A three-dimensional image of shallow subduction: crustal structure of the Raukumara Peninsula, New Zealand

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SUMMARY
Earthquake arrival time data from a 36-station deployment of portable seismographs on the Raukumara Peninsula have been used to determine the 3-D \( V_p \) and \( V_p/V_s \) structure of this region of shallow subduction. A series of inversions have been performed, starting with an inversion for 1-D structure, then 2-D, and finally 3-D. This procedure ensures a smooth regional model in places of low resolution. The subducted plate is imaged as a northwest-dipping feature, with \( V_p \) consistently greater than 8.5 km s\(^{-1}\) in the uppermost mantle of the plate. Structure in the overlying plate changes significantly along strike. In the northeast, there is an extensive low-velocity zone in the lower crust underlying the most rapidly rising part of the Raukumara Range. It is bounded on its arcward side by an upwarp of high velocity. A viable explanation for the low-velocity zone is that it represents an accumulation of underplated subducted sediment, while serpentinization of the uppermost mantle may be responsible for the adjacent high-velocity region. The low-velocity zone decreases and the adjacent high-velocity region is less extensive in the southwest. This change is interpreted to be related to a change in the thickness of the crust of the overlying plate. In the northeast the crust is thinner, and subducted sediment ponds against relatively strong uppermost mantle, while in the southwest the crust is thicker, and the relatively weak lower crust allows sediment subduction to greater depths. A narrow zone of high \( V_p/V_s \) parallels the shallow part of the plate interface. This suggests elevated fluid pressures, with the distribution of earthquakes about this zone further suggesting that these pressures may be close to lithostatic. The plate interface at 20 km depth beneath the Raukumara Peninsula may thus be a closed system for fluid flow, similar to that seen at much shallower depths in other subduction décollements.

Key words: crustal structure, New Zealand, seismic velocities, subduction, underplating.

INTRODUCTION
The Raukumara Peninsula in the eastern North Island of New Zealand lies some 300 km southwest of the junction of the Tonga–Kermadec and Hikurangi subduction zones. To the north of this junction, oceanic crust of normal thickness is subducted at the Kermadec Trench, while to the south subduction is modified by the volcanic Hikurangi Plateau on the incoming Pacific Plate. This plateau has a crustal thickness of about 13 km to the east of the Raukumara Peninsula (Davy & Wood 1994), and the buoyancy of the subducted plateau has resulted in exposure of the forearc above the shallow part of the subduction thrust. The peninsula thus affords an excellent opportunity to study the local subduction process.

Geodetic strain and active faulting indicate that the eastern part of the peninsula is currently extending normal to the plate margin (e.g. Thornley 1996). Also, offshore multibeam bathymetry, side-scan sonar and seismic reflection data indicate that the margin is characterized by tectonic erosion (Collot et al. 1996). At the same time, the axial mountain ranges of the peninsula have been rising rapidly since at least the Pleistocene (e.g. Yoshikawa et al. 1980). What structures and processes at the plate boundary are required to reconcile these disparate observations?

Here we address this question using data from 36 portable digital seismographs installed throughout the Raukumara Peninsula for a five-month period in 1994 (Fig. 1). 28 of these were three-component instruments, and the data recorded, when combined with that from nearby permanent stations of the New Zealand national seismograph network, have allowed detailed 3-D modelling of both the \( V_p \) and the \( V_p/V_s \) structure of the region down to 80 km depth.
Figure 1. Tectonic setting and generalized geology of the Raukumara Peninsula. The arrow indicates the velocity of the Pacific plate relative to the Australian plate (DeMets et al. 1994). Bathymetry is in kilometres, and onshore fault traces are indicated by thin lines. Triangles are permanent stations of the New Zealand national seismograph network, and circles are portable digital seismographs deployed in this study. The star denotes the 1993 Ormond earthquake. Geology: vertical shading—Early Cretaceous greywacke; horizontal ticks—Late Cretaceous and Early Tertiary rocks of the East Coast Allochthon; unshaded region—Neogene marine sedimentary rocks. Abbreviations in the inset are: HP—Hikurangi Plateau; HT—Hikurangi Trough; KT—Kermadec Trench.

TECTONIC AND GEOLOGICAL SETTING

In the region of the Raukumara Peninsula, the Pacific and Australian plates are converging obliquely at about 45 mm yr$^{-1}$ (DeMets et al. 1994; Fig. 1). This convergence is accommodated by subduction of the Pacific plate and deformation of the overlying Australian plate. The general shape of the subducted plate has been revealed by compilations of seismicity located with the New Zealand national seismograph network (e.g. Anderson & Webb 1994). The dipping seismic zone associated with the subducted plate has a strike of 040° (Ansell & Bannister 1996), and extends to about 320 km depth beneath the western Bay of Plenty. At the east coast of the Raukumara Peninsula, the plate interface (as defined by the upper envelope of activity in the dipping seismic zone) is some 15 km deep, and dips to the northwest at about 10°. The peninsula thus overlies the shallow part of the plate interface, which may be capable of producing large subduction thrust earthquakes (e.g. Hyndman et al. 1997).

Structurally, the Raukumara Peninsula can be divided into three units (Fig. 1): (1) a western unit of Early Cretaceous greywacke in the Raukumara Range; (2) the East Coast Allochthon, a belt of Late Cretaceous and Early Tertiary rocks that were thrust towards the southwest over unit (1) during the earliest Miocene; and (3) an eastern unit consisting of Neogene marine sedimentary rocks that overlie the allochthon in the east and are faulted against it in the west. These Neogene rocks have been normally faulted, folded and intruded by diapirs (e.g. Thorneley 1996; Mazengarb 1984). Extension on margin-parallel normal faults has been taking place for at least
5–10 Myr, while diapirism began in the mid–late Miocene and widespread mud volcanism continues today. This indicates relatively high fluid pressures in these rocks, a fact corroborated by pore fluid pressure measurements in petroleum boreholes (Allis et al. 1997).

