoring and Understanding a Dynamic Planet with Geodetic and Oceanographic Tools, IAG Symposium, Cairns, Australia, 22-26 August, 2005, Series: International Association of Geodesy Symposia , Tregoning, Paul; Rizos, Chris (Eds.), Vol. 130, 438-444.

It is reproduced here because the published version is in an IAG Symposium series that is not readily available.

Slow-slip Events on the Hikurangi Subduction Interface, New Zealand

J. Beavan, L. Wallace, H. Fletcher
GNS Science, P O Box 30368, Lower Hutt, New Zealand

A. Douglas
School of Earth Sciences, Victoria University of Wellington, P O Box 600, Wellington, New Zealand

Abstract. In common with other regions where continuously-recording Global Positioning System (CGPS) networks have been established above subduction zones, several aseismic deformation episodes have been observed in New Zealand since 2002. We interpret these episodes to result from slow-slip on the subduction interface, though with the current density of CGPS stations the details of most events recorded to date are not well resolved.

We have observed events with accompanying surface displacements ranging from 5-30 mm magnitude, and lasting several days to more than a year. Modelling suggests that the events are occurring near the down-dip end of the locked seismogenic part of the subduction zone, in the transition zone between the interseismically coupled and creeping portions of the interface.

Comparison of event sizes, inter-event deformation rates, and long-term deformation rates suggest a repeat time of 2-3 years for an October 2002 event recorded near Gisborne in the northern Hikurangi margin. A similar-sized event recorded in November 2004 supports this estimate. Two longer-duration slow-slip events beneath the central and southern North Island may have triggered a series of small to moderate earthquakes over the past two years.

There is preliminary indication of seismic tremor associated with the Gisborne events, as has been observed in Japan and western North America, but more work is needed to confirm this.

The best-documented slow-slip event in the North Island occurred beneath the Manawatu region, and lasted from early 2004 until June 2005. The event caused displacements at up to seven CGPS sites, and probably resulted from up to 300 mm of slip on the subduction interface.

Keywords. GPS, aseismic slip, slow earthquakes

1 Introduction

Transient fault slip episodes, occurring over much longer time periods (days, months) than earthquakes, have been recorded with continuously-recording Global Positioning System (CGPS) instruments located at several subduction margins on the Pacific Rim (e.g., Dragert et al. (2001); Ozawa et al. (2001, 2003); Larson et al. (2004)). The episodes are detected at the Earth's surface as non-linear motion of CGPS sites that is often rapid compared to normal tectonic plate motions. The physics of the deformation mechanisms underlying these so-called slow-slip events, or slow earthquakes, is not yet well understood. The events may trigger other types of deformation, such as actual earthquakes, and may make a significant contribution to moment release in subduction zones. Quantifying their size, location and frequency is therefore a key task in characterizing seismic hazard for subduction zones.

Subduction of the Pacific Plate occurs beneath the North Island of New Zealand, and at least five distinct slow-slip events have been observed at CGPS sites in the North Island over the past three years (e.g., Beavan et al. (2003); Douglas et al. (2005)). The CGPS sites have been established as part of the PositioNZ (www.linz.govt.nz/positionz) and GeoNet (www.geonet.org.nz) networks.

The slow-slip events have occurred in at least three different locations on the subduction interface (Figure 1). Some have lasted only a week, while others have continued for more than a year. Some events have caused small (~5 mm) displacements at the surface, while others have caused CGPS sites to move more than 30 mm. In all cases, the events seem to occur near the down-dip end of the well coupled, or seismogenic, part of the subduction interface, consistent with observations from Cascadia and Japan (e.g., Dragert et al. (2001); Ozawa et al. (2003)). We find that the recurrence intervals of some events may be two to three years, and that other events may take five years or longer to recur. The diverse characteristics of slow-slip events observed thus far at the Hikurangi margin highlights the importance of this location as a natural laboratory for understanding aseismic deformation events at subduction margins.

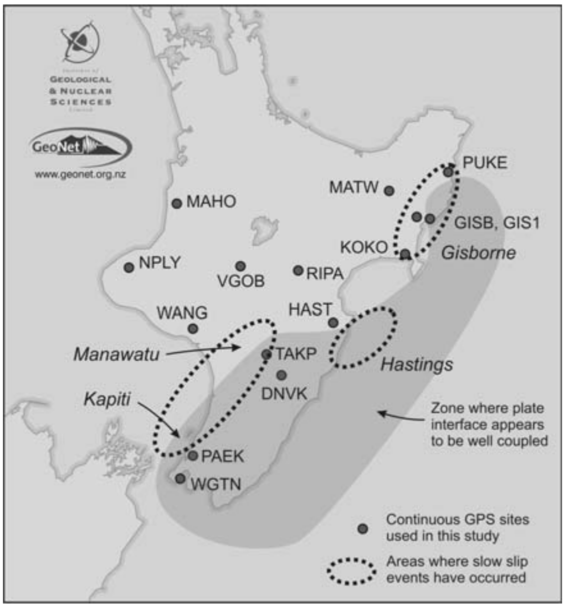

Fig. 1 Continuous GPS stations and regions where slow-slip events have been observed. Dark shaded area approximates the well coupled part of the subduction interface (Figure 3).

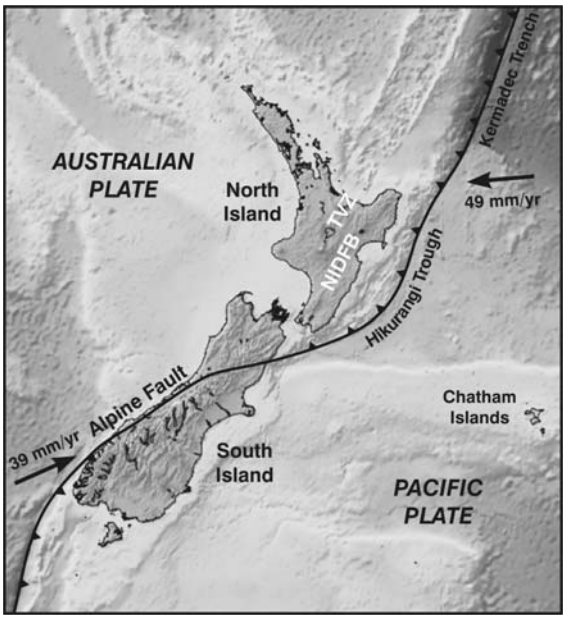

Fig. 2 Tectonic setting of New Zealand. Light coloured areas are submerged continental crust. Pacific Plate subducts beneath North Island at the Hikurangi Trough. TVZ: Taupo Volcanic Zone; NIDFB: North Island Dextral Fault Belt.

