Crustal seismicity in Taranaki, New Zealand using accurate hypocentres from a dense network

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SUMMARY
A large, dense network of three-component, broad-band seismographs was used to determine accurate hypocentres for earthquakes in Taranaki, New Zealand. They allow us to characterize seismicity around Mt Taranaki, a large, dormant, andesite, cone volcano, and to map precisely two major lineations of crustal seismicity. A minimum 1-D velocity model was used to locate 389 local earthquakes using the probabilistic, non-linear earthquake location program NONLINLOC. There are few earthquakes beneath Mt Taranaki itself, and all are relatively small and shallow (≤10 km deep). The shallow seismogenic zone can be explained by the crust being unusually hot, thus causing the base of the brittle–ductile transition to be shallower than normal beneath Mt Taranaki. This is supported by a high heat flow anomaly in this area. The absence of any volcanic earthquakes beneath Mt Taranaki suggests that active volcanic processes are currently unlikely, and the shallow brittle–ductile transition depth means that precursory volcano–tectonic seismicity from any future magmatic intrusion is unlikely to occur below 10-km depth. The permanent seismic network can locate earthquakes in Taranaki reasonably accurately and can reproduce most of the details seen by the temporary seismograph deployment provided that only the best hypocentres are considered. However, beneath Mt Taranaki, which is the most important area for volcano monitoring, hypocentres determined by the permanent network are too deep by 4–12 km. The active Cape Egmont fault zone (CEFZ), west of Mt Taranaki, is the most seismically active area, with earthquakes in the upper crust to about 22-km depth. Spatial and temporal clustering, earthquakes with similar waveforms, and an absence of obvious main shocks imply that earthquake swarms make up a significant proportion of the seismicity in this area. Earthquakes in eastern Taranaki occur primarily along the Taranaki–Ruapehu Line (TRL), thought to be a major boundary across which the crustal thickness changes by about 10 km. These earthquakes are less clustered, have a $b$ value typical of tectonic earthquakes, and occur in the lower crust to a depth of 35 km, with the upper crust almost aseismic. The abrupt cessation of seismicity at 35-km depth is consistent with this boundary marking the Moho, with no earthquakes in the mantle. The concentration of earthquakes in the lower crust requires it to be drier and more mafic than the wet, quartzo-feldspathic composition often used to model crustal rheology. There is no change in maximum earthquake depth across the currently accepted location of the TRL, but there is a 10-km decrease in maximum earthquake depth some 25 km to the north of the currently accepted location. This suggests that the true position of the TRL is 25 km north of the hitherto accepted position.

Key words: $b$ value, crustal rheology, earthquake depth, seismicity, Taranaki, volcano monitoring.

1 INTRODUCTION
The Taranaki region (Fig. 1), in the western part of the North Island of New Zealand, is dominated by Mt Taranaki, a 2518-m, andesite, cone volcano (Neall et al. 1986) that lies within a large sedimentary

Taranaki seismicity

Figure 1. Map of the Taranaki region showing topography, the portable seismograph network (triangles), the Taranaki Volcano-Seismic Network (TVSN; open circles), active faults (black lines) and Neogene faults from oil industry seismic reflection profiles (grey lines). The Taranaki basin lies west of the Taranaki fault. The Taranaki–Ruapehu Line (TRL), as defined by Stern et al. (1987), is shown as a dashed line. The Taranaki volcanoes are Mt Taranaki (MT), Pouakai (P), Kaitake (K) and Sugar Loaf—Paritutu (SLP). The named active faults are Oaonui (OF), Inglewood (IF) and Norfolk (NF). The lower inset shows the location of the study area with the tectonic plates and plate boundary, North Island (NI) and South Island (SI) and the Taupo volcanic zone (TVZ). Basement geology, Eastern Province (EP), Median tectonic zone (MTZ) and Western Province (WP), is after Mortimer et al. (1997). The upper inset shows other permanent seismographs and strong motion recorders from which data were used.

basin containing significant oil and gas reserves (King & Thrasher 1996). Taranaki is seismically active and includes two major lineations of crustal seismicity with historic earthquakes as large as $M_{L}6.1$ (Robinson et al. 1976; Anderson & Webb 1994).

Mt Taranaki is currently inactive, but last erupted in AD 1755 and has a mean period between eruptions (with a volume $>10^{7} \text{ m}^{3}$) of 330 yr, though the mean period for smaller eruptions that are not well-preserved geologically could be significantly shorter (Alloway et al. 1995). Mt Taranaki has a history of small–moderate eruptions over the last 120 000 yr that are characterized by dome emplacement and subsequent collapse, by ash eruption, and by frequent pyroclastic flows and lahars (Neall et al. 1986). Future eruptions would have a significant impact on the local economy, particularly dairy farming, and oil and gas production (Hull 1996), yet nothing is known of precursors to past eruptions as New Zealand was only sparsely settled until the mid-1800s. Long-term monitoring uses the Taranaki Volcano-Seismic Network (TVSN), a network of six short-period, vertical-component seismographs, supplemented by more distant seismographs for larger earthquakes, but little is known of the structure of the crust at the depths at which earthquakes and magmatic processes occur. If volcanologists are to successfully interpret eruption precursors, and issue timely and accurate warnings, then a thorough understanding of local seismicity and structure is essential.