The western part of the peninsula is experiencing broad antiform uplift, reaching a peak (3 mm a⁻¹ over 125 ka) along the crest of the Raukumara Range, arcward of a belt of much slower uplift (0.5–10 mm a⁻¹) between the range front and the east coast (Thornley 1996; Yoshikawa et al. 1980). Rapid uplift of the Raukumara Range has been interpreted by Walcott (1987) as being due to sediment subduction and underplating at the base of the crust of the Australian plate. This interpretation is based on calculations of the sediment budget at the subduction zone since the early Miocene. It is clear from the geometry of the toe of the frontal wedge today that the margin is undergoing tectonic erosion and subduction of material from the front of the wedge. Multibeam bathymetry, side-scan sonar and seismic reflection studies (Collot et al. 1996) indicate that the toe of the margin is indented by 10–25 km to the east of the peninsula, relative to regions to the northeast and southwest. This is inferred to be the result of repeated impacts of the large seamounts that are abundant on the northern Hikurangi Plateau. The two most recent impacts have left the major Ruatoria and Poverty indentations in the margin (Fig. 1).

On a regional scale, geodetic data from triangulation measurements corroborate the geological deformation. These indicate high rates of horizontal strain across the eastern Raukumara Peninsula and insignificant horizontal strain in the greywacke Raukumara Range, and suggest that the incompetent rocks in the east may be sliding towards the trench off the axial ranges (Thornley 1996). The distribution of seismic strain on the peninsula is currently poorly known, and quantifying this was one of the major motivations for the seismograph deployment (Reyners & McGinty 1999). Two low-angle thrust events (Mₚ 5.7, 5.8) near the plate interface have been recognized in the eastern part of the peninsula (Webb & Anderson 1998). The two largest instrumentally recorded earthquakes on the peninsula occurred in 1914 (Mₛ 6.7, 6.5). These events were centred near Cape Runaway, and are interpreted to have occurred in the overlying plate (Reyners 1990).

DATA

When permanent stations of the New Zealand national seismograph network are included, an average station spacing of about 20 km was achieved throughout the peninsula, thus providing good control on earthquake locations. Data for an arrival-time inversion should be high quality and provide spatially distributed ray paths. Events were thus restricted to those with azimuthal gap < 180°, and with at least 10 phases from at least six stations. A spatially distributed data set was then selected which mainly included the best observed events. To improve ray path coverage, data were included from aftershocks of the Mₛ 6.2 Ormond earthquake of 1993 recorded on a network of portable seismographs (Reyners et al. 1998). This event occurred in the uppermost mantle of the subducted plate near the town of Te Karaka (Fig. 1). 80 per cent of the events had 18 or more observations and 39 per cent had 30 or more observations. The data set included 374 earthquakes that provided 6476 P arrivals and 3779 S arrivals. Linear regression on a Wadati diagram gave an initial V_p/V_s ratio of 1.71.

The distribution of seismicity is uneven. Most upper-plate events are in the southeast where the subducted plate is relatively shallow, and seismicity deepens to the northwest. Thus in setting up the velocity grid and inversion procedures, we pay particular attention to assumptions that determine the estimate of velocity in areas of low resolution, and we assess the spatial averaging of velocities among nodes in the 3-D model.

METHOD

Local earthquake data were used in a simultaneous inversion for hypocentres and 3-D V_p and V_p/V_s ratios (Thurber 1983, 1993; Eberhart-Phillips 1990, 1993; Eberhart-Phillips & Michael 1998). Our goal is to obtain a reasonable model for interpreting crust and mantle structure for use in locating earthquakes and computing ray paths (azimuths and take-off angles) for local mechanisms, and for including heterogeneous elastic moduli in deformation modelling. We would like to avoid a model with peculiar or unbelievable velocity anomalies, even if only in areas of low resolution, as such a model would be less straightforward to use for other purposes. We prefer to obtain V_p and V_p/V_s models rather than V_p and V_s models. V_p/V_s is similar to Poisson’s ratio and is an important parameter in characterizing rock properties and rheology. Particularly in cases such as ours, where the S data are less numerous and of poorer quality than the P data, V_s would be poorly resolved compared to V_p, making the interpretation of V_p/V_s variations difficult (Eberhart-Phillips 1990). Solving for V_p/V_s makes the assumption that given a 3-D heterogeneous V_p model and unknown V_s, it would be better to estimate a V_s model from the V_p using a constant V_p/V_s than to assume a homogeneous V_s model. The alternative assumption that homogeneous (1-D) V_s is always the best V_s estimate when there are few S observations would imply crustal lithology that had only 1-D variations in rigidity despite having known 3-D variations in compressibility. This would require large variations in Poisson’s ratio wherever the 3-D V_p pattern differed from the 1-D V_s. This issue can also be dealt with by applying V_p/V_s coupling between the V_p and V_s solutions in a V_p and V_s inversion (Michelini 1991).