2 Tectonic setting and interseismic coupling distribution

The North Island of New Zealand lies in the boundary zone between the obliquely converging Pacific and Australian plates. North Island active tectonics is dominated by westward subduction of the Pacific Plate beneath the eastern North Island at the Hikurangi Trough (Figure 2). Clockwise rotation of the eastern North Island forearc leads to back-arc rifting in the Taupo Volcanic Zone (TVZ), while strike-slip faulting in the eastern North Island dextral fault belt (NIDFB) occurs due to the oblique convergence between the Pacific and Australian plates. Wallace et al. (2004) used campaign GPS data to estimate forearc rotation and interseismic coupling on the Hikurangi subduction interface. The coupled zone is wider beneath the southern North Island, and it narrows and shallows towards the north (Figures 1 and 3). Much of the North Island overlies the "transition zone" between the coupled and creeping portions of the subduction interface (see also Reyners (1998)).

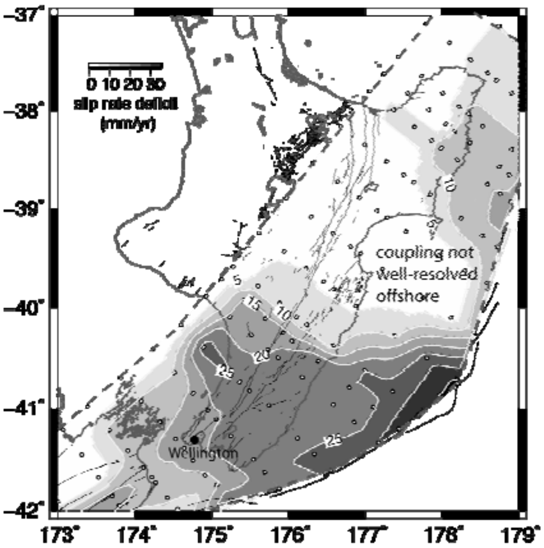

Fig. 3 Distribution of seismic coupling on the Hikurangi subduction interface, estimated from ~12 years of campaign GPS site velocities using an elastic crustal block model. The dots are the nodes at which slip deficit is estimated, with bilinear interpolation between nodes. Darkest areas represent slip deficit of 30 mm/yr, meaning this amount of slip is being stored as elastic energy for eventual release as slip on the subduction interface, probably in a large earthquake. White areas imply slip is occurring steadily with no elastic strain build up. The coupling distribution offshore of the eastern North Island is not well resolved. There is also strong coupling, not shown in the figure, on the strike-slip faults in the northern South Island and southern North Island.

3 Analysis techniques and time series

We process the CGPS data to give daily position estimates using Bernese version 4.2 and 5 software (Beutler et al. (2001); Hugentobler et al. (2004)). We use fixed IGS final orbits then place the daily coordinate results in a global reference frame using a least-squares fit to the ITRF2000 coordinates of a set of regional IGS stations. For the New Zealand sites, we then remove outliers and apply regional filtering to reduce remaining common-mode noise in the daily position time series, in an iterative process (Zhang et al. (1997); Beavan (2005)).

4 Gisborne events

In October 2002, a rapid (compared to normal plate motion) surface deformation event of 20-30 mm magnitude was observed over a 10 day period on two CGPS instruments near Gisborne (GIS1 and GISB; Figure 4). The event occurred in the preliminary stages of CGPS network development above the northern Hikurangi subduction zone so was not well recorded spatially. Forward modelling by Douglas et al. (2005) indicates that the event was probably due to about 180 mm of aseismic slip on the subduction interface just offshore of the Gisborne region. The event was fairly shallow, at ~10-14 km depth, but is consistent with slip occurring near the deeper end of the well-coupled part of the plate interface in this region. By balancing the magnitude of slip and the long-term plate motion, Douglas et al. (2005) predict that events of similar magnitude could recur every 2-3 years. A similar event was, in fact, recorded in November 2004, within the predicted interval.

Slow-slip events in Japan and Cascadia, have been associated with a seismic noise signal, or "subduction zone tremor" (Obara (2002); Rogers and Dragert (2003); Obara et al. (2004)). Douglas (2005) has made a preliminary attempt to identify such signals in regional broadband seismic data for the 2002 and 2004 Gisborne events. There does seem to be an indication of these signals, and further work is planned.

5 Hastings events

Several small displacement events have been observed at a CGPS site at Hastings (HAST) since its installation in late 2002 (Figure 5). These events are generally not observed at many CGPS sites because of poor spatial sampling. There is an indication that some of the Hastings events follow a few months later than Gisborne events, suggesting a possible along-strike migration with time similar to that observed by Dragert et al. (2001) in Cascadia and Ozawa et al. (2003) on the Boso Peninsula. The deformation at HAST in late 2004 and early 2005 may result largely from the Manawatu event discussed below.

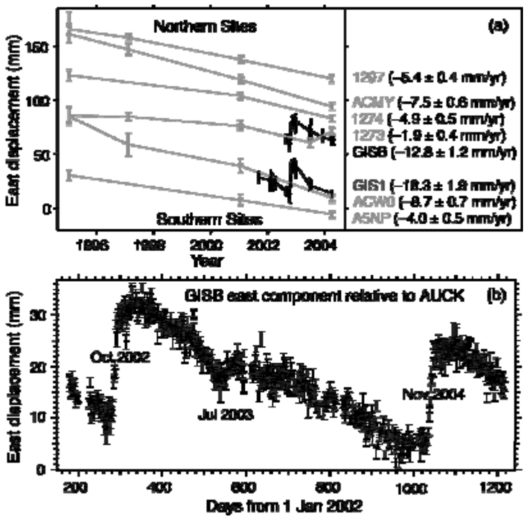

Fig. 4 (a) Position time series relative to a site near Auckland for campaign GPS sites north and south of Gisborne, and for selected days of CGPS sites GISB and GIS1 (Figure 1). The campaigns are too infrequent to detect the events seen at the CGPS sites. (b) GISB time series showing eastward surface displacements of 20-25 mm occurring over 7-10 day periods in 2002 and 2004, with a smaller and slower event in 2003.

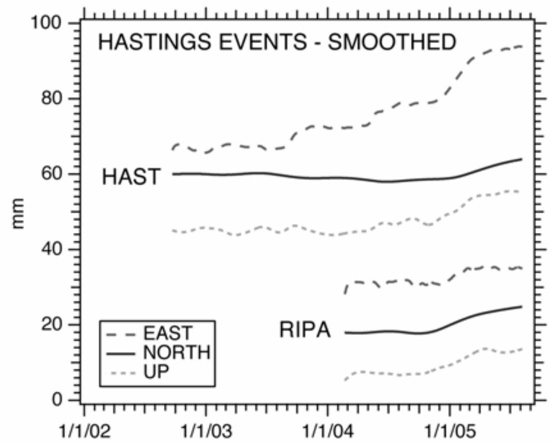

Fig. 5 Smoothed time series showing several deformation events detected at sites HAST and RIPA (Figure 1). Linear fits have been subtracted to make the velocities between events approximately zero. The 2003 event lags the similar Gisborne event by ~2 months. The early 2005 deformation is well modelled as part of the Manawatu slow-slip event.