The Cape Egmont fault zone (CEFZ) marks the western limit of deformation related to the Pacific–Australian plate boundary in the North Island of New Zealand (Fig. 1). The faulting dates from the mid-Cretaceous formation of the Tasman sea (King & Thrasher 1996) and is still active (Nodder 1993). The Cape Egmont fault, the largest individual fault, is visible as a 1–5 m high scarp for 53 km along the seafloor and has had up to 0.8 mm yr$^{-1}$ of dip-slip movement over the last 225 000 yr (Nodder 1993). There is a 150-km-long
lineation of crustal seismicity associated with the CEFZ (Anderson & Webb 1994), but because earthquakes are mostly offshore, their hypocentres and particularly their depths are inaccurate, except in western Taranaki where the CEFZ and seismicity are onshore and less than 20 km west of Mt Taranaki.

An east–west line through a steep north–south gradient in the iso-static gravity anomaly field in eastern Taranaki marks the Taranaki–Ruapehu Line (TRL, Fig. 1), thought to mark the juxtaposition of a thin crust (25 km) to the north against a normal thickness crust (36 km) to the south (Stern et al. 1987). As a crustal boundary, it is poorly understood and has only recently begun to be studied using seismic attenuation and resistivity techniques (Salmon et al. 2003). There is a broad lineation of seismicity associated with the TRL that is thought to occur mainly in the lower crust (Anderson & Webb 1994), but because of the absence of nearby permanent seismographs many depths are fixed in the location process and how these earthquakes relate to the TRL is unknown.

This paper describes the first results of a multifaceted study of seismicity and structure in Taranaki. The overall aim is to understand the structure and background seismicity of the region and to use this information to allow better interpretation of future eruption precursors at Mt Taranaki. The objective of this paper is to report this information to allow better interpretation of future eruption precursors, to characterize better the style of seismicity in the CEFZ, and to use accurate hypocentres to help understand the nature and position of the crustal changes across the TRL.

1.1 Setting and previous work

Taranaki is about 400 km west of the surface expression of the obliquely convergent boundary between the Pacific and Australian plates, and 150 km west of the Taupo volcanic zone (TVZ), the main region of active volcanism in New Zealand (Fig. 1). Mt Taranaki is the most recent of the Taranaki volcanoes, which represent a south– southeast migration of activity commencing at 1.74 Ma (Neall et al. 1986). Volcanism at Mt Taranaki began at about 120 ka, but it has suffered several collapse events and the present cone is less than 10 000 yr old. A dyke-stock complex of solidified andesite forms the core of all the Taranaki volcanoes and extends to a depth of at least 6 km (Locke & Cassidy 1997). There is a warm (up to 30 °C) spring and a cold, bicarbonate spring containing dissolved carbon dioxide of volcanic origin on Mt Taranaki (Allis et al. 1995), and a small heat flow anomaly centred near the oldest of the Taranaki volcanoes that is thought to be of volcanic origin (Allis et al. 1995; Funnell et al. 1996). However, there is no evidence of ongoing volcanic processes (such as magma movement or degassing) and Mt Taranaki is currently inactive.

The Taranaki basin, a large Cretaceous–Tertiary basin with significant oil and gas reserves, is the main geological feature of western and central Taranaki with an average thickness of 6 km of sediments (King & Thrasher 1996). The eastern boundary of the basin is marked by the Taranaki fault (Fig. 1), a Miocene reverse fault that offsets basement rocks by about 6 km vertically, and the basin extends west to at least the continental shelf break (King & Thrasher 1996). Basement lithology is sampled only from a few deep drill holes and is partly extrapolated from the northern part of the South Island (Mortimer et al. 1997). Basement rocks beneath the Taranaki basin are subduction-related calc-alkaline plutons with subordinate volcanic and sedimentary rocks and form part of the Median tectonic zone. East of the Taranaki basin are Eastern Province rocks, which are lithic and feldspathic meta-greywackes with some volcanic and intrusive rocks. There may also be a 15-km-wide strip of pyroxene-rich lavas and sedimentary rocks of the Brook Street Terrane just west of the Taranaki fault (Mortimer et al. 1997). Seismic reflection and fault-trenching data show that faulting in western and central Taranaki is dominantly normal with faults aligned northeast–southwest (Fig. 1; Nodder 1993; Hull 1994).