The velocity of the medium is parametrized by assigning velocity values at the intersections (nodes) of a 3-D grid. The ray paths are calculated with an approximate 3-D ray-tracing algorithm that produces curved non-planar ray paths which are defined by points more finely spaced than the velocity field and unknown elastic moduli in deformation modelling. We would like to avoid a model with peculiar or unbelievable velocity anomalies, even if only in areas of low resolution, as such a model would be less straightforward to use for other purposes. We prefer to obtain V_p and V_p/V_s models rather than V_p and V_s models. V_p/V_s is similar to Poisson’s ratio and is an important parameter in characterizing rock properties and rheology.
than a 3-D model obtained by directly inverting for the final 3-D model from a 1-D initial model (Eberhart-Phillips 1993; Eberhart-Phillips et al. 1995).

The initial velocity model and initial hypocentres were obtained through simultaneous inversion for 1-D velocity, station corrections and hypocentres (Kissling et al. 1994). Layer thicknesses were varied between models, and all reasonable models indicated monotonically increasing velocity down to a Moho at about 34 km depth. The initial velocity model is shown in Table 1. The z grids were chosen at 0, 9, 18, 23, 31, 37, 45, 67 and 100 km. There are very few shallow earthquakes, with only 24 events being shallower than 12 km. Thus there is little information to determine uniquely shallow velocities. The deepest event is at 80 km, so there is almost no information on the velocities at 100 km and they will not be perturbed from the initial model. The 1-D model station corrections show a regional pattern with slower material to the southeast.

The velocity grid is shown in Fig. 2. The coordinates of the grid were chosen to parallel the pattern of velocity heterogeneity as indicated by the 1-D station corrections. The y-axis is positive to the southwest (216°), and the x-axis is positive to the southeast. The 2-D model has constant velocity in the y-direction, and thus the 2-D inversion with these coordinates will assume that, to first order, the regional structure is 2-D trending N36°E. This direction is close to the N40°E strike of the subducted plate in this region determined by Ansell & Bannister (1996). The x- and y-grids are not uniformly spaced.

Their locations are based on the distribution of earthquakes and stations, in order to achieve a relatively even resolution and attempt to image dominant tectonic features such as the plate interface.

The damping parameter for each inversion was chosen empirically by evaluating a trade-off curve of data variance and solution variance, as the damping will vary with the model grid and the data set (Eberhart-Phillips 1986). The selected damping greatly reduces the data variance with only a moderate increase in solution variance. The 2-D model (Fig. 3) achieved a 35 per cent reduction in data variance from the 1-D model. Next a coarse 3-D inversion was performed with eight x-grids

<table>
<thead>
<tr>
<th>Depth below sea level, km</th>
<th>Vp, km s⁻¹</th>
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<tbody>
<tr>
<td>0.0</td>
<td>4.70</td>
</tr>
<tr>
<td>9.0</td>
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<td>18.0</td>
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<td>8.41</td>
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<tr>
<td>100.0</td>
<td>8.56</td>
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</tbody>
</table>

Table 1. Initial velocity, linear velocity gradient between nodes, and initial Vp/Vs = 1.71.

Figure 2. Grid nodes (triangles) and initial locations of earthquakes used in the 3-D inversion. Pluses denote earthquakes shallower than 40 km, and crosses are deeper events. Earthquake symbol size is scaled to magnitude.
Figure 3. A depth section down the dip of the subducted plate of the 2-D model showing inversion nodes (small dots) and relocated hypocentres of the inversion events (pluses). (a) $V_p$, (b) $V_p/V_s$, (d) $V_s$, which is not solved for but is defined by the $V_p$ and $V_p/V_s$ solutions. The estimated position of the top of the crystalline crust of the subducted plate, as defined by the upper envelope of seismicity in the upper plane of the dipping seismic zone, is shown by the dashed line in each depth section. Note that this boundary is not specified in the model; velocities are defined by linear gradients between nodes. Resolution is displayed in (c) for $V_p$ and (f) for $V_p/V_s$ by the spread function and 50 per cent smearing contours for nodes that have significant smearing; that is, for which the resolution contour extends further than adjacent nodes. (e) 50 per cent smearing contours for $V_p$ for all nodes show that where there is adequate resolution (low spread function) the velocity is a reasonable spatial average of the volume centred around each node as defined by the grid spacing.
Station corrections were included in the 3-D inversions, and these are shown in Fig. 5. We evaluated inversions with and without station corrections. Since this data set has very few shallow earthquakes, we are unable to resolve uniquely shallow velocity and it is appropriate to include station corrections.

Figure 4. Map views of the 3-D $V_p$ and $V_p/V_s$ solutions at various depths. $V_p$ is contoured at 0.25 km s$^{-1}$ intervals, $V_p/V_s$ at 0.025. The velocity solution grid is marked by white dots, and station locations are denoted by triangles on the $z=0$ km maps. Pluses are relocated hypocentres of the inversion events, midway out to adjacent gridpoints in depth (thus the $z=18$ km maps include events between 13.5 and 20.5 km deep). 

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An inversion without station corrections had similar patterns to the model with station corrections, but was less smooth with a few strong velocity anomalies at $z = 9$ that may be due to near-surface velocity heterogeneity.

The inversion routine calculates the full resolution matrix. The resolution describes the distribution of information for each node, such that each row is the averaging vector for a parameter. The relative size and pattern of the off-diagonal elements show the way the information is smeared. For a node to be adequately resolved, its resolution should be peaked and should have no significant contribution from nodes that are not adjacent. For a succinct way of assessing the resolution, we calculate the spread function (Michelini & McEvilly 1991; Michelini 1991), which compresses each row of the resolution.
matrix into a single number that describes how strong and peaked the resolution is for that node. As described in detail by Toomey & Foulger (1989), the spread function is a better way to describe the resolution than by solely examining the diagonal element, since the diagonal resolution is very dependent on the grid spacing and the damping. They show that the best image fidelity will be obtained by selecting a grid based on earthquake and station distribution and geological features, even though the diagonal elements will be smaller than for a much coarser grid that would require less damping.