6 Kapiti Coast event

A CGPS site at PAEK on the Kapiti Coast (Figure 6) moved steadily westward at 25 mm/yr (relative to the Australian Plate) from 2000 to about May 2003, when it suddenly slowed to only 15 mm/yr westward. At the same time PAEK began relative uplift at about 10 mm/yr. During the same interval, WGTN continued to show fairly steady westward motion at 30 mm/yr and no clear change in vertical motion. The changes at PAEK lasted about a year then the site resumed approximately its previous motion.

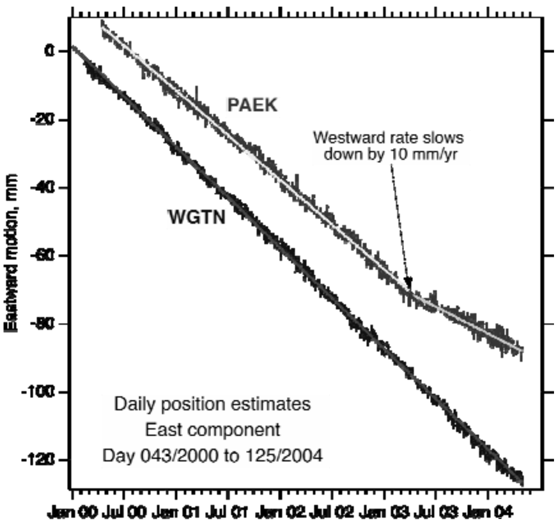

Fig. 6 Eastward motion of sites PAEK and WGTN (Figure 1) relative to the Australian Plate. PAEK slowed by ~10 mm/yr in about April 2003. At the same time its vertical (upward) rate increased by ~10 mm/yr. Soon after the end of this plot, the velocities returned to approximately their earlier values.

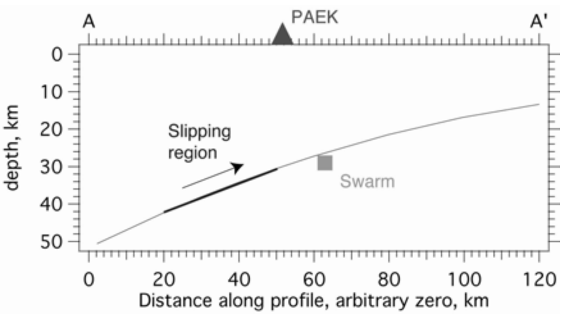

Fig. 7 Cross-section from northwest (A) to southeast (A′) showing inferred region of slow-slip, PAEK CGPS station, and the site of several earthquake swarms located near the top of the subducting plate. Coulomb stress changes calculated from the slip event favour the normal faulting style of the swarm earthquakes.

The signal can be explained by ~500 mm of slip occurring slowly on a deep part of the subduction interface between about 30 km and 40 km depth. In order to produce the observed 10 mm/yr change in horizontal motion at PAEK, while producing little change in motion at WGTN, the area of slip must be fairly small; in particular the southeastern and southwestern edges (see Figure 9) are fairly well constrained. Also, the inferred slip direction is oblique – it is approximately parallel to the plate motion direction (west-southwest) rather than being approximately down dip of the subducted plate (northwest). For oblique slip to occur on the deeper part of the subduction interface is reasonable, since the proposed slip event is occurring to the west of most of the upper-plate strike-slip faults.

Several swarms of earthquakes up to M_L ~5 have been felt in the region from mid 2003 to early 2005. The earthquakes are located up-dip of the supposed slow-slip event (Figure 7). They are predominantly normal faulting events and their hypocentres, estimated by double differencing, line up on two planes dipping approximately southeast within, but near the top of, the subducting plate (M. Reyners, pers. comm., 2004, 2005). Coulomb stress changes calculated for the postulated slow-slip event are in the correct sense to favour this style of faulting (R. Robinson, pers. comm., 2004). A plausible explanation for these observations is that the stress changes produced by the slow-slip event triggered incremental slip on pre-existing faults in the crust of the subducted plate.

7 Manawatu regional event

Up to seven CGPS sites in the central North Island experienced displacements (a few to 30 mm) during an eighteen month period from early 2004 until June 2005 (Figure 8). The sites with the largest movements were TAKP, WANG and DNVK, which straddle the Manawatu region – hence our name for the event. Additionally, TAKP and DNVK underwent at least 20 mm of uplift during this period. In Figure 8 we plot smooth curves as well as the original data. These are smoothing spline fits to the data, with the smoothing parameters chosen by trial and error. We have also subtracted a linear signal from the time series so the inter-event velocities are approximately zero.

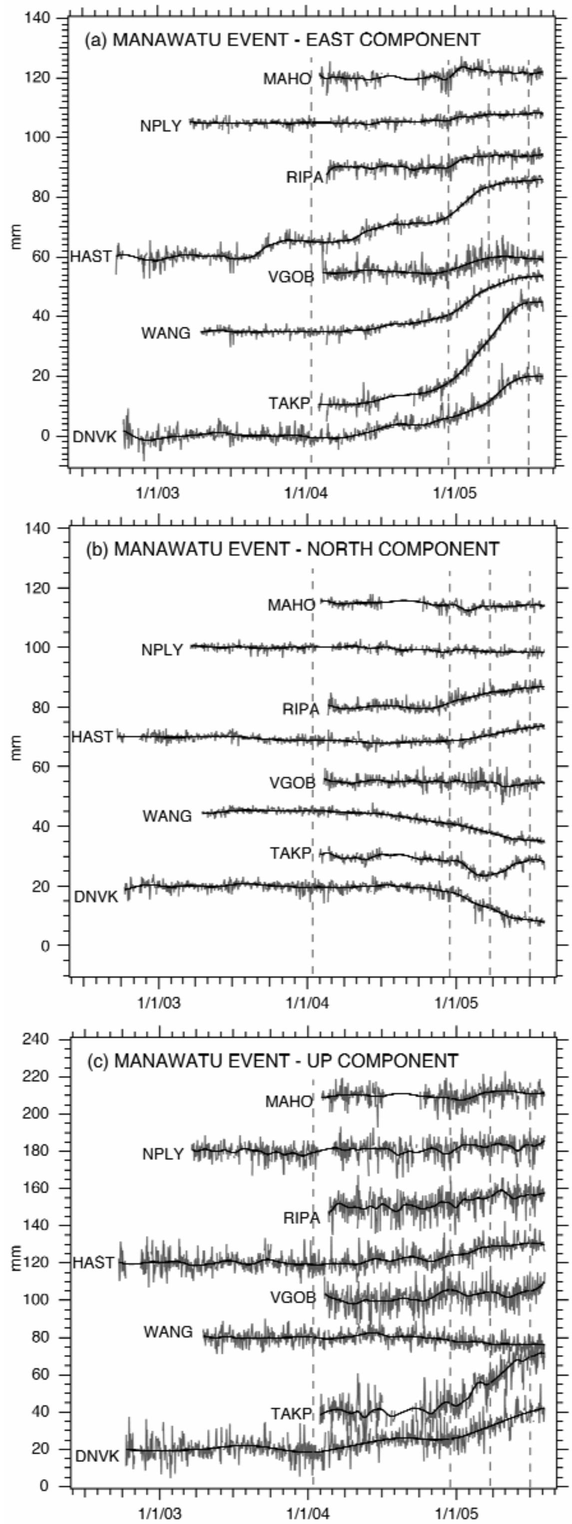

Fig. 8 Daily and smoothed time series from CGPS sites (Figure 1) affected by the Manawatu slow-slip event. The grey dashed lines split the event into three sub-events that we model separately.