Southeast Taranaki lies above the deepest part of the Benioff zone beneath the North Island (Anderson & Webb 1994), with infrequent very deep earthquakes (about 600 km) beneath northeast Taranaki marking a deep, possibly detached, part of the subducted Pacific Plate (Adams & Ferris 1976; Anderson & Webb 1994; Boddington et al. 2004). Seismicity close to Mt Taranaki was first described in detail using data from the TVSN (Cavill et al. 1997), but those results were limited by a small network of only five vertical-component seismometers centred on the volcano and the wider regional context was not discussed. This study has more than 10 times the number of seismographs and the advantage of continuous, three-component data, which allow the hypocentres to be constrained better. Cavill et al. (1997) imposed a 1-D velocity model a priori, whereas we use a self-consistent model derived from the data. We also use a probabilistic, fully non-linear location program that is able to show hypocentral uncertainty more realistically than earlier studies.

2 NETWORK, DATA AND ANALYSIS

A network of 75 three-component, broad-band seismographs, Guralp CMG-6TD and CMG-40T, was operated from 2001 December to 2002 September (Fig. 1), with data recovered from 68 of these (Sherburn & Allen 2002). The seismographs have a near flat response to velocity between 0.03 and 50 Hz and GPS timing. Data were digitised at the field at 100 Hz and downloaded monthly for subsequent processing.

A frequency-domain detector (Gledhill 1985) followed by manual checking was used to find local earthquakes, with S–P times ≤ 10 s at the nearest site, and these were supplemented by data from the TVSN and several other permanent sites in western New Zealand (Fig. 1) operated by the Institute of Geological and Nuclear Sciences (GNS). For phase picking and amplitude measurement, the SEISAN analysis package (Havskov & Ottemöller 1999) was used. P phases were picked only on vertical-component seismograms and S phases only on horizontal-component seismograms, with the first arriving, clearest S phase picked on unrotated seismograms. Picks were weighted (0–4) according to their perceived reliability judged on the basis of the signal-to-noise ratio (S/N) and the estimated uncertainty in pick time. Weights of 0–4 correspond to uncertainties of ≤ 0.05, ≤ 0.10, ≤ 0.25, ≤ 1.00 and > 1.00 s, respectively. P picks were given weights of 0–4 and S picks weights of 1–4. Picks with a weight of 4 were not used in the analysis. Where possible, picks were made on unfiltered seismograms, but where the S/N was too low, seismograms were bandpass filtered at 2–20 Hz to remove strong microseisms before picking. A small phase shift was sometimes introduced by the filtering, but tests showed that this was typically less than the uncertainty assigned to a pick of that weight and, in addition, phases picked on filtered seismograms were given lower weights than equivalent phases picked on unfiltred seismograms. A total of 389 earthquakes were located, using more than 15 000 phase picks, comprising 55 per cent P phases and 45 per cent S phases. The overall data errors, estimated as the weighted mean uncertainty attributed to the phase picks, were 0.10 s for P, 0.20 s for S and 0.15 s for the complete data set.
Amplitudes were measured from synthetic horizontal-component Wood–Anderson seismograms and used to estimate magnitudes. Magnitudes were compared with those determined by GNS ($M_L$, Maunder 2001) for all earthquakes common to both catalogues (150 events, $M_L1.3$–$3.9$) and site-specific magnitude corrections, calculated as the mean difference between GNS magnitudes and those in this study, were calculated for portable sites on a site by site basis (mean correction 0.2, range $-0.2$ to 0.6). After averaging, the final magnitudes ranged from 1.4 to 4.0, with a mean difference from GNS magnitudes of 0.14.

3 EARTHQUAKE LOCATION

3.1 Minimum 1-D velocity model

Preliminary locations were calculated using the standard New Zealand 1-D velocity model (Maunder 2001). A minimum 1-D model, a model that minimizes the overall misfit between the observations and model predictions, was then calculated using the program VELEST (Kissling 1988, 1995). Station corrections were calculated as part of the velocity model and account for 3-D features in the real velocity structure that are not represented by the 1-D approximation. The model was calculated in three steps: first, a $V_p$ model was calculated using only $P$-phase data, with a starting model based on a priori information and apparent velocities present in the data; secondly, a $V_s$ model was constructed using only $S$-phase data, with a starting model based on the $V_p$ model and a mean $V_p/V_s$ ratio for the region; and finally, a joint $V_p$ and $V_s$ model was made using both $P$- and $S$-phase data, and the $V_p$ and $V_s$ models as starting models. For the $V_p$ model, earthquakes were selected on the basis of an azimuth gap of $\leq 180^\circ$ and $\geq 10$ $P$ phases with a weight $\leq 2$. To avoid biasing the 1-D model to overly reflect seismic velocities along ray paths from earthquakes in the most active region west of Mt Taranaki, $\geq 15$ $P$ phases were required for earthquakes from this region to be selected. The $V_s$ model used the same criteria, but considered only $S$ phases. The joint model used the same earthquakes as the $V_p$ model with the $P$ and $S$ phases from those earthquakes. For $V_p$ there were 149 earthquakes with 3907 $P$ phases, for $V_s$ 142 earthquakes with 3142 $S$ phases, and for the joint model 149 earthquakes and a total of 7203 phases (Fig. 2). The coverage by both $P$ and $S$ rays was excel lent, and the earthquakes sampled the whole crustal volume, though not uniformly. The only difference between $P$ and $S$ rays was the lack of $S$ arrivals at sites east of Taranaki, as these seismographs were mostly vertical-component only.