As defined by Michelini & McEvilly (1991), the spread function $S_j$ for a node $j$ is computed from all the elements $r_{jk}$ of the corresponding row of the resolution matrix, $r_j$, weighted by their distance $D_{jk}$ from the node,

$$S_j = \log \left[ |r_j|^{-1} \sum_{k=1}^{N} \left( \frac{r_{jk}}{|r_j|} \right)^2 D_{jk} \right],$$

with $|r_j|$ the norm of the row of the resolution matrix. This combines two issues in evaluating resolution: the amount of information and the smearing. The first factor makes the spread function small for nodes that have large resolution values. The summed terms make the spread function large for nodes that have significant averaging from other nodes and particularly large if they are more distant.

Toomey & Foulger (1989) show that there is not a universally applicable value to define the range of acceptable values of spread function, since the range will differ for each study area and grid. They show that the maximum spread function value representative of acceptable resolution should be selected by examining individual averaging vectors to check for significant smearing beyond the volume immediately surrounding the node. The range of values obtained for the same data set will vary depending on the chosen damping and grid spacing. For a coarser grid, the distances would be larger and $S_j$ would be larger. For greater damping, $|r_j|$ would be smaller and $S_j$ would be larger.

The spread function does not show the direction of smearing. This can be shown by plotting the averaging vectors for individual nodes (Toomey & Foulger 1989; Eberhart-Phillips & Michael 1993); however, such plots cannot be shown for every node. To show the direction of smearing for all nodes in a single plot, Kissling (1988) shows ellipses representing the ray density. For local earthquake tomography, the smearing is often from only one direction such that there is more averaging on one side of the node, rather than the averaging being centred on the node as implied by ellipses. Having the full resolution matrix, we would like to show a summary plot of the smearing effects as determined by the averaging vectors. Thus we create plots that show key contours of the resolution rows only for nodes that have significant smearing from beyond the adjacent nodes, and we call these contours smearing contours. Smearing contours do not indicate the overall resolution with $|r_j|$ the norm of the row of the resolution matrix. This and solution quality, which is given by the spread function, combine two issues in evaluating resolution: the amount of information and the smearing. The first factor makes the spread function small for nodes that have large resolution values. The summed terms make the spread function large for nodes that have significant averaging from other nodes and particularly large if they are more distant.

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Fig. 3(c) shows the spread function and the smearing contours for only those nodes which have smearing, such that the smearing contour extends beyond an adjacent node. For the 2-D model, we estimate that nodes with spread function $<1$ are well resolved and velocity is representative of the immediately surrounding volume. Near $x = -50$, some of the nodes contain vertical smearing, since all the ray paths there are fairly similar as there are few earthquakes above 40 km depth. The spread function and smearing contours for $V_p/V_s$ are shown in Fig. 3(f). Overall the resolution is lower than $V_p$ since there are fewer $S$ data and proportionally fewer from larger distances since it is more difficult to identify clearly an $S$ arrival. There is negligible resolution at the northwesternmost nodes and strong vertical smearing at $x = -50$. The smearing contours can aid in interpreting the velocity solution. For example, below 20 km the northwestern half of the $V_p/V_s$ model has significant vertical smearing so that the broad high-$V_p/V_s$ feature could be a narrower feature that has been blurred. Significant horizontal smearing is evident at the shallow southeasternmost nodes. For the 3-D model, we estimate that nodes with spread function $<2.5$ are well resolved and nodes with $2.5 < S_j < 3$ have acceptable resolution but some smearing.

Since our method simultaneously inverts for $V_p$ and $V_p/V_s$, there will also be some coupling between the $V_p$ and $V_p/V_s$ solutions. This part of the resolution matrix is not included in $S_j$, but a separate coupling function could be computed (Michelini 1991). We have examined the off-diagonal elements related to the coupling across $V_p$ and $V_p/V_s$ and have found that they are very small, generally much less than 10 per cent of the diagonal element and always much less than the off-diagonal elements among the $V_p$ or $V_p/V_s$ nodes that are included in $S_j$. Thus the trade-off between the $V_p$ and $V_p/V_s$ solutions is not a significant factor for evaluating and interpreting the Raukumara results and we do not show plots of the coupling function.

**RESULTS**

The regional structure is shown in the 2-D model (Fig. 3). In interpreting this structure, it is important to differentiate between the subducted and overlying plates. Here we approximate the top of the crystalline crust of the subducted plate with the upper envelope of seismicity in the upper plane of the dipping seismic zone (cf. Robinson 1986), and define the plate interface to be the boundary between this crust and any subducted sediment. Focal mechanisms of earthquakes below this interface indicate down-dip tension, implying that these events lie within the subducted plate, while thrusting is indicated for events at and above the interface (Reyners & McGinty 1999).