To examine its time evolution, we split the Manawatu event into three parts (Figure 8). The first sub-event covers the period from early 2004 to January 2005. The displacements at TAKP, WANG and DNVK are less marked during this period compared to January-June 2005, but they are still significant. The second sub-event occurs from January-March 2005, and the third from March-June 2005. We separate the second and third sub-events by the change in direction of motion at TAKP, which is displaced southwards from January-March, and northwards from March-June.

We use software of McCaffrey (1995) to invert the observed displacements during each sub-event for slip on an array of dislocations in an elastic half-space (Okada (1985)). We assume that the slip occurs on the subduction interface defined from seismicity data by Ansell and Bannister (1996), and constrain the slip directions to be parallel to the direction of long-term relative motion on the interface estimated by Wallace et al. (2004). In the inversion, we solve for the amount of slip at nodes on the fault (Figure 9). Slip on the fault patches between the nodes is estimated by bi-linear interpolation.

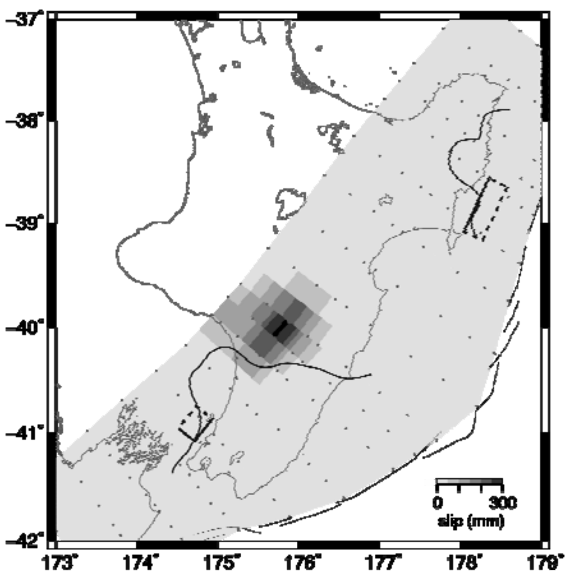

Fig. 9 Slip distribution on the subduction interface over the total duration of the 2004-05 Manawatu slow-slip event. Black lines show the transition from relatively strong to relatively weak coupling in the southern North Island and Gisborne regions, from Figure 3. Black rectangles (solid where well constrained, dashed otherwise) show inferred locations of slow-slip in the Gisborne and Kapiti events. All events lie near or below the coupling transition.

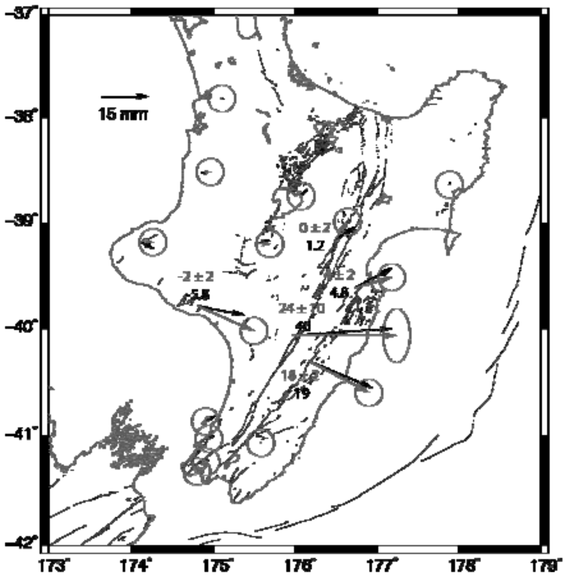

Fig. 10 Fit between observed (grey arrows) and calculated (black arrows) horizontal offsets over the whole duration of the Manawatu event. The numbers beside each site show the observed (grey) and calculated (black) vertical offsets in mm.

The data are best fit by slip of up to 300 mm on the subduction interface, over an area of approximately 100 km by 100 km (Figure 9). The distribution of slip in the event follows the transition zone between the interseismically coupled and creeping portions of the interface estimated from campaign GPS (Figure 3). During early 2004 to January 2005, most of the slip occurred further down-dip on the subduction interface, relative to the later stages of the event. From January-June 2005, the event propagated up dip and then slightly southwards along strike. We show only the total slip distribution from all three sub-events in Figure 9, with the match between observed and predicted displacements in Figure 10. If this aseismic slip had occurred rapidly in an earthquake, it would have resulted in a M_W ~7.0 event.

If the estimated displacement of up to 300 mm is correct, the expected recurrence interval for an aseismic slip event of this type is ~7.5 years, given that the relative plate motion accommodated on this part of the interface is 40 mm/yr (Wallace et al. (2004)). However this calculation assumes there is 100% coupling on the interface in the periods between aseismic slip events, and that these events release 100% of plate motion accumulated between the events. If less than 100% of the plate boundary strain is accumulated between events (and 100% of accumulated strain is released during events), then the actual recurrence interval could be much longer than 7.5 years. Alternatively, if the data can be matched equally well by lower overall slip distributed over a larger portion of the plate interface, the recurrence interval may be shorter.

8 Discussion and Conclusions

All the slow-slip events observed thus far in the North Island occur on the portion of the subduction interface that is in transition between aseismic creep and interseismic coupling. This is shown in Figure 9, where we plot the estimated slip regions for the Gisborne and Kapiti events in addition to the slip distribution of the Manawatu event. This is consistent with observations of slow-slip events in Japan and Cascadia (e.g., Dragert et al. (2001); Obara et al. (2004)), which are also inferred to occur in this transition zone.

The Manawatu and Kapiti slow-slip events occurred during a time period of frequent small to moderate earthquakes in the lower North Island. It is possible that this seismicity was somehow triggered by the slow-slip events, as argued above for the Kapiti case. Robinson (1987) suggested that a change in the character and rate of earthquake activity documented in the Wellington region in mid 1981 was due to a possible episode of slow-slip on the subduction interface. Moreover, Robinson (2003) suggested that the Weber earthquake sequence in 1990 (close to the location of the Manawatu slow-slip event) was triggered by a possible slow-slip event in that region. We may speculate that the 2004-05 Manawatu slow-slip event is a repeat of the slow-slip event that may have triggered the Weber earthquakes. Such speculation implies that large slow-slip events in the southern North Island may occur roughly every 10-15 years.