The initial model had 22 layers with a vertical spacing of 1 km in the upper 6 km of the crust, corresponding to the average thickness of Taranaki basin sediments, increasing to 3 km spacing in the lower crust (Fig. 3). Velocity reversals were not permitted. $V_p$ inversions were carried out using a range of starting models with different velocities in order to sample the whole range of the solution space and to ensure that the final model had a global rms minimum rather than just a local minimum (Kissling 1995). Between 10- and 36-km depth, different initial models converged satisfactorily towards a common model, but at shallower depths, the final model depended strongly on the initial model with both low- and high-velocity models achieving the same overall rms misfit. Below 36-km depth, there was little control on velocities as this is deeper than the deepest hypocentres and 90 per cent of rays left the source above the horizontal, so deeper velocities were poorly sampled. In the final $V_p$ model, the velocity of the upper 10 km was taken as the mean of the high and low extremes, and the bottom of the model was set at 38-km depth with a velocity of 7.4 km s$^{-1}$, similar to the highest apparent velocities observed in the data.

A Wadati plot gave a $V_p/V_s$ ratio of 1.72 ± 0.05 for our data, so an initial $V_s$ model was derived from the $V_p$ model using a constant $V_p/V_s$ ratio of 1.72. The $V_s$ inversion used only $S$ phases; no $P$ phases were used to control hypocentres and a single run produced a $V_s$ model only slightly different from the starting model. In western and central Taranaki, the locations of earthquakes using either $V_p$ or $V_s$ models were similar (<2 km difference), but in eastern Taranaki the influence of limited control from the east pushed hypocentres calculated using the $V_s$ model towards the east. For the joint $V_p$ and
S. Sherburn and R. S. White

Figure 3. Jointly calculated minimum 1-D $V_p$ and $V_s$ models (left) and station corrections (right). $V_p/V_s$ is calculated directly from $V_p$ and $V_s$, with the vertical shaded region marking the initial $V_p/V_s$ ratio of 1.72 ± 0.05. The $V_p/V_s$ ratio above 2 at 2–3 km depth is an artefact of the way in which $V_p/V_s$ was calculated and occurs because $V_p$ increases in that layer, but $V_s$ does not. A more realistic value is 1.9, equivalent to the $V_p/V_s$ ratio of the shallower part of the model. A star marks the reference station at which the $P$ station term was fixed at zero. The area of the Taranaki volcanoes above 500-m elevation is shaded and the Taranaki fault is also shown. $V_s$ models, earthquakes were first relocated using the combined $V_p$ and $V_s$ models calculated earlier and a single inversion gave a final model (Fig. 3).

To ensure that the velocity models are robust estimates of the real velocity structure and do not depend on initial parameters, the models were tested by a combination of relocating earthquakes after randomly shifting their hypocentres, recalculating station corrections after an initial assumption of zero corrections, and calculating new velocity models starting with zero station corrections and randomly shifted hypocentres. The velocity models were successfully reproduced to within 0.25 km s$^{-1}$ of their starting values given a range of different initial parameters, and are therefore judged to be robust and reliable.

Station corrections are consistent with the geology and sediment thickness data (King & Thrasher 1996), and are similar for both $V_p$ and $V_s$ models. There is a change to fast corrections east of the Taranaki fault, which forms the eastern boundary of the Taranaki basin, while variations in sediment thickness are evident in the northern and southern parts of the Taranaki basin (Fig. 3). Fast corrections beneath the Taranaki volcanoes, especially for $V_s$, suggest that the solid andesite core modelled by gravity (Locke & Cassidy 1997) is sampled by our earthquake data and has high velocity.

3.2 Taranaki velocity model

As shown in Fig. 4, at shallow depths the minimum 1-D $V_p$ model is faster than that used by Cavill et al. (1997) because the minimum 1-D model averages velocities both inside and outside the Taranaki basin. In contrast, shallow velocities are much lower than both the standard New Zealand model (Maunder 2001) and that for the western North Island (Stern et al. 1987) as these take no account of surficial low-velocity sediments. However, in the depth range 12–24 km, the Taranaki model is similar to both these, from which we infer that the mid-crust is typical of that in New Zealand as a whole. Despite volcanism in Taranaki having occurred since 1.74 Ma, the minimum 1-D model and TVZ models have little in...
common (Fig. 4). This is not surprising, given that the TVZ is a back-arc spreading centre with extensive volcanism and very high heat flow (Bibby et al. 1995), while Taranaki has continental crust, a dominantly sedimentary environment and, by comparison, volumetrically limited volcanism.

