Lithospheric structure across oblique continental collision in New Zealand from wide-angle \(P\) wave modeling

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[1] Oblique convergence between the Australian and Pacific plates in South Island, New Zealand, has resulted in crustal thickening and distributed deformation within both plates. We measure this thickening and image the deformation with seismic wide-angle data along a 600 km long transect that spans the plate boundary. \(P\) wave arrival times from 34,000 rays are used to construct a two-dimensional model for seismic velocity with depth. Crustal velocities average 6.1 and 6.2 (±0.23) km/s for the Australian and Pacific sides of the boundary, respectively. Upper mantle velocities average 8.1 (±0.25) km/s. Distributed deformation is indicated by the following observations: a reduction in upper crustal and midcrustal velocities by 4% west of the plate boundary, ascribed to flexural bending stresses; a velocity reduction of 8% immediately east of the Alpine fault, linked to high pore fluid pressures; 17 km of crustal thickening to form a 44(±1.4) km deep crustal root that is offset 10–20 km SE of the highest mountains; a velocity reversal below the midcrust in the eastern part of the crustal root, caused either by heating or an artifact from seismic anisotropy; strong thickening of a 7.0(±0.4) km/s fast lower crust within the crustal root; and a low-velocity zone in the Australian upper mantle due to sampling the slow orientation of upper mantle anisotropy. INDEX TERMS: 7218 Seismology: Lithosphere and upper mantle; 8110 Tectonophysics: Continental tectonics—general (0905); 8124 Tectonophysics: Earth’s interior—composition and state; 8150 Tectonophysics: Plate boundary—general (3040); 8180 Tectonophysics: Tomography; KEYWORDS: continental collision, lithospheric structures, ray tracing, South Island, New Zealand


1. Introduction

[2] Structural images of the lithosphere at continental collision zones show us how rocks deform on scales of 10s to 100s of kilometers. The continental plate boundary in South Island, New Zealand, is particularly favorable to study continental collision for a number of scientific and logistic reasons. The orogen is relatively young and uncomplicated [e.g., Norris et al., 1990], the strain history is well known [Walcott, 1998], and the proximity to the sea to both sides of the orogen (central South Island is only 200 km wide) allows the acquisition of efficient, detailed and reversed onshore-offshore data [Okaya et al., 2002]. Finally, a low population density yet a good road infrastructure permit ready access into the mountainous areas.

[3] The geological and kinematic frame work for South Island are both well understood [Norris et al., 1990; Walcott, 1998; Beavan et al., 1999]. Seismicity studies have given us some indication of the deformation and a first order image of the structure [Reyners, 1987; Anderson and Webb, 1994; Leitner et al., 2001]. Furthermore, some exploratory explosion seismology work has shown the crustal thickness to be around 40 km [Smith et al., 1995]. What had been lacking were high-resolution images of the crust and upper mantle adjacent to the boundary and detailed seismic velocity analyses in regions of high strain, to answer questions regarding the strength of the plate boundary, the extent of the deformation, the role of the Alpine fault and coupling of the deformation between crust and mantle. Therefore the joint U.S.-New Zealand funded South Island
Geophysical Transect (SIGHT) experiment [e.g., Stern et al., 1997; Davey et al., 1998] was conducted to address these issues.

[5] High-resolution seismic wide-angle data were collected across the plate boundary in South Island, along two major transects (lines 1 and 2, Figure 1) [Okaya et al., 2002]. This paper presents the seismic data analysis and the structural results from line 2 of the SIGHT program. Line 1 was analyzed by van Avendonk et al. [1999], and a detailed study of the onshore part of SIGHT line 2 was carried out by Kleffmann [1999]. Two further transects were designed as tie lines crossing lines 1 and 2 (Figure 1). Line 3 is presented by Godfrey et al. [2001], and results of line 3W stem from A. Melhuish et al. (personal communication, 1998) and in parts are published by Scherwath et al. [2002]. In addition to the seismic data, a deep magnetotelluric sounding took place [e.g., Wannamaker et al., 2002], petrophysical work was carried out [e.g., Okaya et al., 1995; Godfrey et al., 2000], and gravity data were collected both offshore [Holbrook and Davey, 1996] and on land [Kleffmann, 1999; Chetwin, 1998] to complement the SIGHT experiment.