Slow-slip events at the Hikurangi margin are highly variable in duration, frequency of occurrence, and magnitude of slip. Events near Gisborne occur in as little as 7-10 days, and appear to be the result of at least 200 mm of slip on the interface, recurring every 2-3 years. We have observed smaller magnitude events near Hastings on a regular basis, at least once yearly. These events last for up to a month. The Kapiti and Manawatu events are the slowest, lasting for more than a year, and have occurred near the border of the largest well-coupled portion of the interface. These events also occur at deeper depths (20-40 km) compared to the Gisborne and Hastings events. The Kapiti and Manawatu events are also likely to take longer to recur in comparison to Gisborne and Hastings events, which seem to recur every 1 to 3 years. There seems to be an indication in our data that shallower events occur faster than deeper ones. We speculate that this could represent part of a steady progression from stick-slip behaviour (earthquakes) at shallow depths through to genuinely steady deformation at depths below the deepest slow-slip events.

A rich variety of aseismic deformation events has been observed during just the first few years of GeoNet continuous GPS network development. This emphasises the promise of GeoNet for contributing to our understanding of plate boundary kinematics and dynamics in New Zealand. Improved monitoring, detection and analysis of future aseismic events is important for assessing the seismic hazard posed by the subduction zone, as well as providing greater knowledge and understanding of the fundamental physical processes that control the occurrence of slip (aseismic and seismic) on subduction interfaces.

Acknowledgements

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APPENDIX B PREPRINT

This appendix contains a preprint of Martin Reyners' manuscript "What do small earthquakes tell us about the nature of interplate coupling in the southern North Island, New Zealand?" It expands on the summary given in Section 5 of this report.

What do small earthquakes tell us about the nature of interplate coupling in the southern North Island, New Zealand?

Martin Reyners

GNS Science, PO Box 30368, Lower Hutt, New Zealand

Abstract

To better understand the nature of interplate coupling in the southern North Island of New Zealand, we compare the distribution of relocated small earthquakes with interseismic geodetic locking of the plate interface from GPS and structure within the overlying plate from tomographic inversions. Relocated seismicity shows a good correlation with both the distribution of strong geodetic locking and geological terranes in the overlying plate. It appears that the strong Permian/Triassic Rakaia geological terrane controls frictional coupling at the plate interface. It also modulates the small earthquake distribution within the underlying slab, promoting clustered seismicity in the slab crust and suppressing lower plane seismicity in the mantle. The seismicity and structural data suggest a model where plate coupling is controlled by the ability of fluid to cross the plate interface. When an impermeable terrane in the overlying plate prevents such fluid flow, plate coupling appears to be strong.

1. Introduction

Beneath the North Island of New Zealand, the Pacific plate subducts westward at about 38 mm/yr at the Hikurangi subduction zone (Fig. 1). Campaign and continuous GPS site velocities determined with data since 1991 reveal that portions of the plate interface are locked during the interseismic period [*Wallace et al.*, 2004]. However, there is a wide variation in the down-dip limit of interseismic geodetic locking, with locking extending much deeper in the southern North Island than further northeast along the strike of the subduction zone (Fig. 1). As this lower limit of locking mostly lies beneath the land area, we have good control on its location. Also, the distribution of locking obtained from GPS is consistent with the location of slow-slip events, which to date have all occurred near the down-dip edge of the locked zone [*Beavan et al.*, 2007].

The geodetic transition zone at the down-dip edge of the locked part of the plate interface cuts across isobaths (Fig. 1), and thermal modelling has established that temperature alone cannot explain the distribution of locking [*McCaffrey et al.*, 2007]. So what causes such large changes in the depth of interseismic geodetic locking in the southern North Island? The mechanical behaviour of the subduction thrust will depend on both the large-scale three-dimensional (3-D) distribution of rock types surrounding the thrust, and relatively small-scale structural complexity within the thrust zone itself. The shallow part of the subduction thrust underlies the southern North Island, and is thus amenable to detailed study with land-based seismographs. In recent years we have used dense seismograph deployments to derive detailed 3-D tomographic images of seismic velocities (*Vp* and *Vp/Vs*) and attenuation (Qp^{-1}) surrounding the subduction thrust [*Eberhart-Phillips et al.*, 2005]. One suggestion coming out of this work is that competent terranes in the overlying plate may act as aquicludes, increasing fluid content in the rocks below, and hence might control the distribution of coupling at the plate interface [*Eberhart-Phillips et al.*, 2007]. Here we test this idea by comparing the distribution of abundant small earthquake activity with the distribution of interseismic geodetic locking and structure within the overlying plate.

2. Relocated earthquakes

We relocate all earthquakes in the southern North Island recorded by the GeoNet seismograph network during the period 1990 – 2005 (i.e. a period similar to that for which we have GPS data). We relocate the earthquakes using the 3-D seismic velocity model of *Eberhart-Phillips et al.* [2005], which results in a significant sharpening up of the seismicity distribution compared with the previous routine earthquake hypocentres, which were based on an average 1-D seismic velocity model. As accurate earthquake depths are important in determining in which plate the events are occurring, we only retain events for which the distance to the nearest seismograph is less than twice the earthquake depth. This means that we exclude some shallow events in offshore areas.