Calculating $V_p/V_s$ directly from $V_p$ and $V_s$ may produce significant artefacts due to the differences in the resolution of the two models. Despite this, it is clear that for depths $\leq 3$ km the $V_p/V_s$ ratio is high, up to 1.9 or 2.0, while at greater depths it is generally within the estimated uncertainty of $\pm 0.05$ of the mean value of 1.72 (Fig. 3). The $V_p/V_s$ ratio varies significantly depending on rock properties (Johnston & Christensen 1992; Eberhart-Phillips et al. 1995; Christensen 1996), and in a sedimentary basin can be strongly influenced by high-porosity and fluid saturated sediments that cause low $V_s$, but do not significantly affect $V_p$ (Nur 1972). The high $V_p/V_s$ ratio at shallow depth is therefore ascribed to unconsolidated, fluid saturated sediments, especially those in the Taranaki basin.

### 3.3 Earthquake relocation

All earthquakes were relocated using the minimum 1-D velocity model and the computer program NONLINLOC, a probabilistic, fully non-linear earthquake location program that calculates a maximum likelihood hypocentre by a 3-D search (Lomax 2001; Husen et al. 2003). The hypocentre is expressed as a probability density function (PDF) and includes the effects of location uncertainties due to seismic network geometry, arrival time uncertainty and errors in the calculation of theoretical travel times. Using an oct-tree search algorithm (Lomax 2001) the hypocentral uncertainty is displayed as a scatter diagram with samples drawn from the PDF, with the number of samples proportional to probability. The region of uncertainty may be irregularly shaped and have more than one minimum (e.g. Husen et al. 2003). NONLINLOC also calculates an expectation hypocentre with a 68 per cent confidence ellipsoid that corresponds to the location obtained using a conventional linearized algorithm. In this study, locations are well controlled by sufficient, well-distributed seismographs, and there is little difference between the maximum likelihood and the expectation hypocentres and their respective uncertainty estimates (Fig. 5). Maximum likelihood hypocentres are shown in subsequent figures.

There were no artificial seismic sources of known location to assess independently the absolute hypocentral uncertainty, so differences between hypocentres calculated using different models and randomly shifted hypocentres were used (Haslinger et al. 1999). The estimated uncertainty is $\pm 1$ km in position and $\pm 2$ km in depth. The relative depth uncertainty was usually larger than the epicentral uncertainty, but neither was sufficient to alter the observed seismicity pattern.

![Figure 5. Hypocentral uncertainty for a typical shallow earthquake located in central Taranaki. Map (x–y), x-depth and y-depth views are shown. Grey circles represent 1000 samples drawn from the solution PDF. The star is the maximum likelihood hypocentre, while the open circle and ellipsoid represent the expectation hypocentre and 68 per cent confidence limits, respectively. The lower right plot shows the maximum likelihood hypocentre (star) and seismographs that recorded the earthquake (grey circles). Hypocentral parameters are shown at the top of the figure. The greatest angle without an observation (azimuth gap) is 102°, the distance to the closest seismograph is 8.3 km, and 15 $P$ phases and 10 $S$ phases were used in the location. The maximum likelihood hypocentres are shown in subsequent figures.](https://academic.oup.com/gji/article-abstract/162/2/494/552780)
No attempt was made to relocate permanent network hypocentres using the minimum 1-D velocity model because there were no station corrections for S phases picked on vertical-component stations of the permanent TVSN. S phases picked on these seismographs are routinely used in locating Taranaki earthquakes, but were not used in this study because they were of much lower quality than those from the three-component portable network.

4 RESULTS

4.1 Hypocentre distribution

The hypocentre distribution (Fig. 6) is similar to the long-term seismicity pattern (Anderson & Webb 1994), but with much less scatter. Despite the overall northeast–southwest lineation of the CEFZ seismicity, the onshore portion of this activity occurred mainly in four semi-distinct clusters at 10–20 km depth, with no obvious lineations.

The level of seismicity in central Taranaki, beneath Mt Taranaki and almost as far east as the Taranaki fault, was very low. All earthquakes were shallower than 10 km and relatively small (largest $M_{2.2}$). In all cases, the PDF for these earthquakes had a single minimum and was ellipsoidal in shape, generally being well represented by the 68 per cent confidence ellipsoid (Fig. 5), so the anomalously shallow depths could not be explained by a PDF with both deep and shallow minima, and these earthquakes are well constrained by our observations to be this shallow.