$V_p$ contours in the subducted plate tend to parallel the plate interface, particularly above 40 km depth, where resolution is good. The uppermost mantle of subducted plate is imaged as relatively high-velocity material. Crustal material (depth $<20$ km) is lower velocity southeast of $x = -13$, and to the northwest there is upwarping of higher-velocity material centred at approximately $x = -31$. The 2-D $V_p/V_s$ shows high $V_p/V_s$ just above the plate interface, particularly at the shallow part of this interface in the southeast. Where the $V_p/V_s$ is highest it appears to be bounded above and below by distinct shallowly dipping zones of seismicity, and the $V_p$ is 6.6 km s$^{-1}$. This high $V_p/V_s$ zone extends upwards at $x = 0$, so that the lateral gradient in $V_p$ is also associated with a lateral gradient in $V_p/V_s$. Also shown is the 2-D $V_s$ model, which is defined by the $V_p$ and $V_p/V_s$ solutions. As noted above, solving for $V_p$ and $V_p/V_s$ means the $V_s$ structure will be the same as the $V_p$ structure except in regions of resolved $V_p/V_s$ heterogeneity. Here the main difference between the $V_p$ and $V_s$ models is that the lower-velocity material just above the plate interface near $x = 13$ is more prominent in $V_s$.

We can consider the reliability of the velocity features by using the spread function and resolution contours. The crustal velocities southeast of $x = -20$ and above 40 km depth are well resolved in $V_p$ and moderately well resolved in $V_p/V_s$ (Figs 3c and f). The lateral gradient in $V_p/V_s$ near $x = -10$ is well resolved. The $V_p$ resolution near the plate interface is good to 40 km depth and we can have confidence in the dip of the $V_p$ contours in the crust and uppermost mantle of the subducted plate. The uppermost mantle of the subducted plate is certainly imaged as high velocity. At depths above 50 km, the resolution is good and the velocities are reasonable spatial averages. Below 50 km, this uppermost mantle is still imaged as a broad high-velocity feature but its actual shape and velocity are not as well resolved, so minor velocity variations cannot be interpreted.

The $V_p/V_s$ resolution degrades with depth and to the northwest. Below 20 km depth and northwest of $x = -10$, there is some information as indicated by the spread function less than 3.5 (Fig. 3f), but there is significant smearing across adjacent nodes, especially vertically. Thus the $V_p/V_s$ values are realistic average values, but over larger volumes than indicated by the grid spacing. The damped inversion provides a solution that is similar to having a coarser grid where there is weak resolution. In interpreting the $V_p/V_s$, we consider that the high observed just above the plate interface in the southeast extends northwest to at least $x = -13$, where it is imaged as a broad feature centred at 31 km depth (Fig. 3b). From the traveltime inversion image, we are unable to discern whether this actually is a broad feature or whether it is a narrower, stronger high that has been blurred into a broader, lower-amplitude feature.

The velocity will be marginally perturbed, if at all, from the 1-D initial model in areas of very low resolution. Thus the northwest peripheral nodes above 20 km are close to the initial model and we cannot really interpret the slight lateral velocity variations from $x = -49$ to $x = -75$. At 67 km depth, the only nodes with significant resolution and thus significant velocity perturbations are from $x = -49$ to $x = -13$. These show a lateral velocity gradient which implies relatively high velocity in the subducted plate and relatively low velocity in the overlying mantle. We know that the subducted Pacific plate extends deeper and further to the northwest, but we are unable to image that with our data.

The crustal structure imaged in the 3-D model shows variations from the 2-D model. The shallow southeastern low-velocity region varies along the coast. It is most prominent in the $y = -22$ and $y = -6$ cross-sections (Figs 6g and i), with velocities less than 5 km s$^{-1}$ above 10 km depth. The low-velocity region appears to be centred just inland from the coast, but note that since the resolution is somewhat lower at the peripheral node ($x = 55$, Figs 7g and i), the low velocity may possibly extend uniformly offshore to the southeast. In the northeast of our study area, where the coast trends to the north, the shallow low-velocity region is also centred just inland from the coast ($y = -80$, Fig. 6a), with velocities less than 5 km s$^{-1}$ above 10 km depth, forming a strong lateral gradient.
Figure 6. Depth sections down the dip of the subducted plate of the 3-D $V_p$ and $V_p/V_s$ model. Inversion nodes are shown by small dots, and pluses are relocated hypocentres of all events located, not just those used in the inversion. Each section shows hypocentres midway to the next section (thus the $y = -38$ section includes events between $y = -48.5$ and $y = -30$ km). Stars at the tops of the sections denote the coastline.
Crustal structure of Raukumara Peninsula, New Zealand

Figure 6. (Continued.)
Figure 7. Spread functions and 70 per cent smearing contours for depth sections of the 3-D $V_p$ and $V_p/V_s$ model.

at $x=0$. Again we might infer that the low velocity extends uniformly offshore southeastwards, but that is unresolved (Fig. 7a). In the southwestern cross-sections ($y=18$ to $61$, Figs 6m–q), a shallow low-velocity region is also imaged, but is somewhat less prominent and extends further inland from the coast.

The high-velocity region on the northwestern side of the model shows some strong changes from the 2-D model. There is
a particularly distinctive high-velocity feature in the $y = -22$ cross-section (Fig. 6g), where there are higher velocities from 0 to 45 km depth, with a velocity of 7 km s$^{-1}$ at 15 km depth. This feature apparently has a distinct vertical boundary to the southeast, and it appears to be spatially restricted. It is only slightly apparent to the northeast in the $y = -38$ cross-section (Fig. 6e), and there is a clear change to the southwest so that it is not imaged in the $y = -6$ cross-section (Fig. 6i).
spatial limits of the high-velocity feature in the \( y = -22 \) cross-section are not correlated with changes in resolution. This distinct high is part of the broader region of uplifted \( V_p \) contours, as shown by the 6.0 km s\(^{-1}\) contour (Figs 6a–q). However, the broad high has relatively lower velocity in its southeastern part at \( x = -22 \), including a low-velocity zone (LVZ) centred near \( z = 23 \) imaged from \( y = -59 \) to \(-22 \) (Figs 6c–g). The \( y = -80 \) cross-section has lower resolution (Fig. 7a) and so could be consistent with the LVZ extending northeast, but the LVZ decreases southwest of \( y = -22 \) since the resolution is fine for \( y = -6 \) and it is much less obvious there. The \( y = -6 \) to 18 cross-sections (Figs 6i–m) are similar to each other, even though the resolution is reasonable. Thus the velocity structure there is fairly uniform and fitted well by the 2-D model.