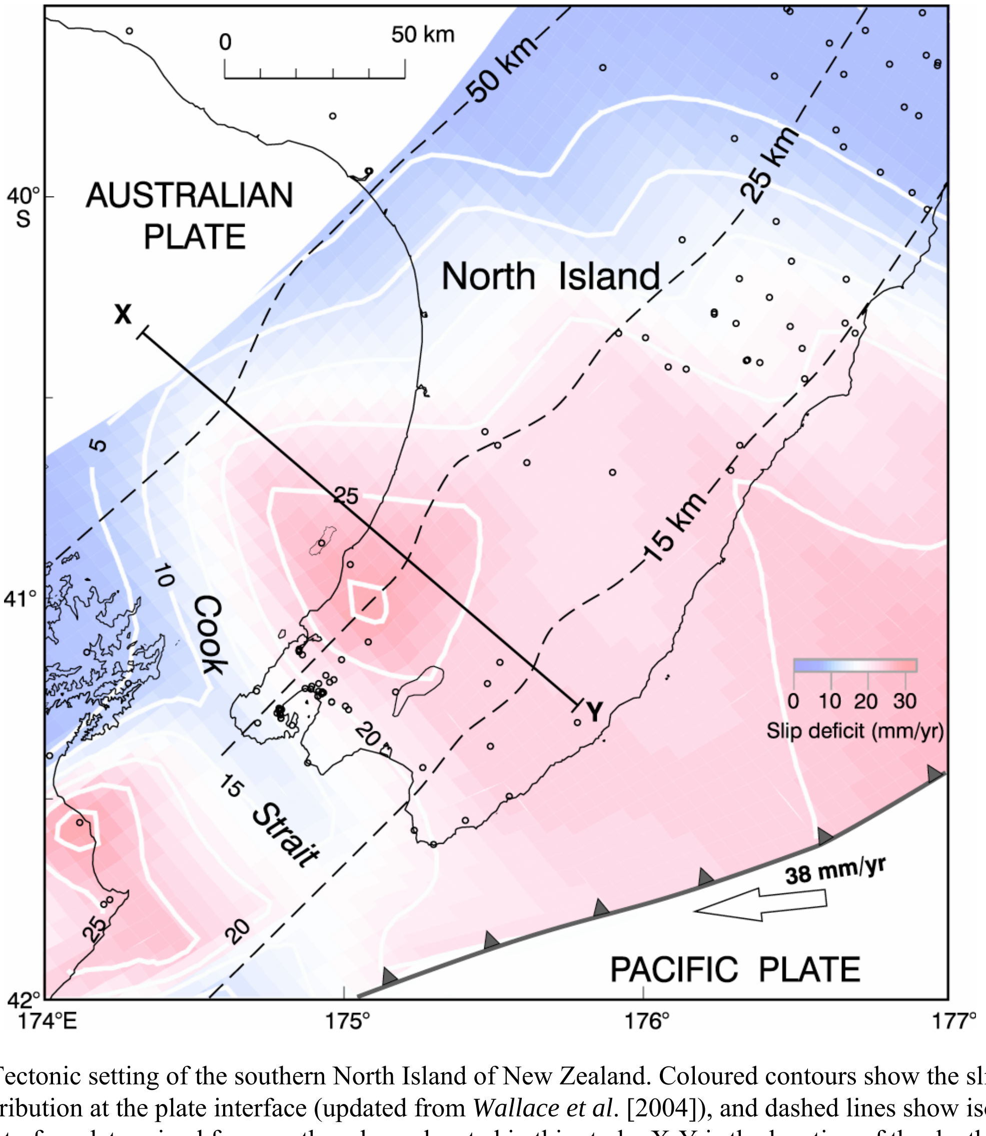

Figure 1. Tectonic setting of the southern North Island of New Zealand. Coloured contours show the slip rate deficit distribution at the plate interface (updated from *Wallace et al.* [2004]), and dashed lines show isobaths of the plate interface determined from earthquakes relocated in this study. X-Y is the location of the depth section shown in Fig. 3.

We next partition the relocated earthquakes into those in the crust and mantle of the overlying plate, those in the upper plane of the dipping seismic zone within the slab, and those within the lower plane of the dipping seismic zone (Fig. 2). We do this by plotting 20 km-wide depth sections of the events oriented down the dip of the subducted plate. We define the plate interface by the upper envelope of seismicity in the upper plane of the dipping seismic zone. This choice of plate interface is consistent with independent data on the location of the interface, including offshore seismic reflection results [*Davey*, 1987; *Davey et al*., 1986; *Stern and Davey*, 1989] and onshore seismic reflection and refraction results [*Chadwick*, 1997; *Davey and Smith*, 1983]. We use a depth of 15 km below the plate interface to separate upper plane and lower plane events. A 15 km thickness for the upper plane contains all events making up clusters defining steeply dipping faults within the top of the slab (Fig. 3). It is also consistent with the crustal thickness of the Hikurangi Plateau capping the Pacific plate east of the trench determined from gravity modelling [*Davy and Wood*, 1994].

2.1 Earthquakes in the overlying plate

As the northwestern part of our study region includes the shallow part of the mantle nose in the overlying plate, we have partitioned overlying plate events into those within the crust (Fig. 2a) and mantle wedge (Fig. 2b), using a Moho depth of 40 km [*Stern and Davey*, 1989]. The distribution of crustal events is rather heterogeneous. The largest cluster defines the aftershock zone of an M_W 6.4 earthquake in May 1990. This earthquake occurred directly above an M_W 6.2 earthquake in the crust of the subducted plate three months earlier (see Fig. 2c for its aftershock zone). There was no seismicity at the plate interface between these earthquake sequences, and *Robinson* [2003] has suggested that the entire earthquake sequence was triggered by a slow-slip event further downdip on the plate interface. This explanation is consistent with the earthquake sequence having occurred on the edge of the region of strong plate coupling.

Interseismic geodetic locking at the plate interface extends deepest where the Permian/Triassic Rakaia terrane overlies the slab (see *Mortimer* [2004] for a description of New Zealand geological terranes). To delimit geological terranes in the southern North Island at the depth of the crustal seismicity, we can use the ambient noise Rayleigh wave tomography results of *Lin et al*. [2007]. At 13 s period, this technique samples the mid-crust (~10-15 km). The Permian/Triassic Rakaia terrane and the Jurassic Kaweka terrane are distinguished by high group velocity (> 2.9 km/s). The Rakaia terrane also has high *Vp* (> 6.5 km/s) and high *Qp* (> 500) at 15 km depth in the region where locking is deepest (Fig. 2a, Fig. 3). This region also has low *Vp/Vs* (< 1.68) at 15 km depth [*Eberhart-Phillips et al*., 2005]. Thus the Rakaia terrane in this region can be interpreted as a competent, cold and dry terrane impacting the subducted plate.

There seems to be relatively little crustal seismicity associated with the Rakaia terrane (Fig. 2a, Fig. 3). Rather, crustal seismicity concentrates to the southeast, in the Cretaceous Pahau terrane. This terrane has a lower Rayleigh wave group velocity (~ 2.7 km/s), a lower *Vp* (~ 6.0 km/s), lower *Qp* (< 250) and higher *Vp/Vs* (> 1.78) at 15 km depth than the Rakaia terrane [*Eberhart-Phillips et al*., 2005; *Lin et al*., 2007; Fig. 3]. Thus the seismic Pahau terrane can be interpreted as less competent and more fluid-rich.

Seismicity in the mantle nose (Fig. 2b) concentrates directly downdip of the region of deepest locking. As this activity lies offshore, we must be certain that it is simply not mislocated slab activity. We have tested this possibility by comparing the seismicity distribution over two periods: 1990-1994, when the seismograph network was sparse, and 2001-2005, when the network was much denser. The seismicity distribution is similar for both periods. Given the depth of these events (> 40 km), the station geometry is still good even though they are offshore, and formal standard errors in the locations of these events are similar to those of nearby slab events.