Southern Taranaki (south of $39.4^\circ$S and west of the Taranaki fault) was almost completely aseismic during the observation period. The permanent network detects only a few well-located earthquakes here (Fig. 6), so the level of seismicity here is always very low.

About $174.3^\circ$E, just west of the Taranaki fault, there is an abrupt increase in the level of seismicity and a corresponding increase in hypocentral depth to $25–30$ km (Fig. 6). There is little seismicity 10–20 km east of the Taranaki fault, then there is a further increase in the level of activity and in the depth to about 35 km. These earthquakes occur in a band 40–50 km wide centred on the TRL (Fig. 1). Although there is some clustering, it is noticeably less than in western Taranaki. As a lineation of seismicity, the TRL is
clearly distinct from the CEFZ and does not start at Mt Taranaki as suggested by its name and by some authors (Anderson & Webb 1994), but 30–40 km east of the summit.

In an west–east cross-section, there are three first-order changes in hypocentral depth, an abrupt shallowing beneath the western flank of Mt Taranaki, a deepening near the Taranaki fault and then a change to even deeper earthquakes about 20 km further east (Fig. 7). Although there is some small variation in relative depth uncertainty among these earthquakes, the uncertainties alone are insufficient to explain such large depth differences.

A north–south cross-section across eastern Taranaki (Fig. 8) shows that earthquakes south of the position of the TRL of Stern et al. (1987) occur to almost the estimated crustal thickness of 36 km (Stern et al. 1987). Earthquakes exhibit the same maximum depth for 10 km north of that boundary and are still 30 km deep some 25 km to the north, where the crustal thickness has previously been estimated as 25 km (Stern et al. 1987). A small group of earthquakes, about 50 km north of the TRL, is 15–25 km deep and is separated by 20 km from other relocated seismicity in eastern Taranaki.

4.2 Frequency–magnitude distribution and seismicity style

The size distribution of a group of earthquakes can be described by a power law known as the frequency–magnitude distribution (FMD) (Gutenberg & Richter 1954):

\[ \log_{10} N = a - bM, \]

where \( M \) is magnitude, \( N \) is the cumulative number of earthquakes larger than \( M \), and \( a \) and \( b \) are constants. The slope of the relationship is known as the \( b \) value and measures the relative numbers of small and large earthquakes. For tectonic earthquake sequences, \( b \) typically has a value of about 1. The FMD was calculated for Taranaki earthquakes using the ZMAP software (Wiener 2001) with \( b \) calculated using the maximum likelihood method (Aki 1965), excluding earthquakes more than 20 km offshore to avoid biasing the results (Wiener & Wyss 2002). For all earthquakes and for the TRL separately, \( b \) was 1.12 and 1.11, respectively, but for the CEFZ it was 1.34, significantly different at the 95 per cent confidence level (Table 1). The CEFZ seismicity occurred in four semi-distinct clusters (Fig. 9), which were analysed separately, giving 1.11 for clusters B and D together and 1.72 for the remaining earthquakes, significantly different at the 97 per cent confidence level.

For earthquakes in the GNS catalogue for the period 1994–2001, \( b \) was 1.17 (Table 1), not significantly different from the value calculated for all earthquakes in this study, even at the 90 per cent confidence level. The FMDs for the whole region and for the TRL are therefore similar to the long-term value, and both are typical of tectonic earthquakes. However, there is significant spatial variability in the FMD in the CEFZ, with areas in which \( b \) is similar to the whole region and areas where it is significantly higher.

The rate of activity in the CEFZ was also variable, with periods of almost constant rate and short periods with a very high rate of
Table 1. Frequency–magnitude distribution (FMD) for various Taranaki data sets. \( a \) and \( b \) are defined in eq. (1), \( \delta b \) is the standard deviation of \( b \), and \( M_C \) is the magnitude of completeness. ‘GNS’ data cover the same area as ‘all’, but are for the period 1994–2001.

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Figure 9. Earthquake clusters west of Mt Taranaki for 2002 March 1—September 10. Four circles (A–D) are drawn around obvious clusters of epicentres. Symbols are scaled according to magnitude. The coast, the area of the Taranaki volcanoes above 500-m elevation (shaded) and the surface trace of the Oaonui fault (OF) are shown.

5 DISCUSSION

Seismicity in the Taranaki region is concentrated in two main lineations, the CEFZ and the TRL, with relatively few earthquakes outside those areas. The Inglewood and Norfolk faults, last active 3.3–3.5 ka producing a \( M_{w}6.6–7.1 \) earthquake (Hull 1994), are seismically active, the first time earthquakes associated with these faults have been detected with confidence (Cavill et al. 1997). There is a short lineation in hypocentres trending northeast beneath Mt Taranaki that appears to link the Oaonui fault with the Inglewood and Norfolk faults, tentatively supporting an interpretation that there might be a fault, or fault zone, beneath Mt Taranaki (Cavill et al. 1997).