Lower velocity at the top of the subducted plate is evident to 67 km depth in the \( y = -22 \) cross-section. The \( y = -59 \) to \(-22 \) sections have the most resolution at depth so we may be seeing a better-defined deeper image there. Lower-velocity material (8.0 km s\(^{-1}\) at 45–67 km depth) apparently forms the mantle beneath the subducted and overlying plates. However, a high mantle \( V_p \) of 8.6 km s\(^{-1}\) in the overlying plate is evident in the northwest at 37 km depth in the \( y = -59 \) to \(-22 \) cross-sections. As shown by the spread function (Figs 7e and g), the resolution is adequate in the \( x = -38 \), \(-22 \) grids for 37 and 45 km depth at \( x = -49 \). Thus this high-velocity mantle in the overlying plate is believable and constrained in depth. The resolution degrades farther northwest (\( x = -75 \)) and farther southwest (\( y = -6 \)) so we cannot tell whether the high velocity is more extensive laterally.

For \( V_p/V_s \), the resolution is best for \( y = 6 \) to 18 in the upper 30 km (Fig. 7). These cross-sections (Figs 6l and n) show further increased \( V_p/V_s \) at 23 km depth in the southeast, relative to the 2-D model. Thus this high \( V_p/V_s \) just above the plate interface is a depth-limited feature that does not appear to extend above 18 km, and may possibly be limited to only a few kilometres in depth extent.

The shallower \( V_p/V_s \) pattern shows spatial variations within the 3-D model. The apparently vertical high-\( V_p/V_s \) feature at \( x = 0 \) appears very prominent at \( y = 6 \) to 39. This high-\( V_p/V_s \) feature is on the northwest edge of the region of relatively low \( V_p \) in the upper 20 km. This is a spatially limited feature and could not represent high \( V_p/V_s \) within the whole relatively low \( V_p \) region. The resolution is relatively even from \( x = -30 \) to 20, and we do not see broadening of the high to the southeast but rather decreasing \( V_p/V_s \). There is some near-vertical smearing upwards towards the northwest (Fig. 7n), so it may be possible that the high \( V_p/V_s \) is somewhat more spatially limited in its vertical extent.

As another means of evaluating the \( V_p/V_s \) image we ran a different inversion where the final 3-D inversion used a uniform \( V_p/V_s \) initial model. The resulting \( V_p/V_s \) image is similar to the progressive inversion where there is acceptable resolution with spread function less than 3, but it has little detail elsewhere. Thus the plate interface high \( V_p/V_s \) is not imaged offshore at \( y = -80 \) to \(-38 \) and the high \( V_p/V_s \) is not imaged below 30 km depth at \( y = 18 \) to 61. The shallow \( V_p/V_s \) pattern is very similar to the progressive inversion from \( y = -6 \) to 61, including the prominent vertical high-\( V_p/V_s \) feature. Further northeast, the \( y = -80 \) to \(-22 \) sections do not have as pronounced shallow features. There the uniform initial \( V_p/V_s \) solution shows similar features to the progressive inversion but some are centred slightly deeper or shallower as might be expected for vertical smearing. Since there is relatively little horizontal smearing, the horizontal position of the shallow features does vary with the initial model.

In the uniform initial \( V_p/V_s \) solution, the plate interface high \( V_p/V_s \) is imaged in all \( y \) sections onshore where there is some resolution, but it is only weakly shown where there is low resolution. For instance, in the northeast peripheral grid (\( y = -80 \)) it is just 1.76 instead of 1.80 as in the progressive inversion. Only well-constrained hypocentres were used in the inversion (Fig. 2), and these tend to be onshore. However, the complete set of earthquakes recorded during the seismograph deployment, shown in Fig. 6, contains numerous offshore earthquakes. We are evaluating converted phases from these events for additional information on plate interface properties, and the results show that the high \( V_p/V_s \) at the plate interface does indeed extend offshore (Eberhart-Phillips & Reyners, in preparation). Thus the assumptions made in the progressive inversion are reasonable and give a more useful estimate of material properties in the region than a 3-D inversion with a uniform initial model.

**DISCUSSION**

**Subducted plate**

The subducted plate is imaged as a northwest-dipping feature, with \( V_p \) consistently greater than 8.5 km s\(^{-1}\) in the uppermost mantle of the plate. Such a high velocity is consistent with uppermost mantle velocities determined to the southwest along the strike of the subducted plate from both earthquake and refraction studies (Galea 1992; Chadwick 1997). Because velocities below 50 km are not well resolved, the lower limit of this high-velocity region is poorly defined in this study. However, surface wave dispersion studies using broad-band stations deployed in this study indicate that the high-velocity region in the uppermost mantle cannot be much thicker than about 10 km, and is underlain by mantle with a more normal \( V_p \) of approximately 8.2 km s\(^{-1}\) (Brisbourne & Stuart 1998). Such a high-velocity lid in the uppermost mantle of the subducted plate provides a ready explanation for the efficient propagation of high-velocity phases observed from earthquakes along the strike of the plate in the Tonga–Kermadec region (Ansell & Gubbins 1986).