[5] Knowledge of the crustal structure in this part of New Zealand was long drawn from seismicity and velocity-depth functions to locate hypocenters [Haines et al., 1979; Reynolds, 1987], combined with S to P phase conversion [Calhaem et al., 1977] and gravity modeling [Reilly, 1962; Woodward, 1979; Allis, 1986] and supported by shallow seismic velocity measurements [Garrick and Hatherton, 1973]. A steady increase in the amount of data that have become available and a corresponding increase in model resolution has led from the development of simple models with just one depth for the base of the crust [Reilly, 1962; Calhaem et al., 1977] to relatively detailed models [Smith et al., 1995; Kleffmann, 1999]. The two-dimensional (2-D) model presented in this paper exhibits a further advance in detail and extent of the lithospheric structure across the collision zone. Alongside this 2-D model, the first 3-D models have emerged [Eberhart-Phillips and Bannister, 2002; Kohler and Eberhart-Phillips, 2002]. The construction of these 3-D models, however, was restricted to the use of first arrivals and hence omits reflectors, i.e., velocity discontinuities, and in addition may be affected adversely by seismic anisotropy.

2. Tectonic Setting

[6] New Zealand sits astride two tectonic plates, the Pacific plate in the south and east and the Australian plate in the north and west (Figure 1). About 45 Ma the Australian plate accelerated away from the Antarctic plate, and this introduced strike-slip movement and hence the plate boundary [Walcott, 1998; Sutherland et al., 2000]. Bending of the presumably once straight Matai terrane and its associated magnetic anomaly [Hunt, 1978; Sutherland, 1999] suggests 850 (±100) km of dextral movement [Sutherland, 1999; Molnar et al., 1999]. At least 460 (±20) km of this motion has been accommodated by the Alpine fault [Sutherland, 1999].

[7] At about 6.4 Ma ago the previously minor compression across the Alpine fault increased markedly [Walcott, 1998]. This event initiated the Southern Alps orogen [Norris et al., 1990] and the formation of a crustal root [e.g., Woodward, 1979]. The total amount of lithospheric shortening in central South Island is about 90 km [Walcott, 1998]. Total rock uplift has been estimated between 30 km [Wellman, 1979] and 60 km [Kleffmann et al., 1998]. Most of it has been eroded, leaving Mount Cook at a height of 3764 m as the highest point of the Southern Alps. Current uplift rates are estimated at up to 10 mm/yr [Wellman, 1979; Norris and Cooper, 2001], but high erosion rates [e.g., Adams, 1980; Willet, 1999] counteract an increase in surface elevation.
Current relative plate motion consists of 34–37 mm/yr of strike-slip and 10–13 mm/yr of compression [DeMets et al., 1994; Walcott, 1998; Beavan et al., 1999]. It is estimated that currently up to 70% of the plate motion is partitioned onto the Alpine fault [e.g., Sutherland, 1994; Beavan et al., 1999].

Geologically, central South Island can be divided into two provinces, consisting of Precambrian to late Mesozoic crystalline rock west of the Alpine fault, and mainly Mesozoic to early Cretaceous graywacke and methagraywacke basement rock in the east [Nathan et al., 1986; Field et al., 1989; Adams and Kelley, 1998]. From pressure-temperature [e.g., Grapes and Watanabe, 1992] and fission track [e.g., Tippett and Kamp, 1993] studies it has been estimated that the now outcropping metamorphosed rocks originate from depths of up to 21 or even 35 km.

3. Wide-Angle Data of Line 2

SIGHT line 2 was set out as a 600 km long transect to transgress both the deepest part of the South Westland sedimentary basin and the Main Divide near its highest mountain. The onshore part of line 2 describes a crooked line, and offshore the ship tracks of line 2 were a straight continuation of the line of instruments on land (Figure 1).

Land receivers for the recording of the onshore explosions consisted of IRIS/PASSCAL Reftek stations and Canadian PRS recorders, with a nominal station spacing of 400 m. For recording the offshore air gun shots on land, only the Reftek stations were used, nominally spaced at 1.5 km. Offshore, the air gun pops were recorded on seventeen ocean bottom seismometers and ocean bottom hydrophones; these instruments, hereafter referred to as OBS/H, were positioned at various spacings (Figure 1). Shot spacings for the offshore air gun shots were 50 m (fired every 20 s), and the seven shots of line 2 on land were distributed as shown in Figure 1.

Seismic data from line 2 consist of four complementary data sets, comprising approximately three quarters of a million traces. They are: 2 common depth point (CDP) stacked sections off both coasts (~30,000 traces); 17 OBS/H gatherings off both coasts (~65,000 shots); 7 shot gathers onshore (~2,800 traces); and 178 onshore-offshore receiver gathers (~678,000 traces).