5.1 Earthquake depths

In continental crust, away from subduction zones, earthquakes are often concentrated in the upper crust and the lower crust is aseis-
the brittle–ductile transition depth and the uncertainty that so many variables introduce, the difference in maximum earthquake depths between western and central Taranaki might be explicable by a combination of differences in strain rate and temperature gradient without requiring a change in lithology, but the larger depth differences between central and eastern Taranaki cannot.

The work of Chen & Molnar (1983) suggested that earthquakes occur in the continental mantle, but improved crustal thickness estimates and focal depths have led some authors to suggest that these earthquakes are actually in the lower crust (Maggi et al. 2000; Jackson 2002; Roecker et al. 2003), while others contend that recent data confirms that the mantle is seismically active (Chen & Yang 2003). In New Zealand, Reyners (1987) has shown that earthquakes occur beneath the central South Island to a depth of more than 70 km and that gravity modelling suggests these are in the uppermost mantle. Whether the continental mantle is seismically active remains a hotly debated topic. Fortunately, the maximum earthquake depth in eastern Taranaki is about 35 km, almost the same as the estimate of Stern et al. (1987) of crustal thickness in this area and similar to that estimated 200 km farther south (Garrick 1968), so we can safely assume that these earthquakes occur in the lowermost crust.

A consequence of these earthquakes being in the crust is that, in contrast to what is normally observed, the upper crust in eastern Taranaki is almost aseismic, with activity concentrated in the lower crust. There are no heat flow measurements in eastern Taranaki, but based on those nearby, the heat flow is probably about 60 mW m$^{-2}$ (Pandey 1981; Funnell et al. 1996). A heat flow of even 50 mW m$^{-2}$ suggests a temperature of more than 500 $^\circ$C at 35-km depth so that rocks would be ductile and the region would be aseismic. The most likely explanation for the deep earthquakes is therefore that the crust is drier and more mafic than the wet, quartzo-feldspathic composition often used to model crustal rheology, as this gives high frictional strength at depths where rocks would normally deform by plastic flow. Jackson et al. (2004) use the same argument to explain earthquakes in the lower crust of the Indian shield at depths of 70–80 km beneath southern Tibet and conclude that anhydrous granulite, a fine-grained, basic metamorphic rock, forms the lower crust. A change in rock type as the main factor causing the large change in earthquake depths across the Taranaki fault is supported by a change in basement lithology across the fault (Mortimer et al. 1997), and the difficulty in explaining the change by a combination of non-uniform strain rate and a change in the temperature gradient.

5.2 Taranaki–Ruapehu Line
The 40–50 km wide band of seismicity in eastern Taranaki suggests that deformation related to the TRL is distributed over a wide area. Stern et al. (1987) estimated the position of the TRL and the crustal thickness on either side of the boundary by simple modelling of the North Island isostatic gravity field constrained by seismic estimates of crustal thickness; the position of the TRL being controlled by the location of the zero isostatic anomaly contour. If we accept the TRL position of Stern et al. (1987) and their estimate of the crustal thickness on either side of the boundary, then the occurrence of earthquakes 30–35 km deep as far as 25 km north of the boundary would mean that those earthquakes were in the uppermost mantle, yet the earthquakes to the south of the boundary would be in the lowermost crust (Fig. 8). A simpler explanation might be that all the earthquakes are in the lowermost crust and that the change in crustal thickness across the TRL occurs about 25 km north of the estimate of Stern et al. (1987). This argument is supported by Salmon et al. (2003) who made magnetotelluric and attenuation measurements.
along a profile across the eastern part of the TRL. They found low resistivity (<300 Ωm) deeper than 25-km depth to the north of the boundary but an absence of such low resistivities to the south. They also report high attenuation (path averaged quality factor ≈200) to the north and low attenuation (quality factor >1500) to the south. Both boundaries are about 20 km north of the location of the TRL published by Stern et al. (1987), though magnetotelluric measurements and seismograph station spacings were ca 20 km for these profiles so the uncertainty in the position of the resistivity and attenuation boundaries is about ±10 km.

5.3 Cape Egmont fault zone

Significant spatial variations in b are observed in the CEFZ, with some areas of very high b and other areas in which b is similar to that found in other parts of Taranaki. Although spatial variations in b have been observed in many areas, the physical basis for such variations is often poorly understood (Wiemer & Wyss 2002). Those authors suggest that high b values might occur in conditions as diverse as low ambient stress, a strongly cracked environment and high pore-pressure conditions. Although many cracks are obviously likely within an active fault zone, and high fluid pressures are possible if the permeability is low and fluids are trapped by fault gouge, the reason for the small areas of high b in the CEFZ is not obvious.