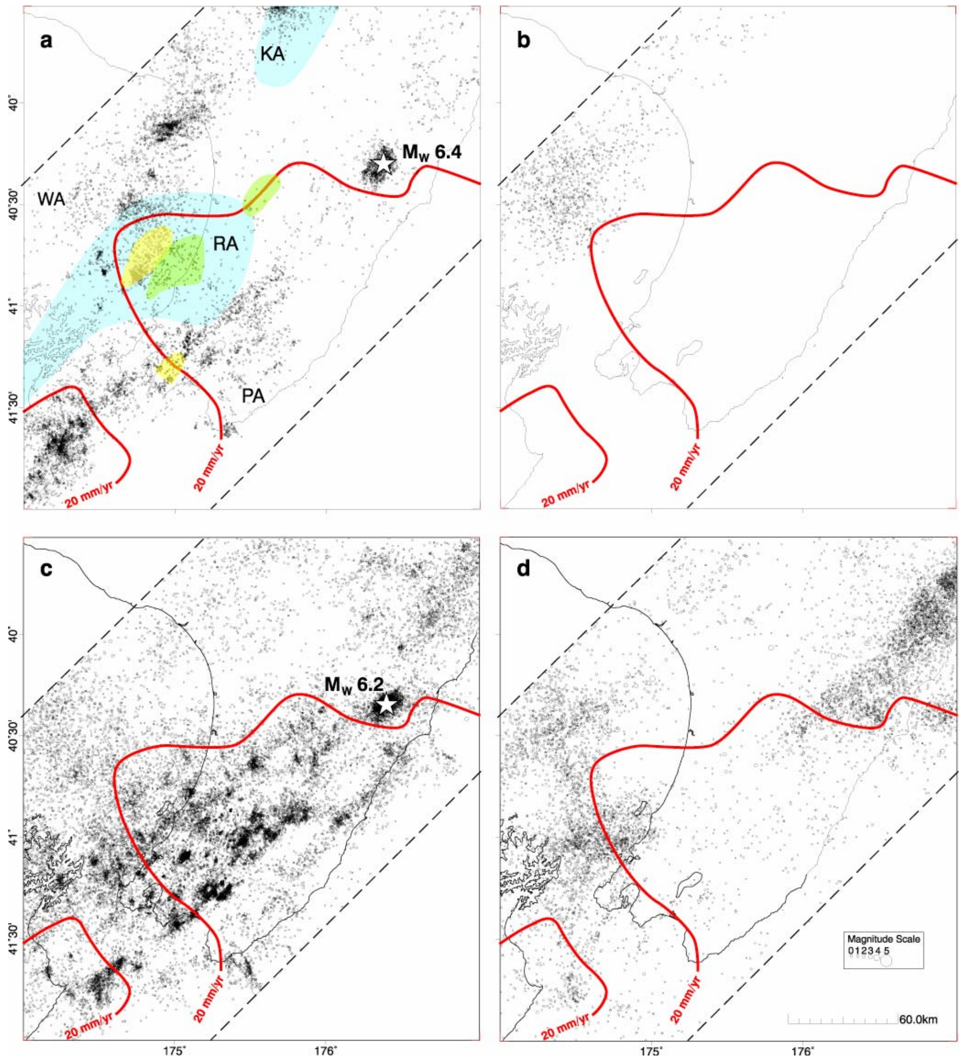

Figure 2. Relocated seismicity for the period 1990-2005. In all plots, dashed lines show the limit of the earthquake sample and red lines are the 20 mm/yr slip rate deficit contours on the plate interface determined from GPS (see Fig. 1). Stars show the two largest earthquake sequences during the study period. **a**) Earthquakes in the overlying plate crust (depth < 40 km). Legend for geological terranes: KA – Kaweka; PA – Pahau; RA – Rakaia; WA – Waipapa. Coloured regions show seismological evidence for strong terranes in the mid crust: blue – group velocity > 2.9 km/s for 13s Rayleigh waves (sampling range ~10-15 km; *Lin et al*. [2007]; yellow – *Vp* > 6.5 km/s and green – *Qp* > 500 (both at 15 km depth; *Eberhart-Phillips et al*. [2005]). **b**) Earthquakes in the mantle nose of the overlying plate (depth ≥ 40 km). **c**) Earthquakes in the upper plane of the dipping seismic zone (i.e. within 15 km of the plate interface). **d**) Earthquakes in the lower plane of the dipping seismic zone.

2.2 Earthquakes in the upper plane of the dipping seismic zone

The distribution of earthquakes in the upper plane of the dipping seismic zone (i.e. less than 15 km below the plate interface) is again heterogeneous (Fig. 2c). The largest cluster denotes aftershocks of the February 1990 M_W 6.2 earthquake, the largest event in the upper plane during the period. Apart from this, earthquake clustering in the southern North Island appears to be most pronounced in the downdip part of the strongly coupled region. There is a prominent southeastern limit to this clustered seismicity, which corresponds to the northwestern limit of significant seismicity in the overlying plate (Fig. 2a). Also, the small earthquake clustering appears to terminate along a NW-SE line through Cook Strait, ~ 20 km southwest of the 20 mm/yr slip deficit contour on the plate interface determined from GPS.

2.3 Earthquakes in the lower plane of the dipping seismic zone

There is very little lower plane seismicity beneath the region of strong geodetic locking in the southern North Island (Fig. 2d). Abundant lower plane seismicity along the strike of the subduction zone in the northeast terminates abruptly at the edge of the locked zone, and significant activity only reappears again at the southwestern, downdip edge of the locked zone. *Qp* results suggest that the dehydration embrittlement model for the generation of seismicity in the subducted slab [*Kirby et al.*, 1996] provides a good explanation for the abundant earthquake activity in the northeast. Furthermore, low *Qp* in the forearc in this region suggests that fluid released from the uppermost mantle passes across the plate interface and weakens the forearc [*Eberhart-Phillips et al.*, 2007]. This implies that for some reason dehydration embrittlement within the uppermost mantle has been suppressed beneath the region of strong geodetic locking.

3. Discussion

The relocated seismicity shows a good correlation with the distribution of strong geodetic locking and terranes in the overlying plate. The concentration of clustered seismicity in the crust of the slab beneath the Rakaia terrane further suggests that this is a region of strong frictional coupling which controls plate locking within the southern North Island. But how does this terrane modulate seismicity in the slab? *Hasegawa et al.* [2007] suggest that the anomalous deepening of a belt of intraslab earthquakes (~100-140 km deep) in the Pacific plate under Kanto, central Japan, is caused by contact with the overlying Philippine slab. Contact with this cold slab hinders the heating of the Pacific slab crust by the hot mantle wedge, which retards dehydration reactions and the resulting seismicity. While the Rakaia terrane is the oldest terrane in the southern North Island and can be expected to be the coldest, the occurrence of abundant, clustered seismicity below it suggests that temperature is not the controlling influence on the seismicity distribution. Rather, the seismicity distribution is more indicative of the high-*Qp* Rakaia terrane being impervious, and trapping fluid in the top of the slab below it. The clustered seismicity in the top of the slab defines high-angle faults oriented along the strike of the subduction zone, consistent with high pore pressures on pre-existing faults (Fig. 2c; *Du et al*., [2004]). Swarm activity on one of these faults has been interpreted as incremental slip on adjacent fault patches [*Reyners and Bannister*, 2007], consistent with fluid movement up the fault. Also, the lack of seismicity within the Rakaia terrane directly overlying the clustered seismicity (Fig. 2a, 2c, 3) suggests fluid does not cross the plate interface in this region.

In contrast, downdip of the Rakaia terrane (i.e. beneath the Jurassic Waipapa terrane), the clustered seismicity in the crust of the slab ceases, lower plane seismicity increases significantly, and seismicity extends across the plate interface into the mantle nose and crust of the overlying plate (Fig. 2, 3), suggesting fluid movement across the plate interface. The concentration of both lower plane and mantle nose seismicity directly downdip of the Rakaia terrane (Fig. 2b, 2d) further suggests that the terrane has had the effect of delaying dehydration reactions within the mantle of the slab. Indeed, the distribution in map view of lower plane seismicity (Fig. 2d) is consistent with the suggestion that dehydration in the mantle of the slab is delayed beneath the region of strongest locking. How locking at the plate interface could affect dehydration reactions in the slab mantle more than 15 km deeper remains an open question. Updip of the Rakaia terrane, seismicity extends across the plate interface and into the crust of the overlying plate, again suggesting fluid movement across the plate interface (Fig. 2a, 3). As mentioned earlier, tomographic results suggest that the Pahau terrane in this region is less competent and more fluid rich than the Rakaia terrane.