Relocated earthquakes in the upper plane of the dipping seismic zone have a relatively uniform distribution along the strike of the subducted plate (Fig. 6). However, the same is not true for earthquakes in the lower plane—these are more numerous in the northwestern half of the study area from \( y = -22 \) northeastwards. In the southeastern half of the area where the subducted plate is shallow, relocated seismicity often extends from the upper plane down to the lower plane. This indicates that brittle fracture can occur throughout the region where \( V_p > 8.5 \) km s\(^{-1}\). A demonstration of this was provided by the \( M_w \) 6.2 Ormond earthquake of 1993, which ruptured from the upper plane to the lower plane (Reyners et al. 1998). Late aftershocks of this event are evident in Fig. 6k, near \( x = 20 \). There has been much debate as to the genesis of deep seismic zones, with both stress and structure being invoked to explain the distribution of seismicity (e.g. Fujita & Kanamori 1981; Kao & Liu 1995). The structure of the uppermost mantle of the subducted plate revealed in this study is relatively uniform.
both along strike and down dip. Thus the variations that we see in the distribution of seismicity are more likely to reflect local stress changes (such as changes in coupling at the plate interface) than changes in structure.

Overlying plate

Shallow velocity structure in the overlying plate (Figs 4a–d) correlates well with surface geology (Fig. 1). Both V_p and V_p/V_s are relatively high in basement and Cretaceous–Palaeocene cover rocks in the western part of the peninsula, while they are relatively low in the Neogene sedimentary rocks in the southeast. Intermediate velocities occur in the region of the East Coast Allochthon. The strong gradient in V_p in the region of East Cape is consistent with the crustal seismic reflection and gravity measurements of Davey et al. (1997) off the northern part of the peninsula. These suggest that a major transcurrent fault (named the East Cape fault by Davey et al. 1997) occurs in this region, juxtaposing low-density forearc sediments to the east against a passive margin sedimentary sequence with a high-density basement to the west.

The region on the eastern edge of the Raukumara Peninsula with V_p < 5 km s^{-1} which parallels the coast between Mahia Peninsula and East Cape shows significant lateral variation (Fig. 4c). It is most pronounced between Gisborne and Tolaga Bay, where it extends to 12 km depth (Figs 6g and i). This change in structure along strike may possibly be the result of the subduction of large seamounts. Collot et al. (1996) have interpreted both the Ruatoria and Poverty indentations in the offshore part of the margin (Fig. 1) as evidence of recent subduction of large seamounts. V_p is relatively higher along the coast downstream of these indentations in the direction of relative plate motion (Fig. 4c), suggesting that passage of such large seamounts has caused local uplift of higher-velocity rocks in these areas. While this pattern is resolved in the onshore region, we cannot be certain of the pattern offshore where the resolution is poor.

Deeper within the overlying plate, the major feature of the 3-D V_p structure is the low-velocity zone extending northeast from y = –22, and the upwarp of higher velocity to the northwest (Figs 4g and 6a–g). The fact that this low-velocity zone underlies the most rapidly rising part of the Raukumara Range (Walcott 1987) suggests a direct relationship between the two. A viable explanation for the low-velocity region is that it represents an accumulation of underplated subducted sediment, as postulated by Walcott (1987) based on calculations of the sediment budget at the subduction zone since the early Miocene. Taking the area between the V_p 6.0 and 6.5 km s^{-1} contours in Fig. 6(g) as a guide, in excess of 20 km of underplated sediment may have accumulated. The situation beneath the Raukumara Range may thus be analogous to that beneath the Kodiak anticlinorium in Alaska, where seismic refraction and reflection data suggest that up to 24 km of underplated low-velocity rocks lie above the subducting Pacific plate (Ye et al. 1997).

What constitutes the higher-velocity ‘backstop’ to the low-velocity zone? This zone has a V_p of 7.0 km s^{-1} extending up to 15 km depth at y = –22, where resolution is good. The boundaries of the zone to the southeast and southwest are well resolved, but its spatial extent to the northwest is unconstrained. The large vertical extent of this high-velocity region suggests large-scale alteration of the crust and/or mantle of the overlying Australian plate. An appealing candidate is serpentinization of the uppermost mantle, caused by hydration of mantle olivine by the large amount of water carried down with the oceanic crust and the nearby accumulation of subducted sediments (e.g. Peacock 1993; Hyndman et al. 1997). The fact that the high-velocity region has its greatest vertical extent adjacent to the region where the low-velocity zone is thickest (i.e. at y = –22) is consistent with this interpretation. An upper-mantle velocity of 7.0 km s^{-1} corresponds to 30 per cent serpentinization, while 7.5 km s^{-1} corresponds to 15 per cent (Christensen 1966).

The fact that the low-velocity zone decreases and the adjacent high-velocity region is less extensive southwest of y = –22 suggests a major change in the structure of the overlying plate along strike. A possible explanation for this is a change in the thickness of the crust of the Australian plate. The gravity modelling of Davey et al. (1997) suggests a relatively thin crust of 17–19 km off the west coast of the peninsula in the northeast. A thinner crust to the northeast is also suggested in this study. For instance, the y = –38 section has a V_p of 7.6 km s^{-1} at 31 km depth (Fig. 6e at x = –31), suggesting upper-mantle material, while the equivalent node in the y = –6 section (Fig. 6i) has a V_p of 6.7 km s^{-1}, indicative of crustal material. A thicker Australian crust to the southeast is also suggested by the gravity modelling of Bannister (1989), which indicates a crustal thickness of 36–37 km in the southwesternmost part of our study area.