5.4 Long-term seismicity

The hypocentre distribution recorded by the temporary network is broadly similar to the long-term seismicity pattern (Figs 6, 7 and 8), with most differences reflecting greater uncertainty in hypocentres derived from the sparse permanent seismograph network. A one-to-one comparison of earthquakes recorded by both networks shows that 80 per cent of GNS earthquakes are within 10 km of the temporary network epicentres from this study and within 8 km of the temporary network depths. The epicentral difference for a few earthquakes is very large (>40 km) and is typically caused by a poorly constrained GNS location, usually due to insufficient phase picks.

The most obvious difference between the temporary and permanent network hypocentres is that the permanent hypocentres in central Taranaki are deeper than those derived from the temporary network (Fig. 7). This is probably because the permanent network is too sparse to locate such shallow earthquakes accurately and the lack of low, near-surface velocities in the standard New Zealand velocity model combined with late S picks from the vertical-component seismographs tends to push the locations of these earthquakes deeper. This is confirmed by three earthquakes located in central Taranaki by the permanent network during the temporary deployment that are 4–12 km deeper than the same earthquakes located by this study (Fig. 7). There is a small difference in the position of the seismicity in western Taranaki, with GNS hypocentres from the permanent network displaced to the west, again probably because of differences in the velocity models. The permanent network hypocentres beneath eastern Taranaki are almost all shallower than 33 km, this being an effect of the base of the crust in the standard New Zealand velocity model (Fig. 4), but are still concentrated in the lower half of the crust.

5.5 Volcano monitoring

All earthquakes located in Taranaki were normal tectonic events producing clear P and S phases, and even beneath Mt Taranaki we recorded no volcanic earthquakes (for which we would expect to see events with a harmonic waveform, a dominant frequency content of ca 1–5 Hz, indistinct phases (McNutt 2000a)). The broad-band seismographs meant that very long period volcanic earthquakes and tremors (those with periods longer than about 10 s) that are invisible to the short-period permanent network might have been recorded. A systematic search for such activity was conducted by low-pass filtering continuous, day-length seismograms and processing to produce Real-time Seismic Amplitude Measurement (RSAM)–like (McNutt 2000b) data in which long-period signals appear as spikes above the background long-period noise. However, there was no evidence for locally sourced long-period signals because everything could be explained by arrivals from distant earthquakes. Although only a short-term deployment, the large, dense, portable network meant that the capability to detect volcanic earthquakes beneath Mt Taranaki was significantly better than that of the permanent network. The lack of either long or very long period volcanic earthquakes, together with the low level of tectonic earthquake activity suggests that there are currently no active volcanic processes beneath Mt Taranaki. If significant seismicity is recorded beneath Mt Taranaki, particularly if some earthquakes appear to be volcanic in nature, then they should be treated as a possible eruption precursor.

The very shallow depth of earthquakes beneath Mt Taranaki itself is important for estimating the maximum lead-in time for precursory volcano–tectonic seismicity. If there is a magmatic intrusion, then the shallow brittle–ductile transition depth means that volcano–tectonic earthquakes are unlikely to occur deeper than 10 km. Volcanic earthquakes or tremor may occur at greater depths, as they do at Kilauea (Klein et al. 1987), but these may be difficult to locate accurately and may not allow the progress of the intrusion to be followed.

The permanent seismic network is capable of being used to locate earthquakes in Taranaki reasonably successfully and can reproduce most of the details seen by the temporary deployment provided that only the best hypocentres are considered; at least eight phase picks are necessary and no earthquakes with fixed depths should be used. However, depth estimates beneath Mt Taranaki from the permanent network alone are all too great. Earthquakes here form only a small fraction of the total Taranaki seismicity, but are the most important in terms of monitoring seismicity that might be precursory to a future eruption.

6 CONCLUSIONS

A large deployment of three-component, broad-band seismographs in Taranaki, New Zealand has confirmed the long-term distribution of seismicity, but has also allowed us to observe some important patterns not apparent in the long-term data set. The absence of earthquakes deeper than 10 km beneath Mt Taranaki can partly be explained by increased temperatures that are consistent with a nearby high heat flow anomaly and has important consequences for future volcano monitoring. Volcanic earthquakes are absent beneath Mt Taranaki, suggesting that active volcanic processes are currently unlikely. Seismicity in the CEFZ in western Taranaki exhibits previously unrecognized spatial and temporal clustering with earthquake swarms making up a significant proportion of the activity. Earthquakes in eastern Taranaki are concentrated along the TRL to 35-km depth, with the upper crust almost aseismic. This requires that the lower crust be drier and more mafic than the wet, quartzo-feldspathic composition often used to model crustal rheology. The absence of a change in maximum earthquake depth across the currently postulated position of the TRL suggests that this boundary is about 25 km north of the previously accepted location.

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Seismicity is just one of several aspects of Taranaki geology and geophysics that need to be understood better if volcanologists are to successfully interpret precursors to a future eruption. Further work will include imaging the 3-D structure, with an obvious target being the region of shallow seismicity beneath Mt Taranaki.

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