Our conceptual model of how the Rakaia terrane might control fluid movement and locking at the plate interface in the southern North Island is shown in Fig. 4. Plate locking in a region of concentrated fluid (and thus presumably high pore pressure) in the top of the subducted plate below the Rakaia terrane at first seems counterintuitive. Normally, high pore pressures in the subducted crust have been interpreted as generating slow-slip at the plate interface (e.g. *Kodaira et al*., [2004]). However, recent numerical simulations using rate- and state-dependent friction laws indicate that high pore pressure at the plate interface will produce stick-slip behaviour (i.e. interseismic locking), while a release of this overpressure will result in slow-slip [*Mitsui and Hirahara*, 2007].

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Our model is put forward as a basis for further work. In determining the general applicability of our interpretation that strong terranes in the overlying plate may control frictional coupling to other subduction zones, it needs to be remembered that the ca. 120 Myr old Pacific slab being subducted beneath the southern North Island is relatively cold, as suggested by its high *Qp* of 900-1200 [*Eberhart-Phillips et al*., 2007]. This is why we see seismicity in the shallow part of the mantle nose and the overlying lower crust (e.g. Fig. 3) – this region is cold enough to support brittle failure. In young, warm subduction zones, this is an aseismic region where episodic tremor occurs (e.g. *Kao et al*. [2006]; *Obara* [2002]). No episodic tremor has yet been identified in the Hikurangi subduction zone [*Delahaye et al*., 2007], suggesting hydroseismogenic processes there differ from those in a warm subduction zone. Nevertheless, even in the warm Cascadia subduction zone regional variations in the character of episodic tremor and slip appear to be spatially correlated with geological terranes in the overlying plate [*Brudzinski and Allen*, 2007].

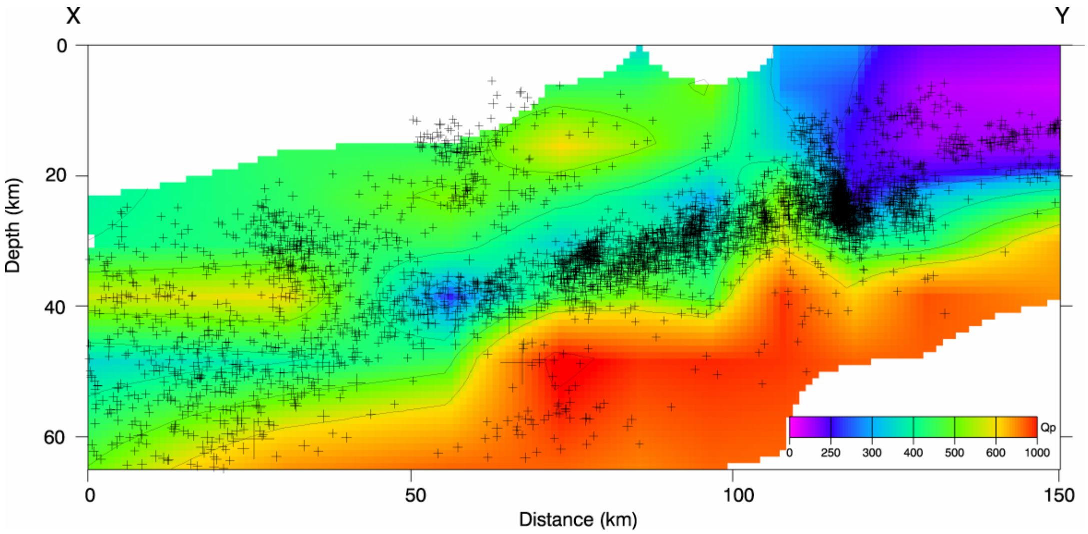

Figure 3. Down-dip depth section through the region of strong plate coupling (see Fig. 1 for location). Relocated earthquakes within 10 km of the section are shown, together with the distribution of *Qp* [*Eberhart-Phillips et al.*, 2005].

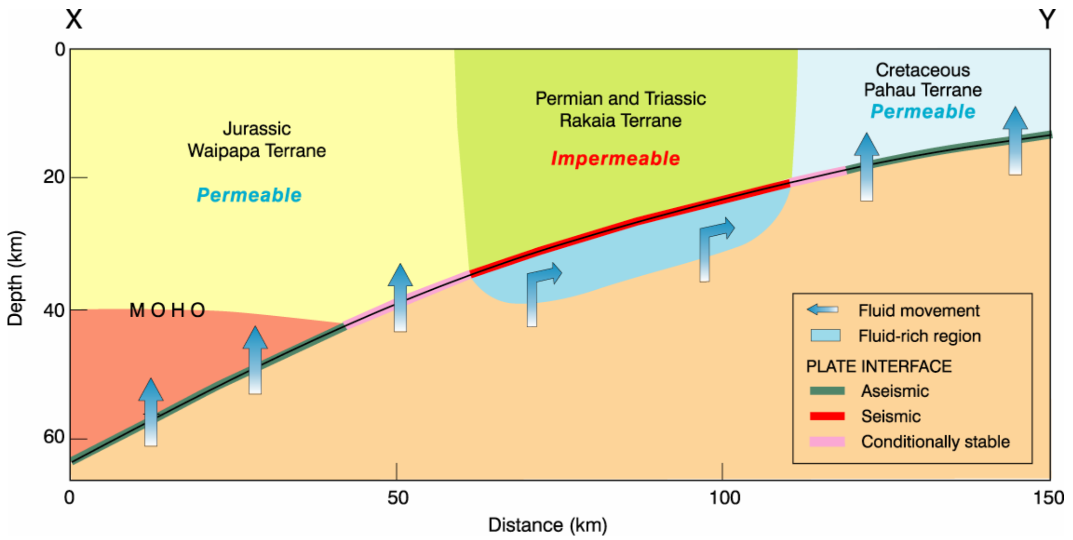

Figure 4. A conceptual model of fluid movement and the nature of plate coupling beneath the southern North Island, along the depth section shown in Fig. 3.

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www.gns.cri.nz

Principal Location

1 Fairway Drive
Avalon
PO Box 30368
Lower Hutt
New Zealand
T +64-4-570 1444
F +64-4-570 4600

Other Locations

Dunedin Research Centre
764 Cumberland Street
Private Bag 1930
Dunedin
New Zealand
T +64-3-477 4050
F +64-3-477 5232

Wairakei Research Centre
114 Karetoto Road
Wairakei
Private Bag 2000, Taupo
New Zealand
T +64-7-374 8211
F +64-7-374 8199

National Isotope Centre
30 Gracefield Road
PO Box 31312
Lower Hutt
New Zealand
T +64-4-570 1444
F +64-4-570 4657