Herein lies an explanation for the change in the distribution of the low-velocity zone along the margin. In the northeast, where the overlying plate crust is thinner, subducted sediment will pond against relatively strong uppermost mantle. However, to the southwest, where the crust is thicker, the relatively weak lower crust will allow sediment subduction to greater depths. The proposed model is illustrated schematically in Fig. 8.

The most prominent feature in V_p/V_s in the overlying plate is the vertical high near x = 0 (Figs 4d and 6). This feature also shows variation along strike, being most prominent in the southwest from y = 6 to 39. Nur (1972) has shown that elevated fluid pressures can increase V_p/V_s, and strong overpressuring has been measured in boreholes up to 4 km deep along the eastern half of the Raukumara Peninsula (Allis et al. 1997). Thus excess fluid pressure, presumably derived from the dehydration of subducted sediments and oceanic crust, provides a viable explanation for the vertical high-V_p/V_s region. Indeed, the variation in this region along strike may be reflecting the fluid budget. In the northeast, some of the fluid from dehydration will be lost in serpentinization of the Australian plate mantle, with the remainder being expelled through the forearc east of the Raukumara Range. In contrast, in the southwest the bulk of this fluid may find its way back to the forearc, giving rise to the strong V_p/V_s anomaly seen there (see Fig. 8). If the vertical high-V_p/V_s region is indeed a region of high fluid pressure, it will be an important weak zone in the overlying plate, along which deformation will be localized. The fact that it occurs on the northwestern edge of a region of low V_p along the margin suggests that it represents a major structural feature. In the central part of the peninsula, it correlates with the eastern edge of the East Coast Allochthon, a region of significant mud diapirism which indicates overpressuring at depth (e.g. Thorney 1996), while in the north of the peninsula it correlates with the East Cape fault (Davey et al. 1997).
Plate interface

The most striking feature at the plate interface is the narrow zone of high $V_p/V_s$, which extends along the margin (Figs 4f and 6). A viable explanation for this zone is again elevated fluid pressure, presumably in subducted sediment at the plate interface. Apart from in the $y = -59$ and $-38$ depth sections, earthquakes tend not to occur where $V_p/V_s$ is highest, but rather in two distinct, shallowly dipping zones on either side—one in the overlying plate and the other in the subducted plate region (e.g. Fig. 3). In the creeping section of the San Andreas fault, Thurber et al. (1997) similarly find a high-$V_p/V_s$ zone with seismicity occurring on the edge of the zone and not within it. This suggests that fluid pressures at the plate interface may be close to lithostatic, promoting stable sliding, with stick-slip only occurring further from the interface where fluid pressures are lower. The concentration of high $V_p/V_s$ at the plate interface and the possibility that fluid pressures are close to lithostatic require that fluids at the interface are isolated from the overlying plate by some permeability barrier. The plate interface at 20 km depth beneath the Raukumara Peninsula may thus be a closed system for fluid flow, similar to that seen at much shallower depths in other subduction décollements, such as Barbados (Moore et al. 1987).

The occurrence of earthquakes within the highest $V_p/V_s$ region in the $y = -59$ and $-38$ depth sections suggests a change in the nature of the plate interface there. These earthquakes form a cluster beneath the coast centred near $x = 30$, $y = -47$ (Fig. 4f). They may represent deformation within a subducted seamount, possibly one associated with the formation of the Ruatoria Indentation in the margin. Such a seamount is not resolvable in the 3-D $V_p$ image, possibly because $V_p$ of a
basaltic seamount increases only moderately during subduction and may match the low velocities of nearby underplated rocks (e.g. Ye et al. 1997).

The model of the plate interface suggested is one of subducted seamounts surrounded by subducted sediment in which fluid pressures are close to lithostatic. This is similar to the model of the plate interface 100 km southwest of the Raukumara Peninsula along the strike of the subduction zone, deduced from a study of the M 5.6 Tikokino earthquake of 1993. This earthquake had very few aftershocks, leading Reyners et al. (1997) to suggest that it may have initiated at an asperity (seamount?) at the plate interface, and ruptured into subducted sediment lying in the conditionally stable frictional field.

CONCLUDING REMARKS

The 3-D inversion of arrival time data recorded during the dense seismograph deployment on the Raukumara Peninsula has successfully imaged major elements of the structure of the shallow part of the subduction zone. It has also revealed a major change in structure along strike in the overlying plate, which appears to be related to a lateral change in the thickness of the crust of the overlying plate. Further constraint on this change in crustal thickness should be provided by a subsequent dense seismograph deployment in the Taupo Volcanic Zone, immediately west of the area of this study (S. Bannister, personal communication, 1997). The lateral changes in structure in the overlying plate should be reflected in changes in coupling at the plate interface, and this has been investigated in a companion study (Reyners & McGinty 1999). The lateral change in crustal thickness also implies a change in the downdip length of the seismogenic zone (e.g. Hyndman et al. 1997), with this zone being longer in the southwest. This has implications for the hazard of large interplate earthquakes in the region.

It is noteworthy that the image of shallow subduction obtained in this study is very different from that obtained in a similar study in the northern South Island, some 500 km to the southwest along the strike of the subduction zone (Eberhart-Phillips & Reyners 1997). In that region there is no concentration of high Vp/Vs at the plate interface, and fluid derived from dehydration of the subducted crust is interpreted as migrating into the overlying plate. Also, there is no indication of subduction of sediment. These changes can be related to the significantly greater thickness (and hence buoyancy) of the subducted Hikurangi Plateau in that region (Davy & Wood 1994).

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