As line 2 on land deviates from a straight line, analyzing the line in two dimensions meant introducing distortion for some shot-receiver pairs. True shot-receiver offsets need to be preserved in order to avoid a systematic shortening of ray paths [Zelt, 1999]. To exclude severe cases of geometric distortion, theoretical geometrical considerations resulted in the omission of about 10% of shot-receiver pairs onshore of line 2 [Scherwath, 2002].

Overall, the data quality is good. Minimal data processing was applied to improve the phase identification and arrival time picking. The onshore-offshore data processing consisted mainly of bandpass filtering and deconvolution, and the ocean bottom instrument data processing included bandpass and velocity filtering. The offshore multichannel seismic (MCS) data were processed by A. Melhuish (personal communication, 1998) and Harrison [1999], and the onshore shot data were processed by Kleffmann [1999]. To improve coherency, a 2:1 trace summation was applied to all but the MCS data, reducing the total number of seismograms to 400,000 traces.

Table 1. Summary of Arrival Time Picks

<table>
<thead>
<tr>
<th>Phase</th>
<th>Total Picks</th>
<th>$T_{avg}$ s</th>
<th>$X_{avg}$ km</th>
<th>$\Delta T_{avg}$ s</th>
</tr>
</thead>
<tbody>
<tr>
<td>Seafloor reflection</td>
<td>261</td>
<td>1.3</td>
<td>0</td>
<td>0.050</td>
</tr>
<tr>
<td>First sediment refraction</td>
<td>136</td>
<td>2.8</td>
<td>5</td>
<td>0.050</td>
</tr>
<tr>
<td>First sediment reflection</td>
<td>73</td>
<td>1.9</td>
<td>0</td>
<td>0.053</td>
</tr>
<tr>
<td>Second sediment refraction</td>
<td>172</td>
<td>3.4</td>
<td>7</td>
<td>0.055</td>
</tr>
<tr>
<td>Second sediment reflection</td>
<td>171</td>
<td>2.4</td>
<td>1</td>
<td>0.055</td>
</tr>
<tr>
<td>Third sediment refraction</td>
<td>224</td>
<td>4.6</td>
<td>11</td>
<td>0.079</td>
</tr>
<tr>
<td>Third sediment reflection</td>
<td>340</td>
<td>3.1</td>
<td>2</td>
<td>0.063</td>
</tr>
<tr>
<td>$P_g$</td>
<td>9702</td>
<td>15.0</td>
<td>80</td>
<td>0.114</td>
</tr>
<tr>
<td>$P_P$</td>
<td>8014</td>
<td>23.5</td>
<td>125</td>
<td>0.174</td>
</tr>
<tr>
<td>$P_mP$</td>
<td>6644</td>
<td>20.0</td>
<td>100</td>
<td>0.150</td>
</tr>
<tr>
<td>$P_n$</td>
<td>7502</td>
<td>26.6</td>
<td>156</td>
<td>0.143</td>
</tr>
</tbody>
</table>

Total number of picks, average arrival times ($T_{avg}$), offsets ($X_{avg}$), and picking uncertainties ($\Delta T_{avg}$) are given. Arrival time picks from MCS data are not included.

[8] In summary, the phases used (identified and picked) are (1) shallow reflections from MCS and OBS/H data of water bottom, sediment and basement reflectors; (2) sediment refractions from OBS/H and onshore shots; (3) crustal refractions ($P_g$) from onshore-offshore, OBS/H and onshore shot data; (4) reflections off the lower crust ($P_P$) from onshore-offshore, OBS/H and onshore shot data; (5) reflections off the Moho ($P_mP$) from onshore-offshore, OBS/H and onshore shot data; and (6) mantle refractions ($P_n$) from onshore-offshore data.

4. Modeling Method and Strategy

The method used here is Zelt and Smith’s [1992] popular and readily available ray-tracing method for seismic travel time inversion. It utilizes arrival times for any body wave phase and performs forward ray tracing through an isotropic velocity model to calculate root-mean-square (RMS) misfits between predicted and measured arrivals. The velocity model can then be adjusted manually to fit the data by forward modeling, or automatically by error minimizing in a least squares sense. As with travel time tomography, a starting model is required prior to modeling, and the inversion process needs to be performed iteratively. As the automatic model inversion is linearly minimizing the travel time residuals, the result is a local rather than global
minimum for travel time residuals, and linear inversion therefore necessitates an effective starting model.

[19] Limitations exist in the computing algorithm applied to the model for predicting the data. These limitations are, for instance, the assumptions of infinite frequency in the ray theory [Aki and Richards, 1980], or the misrepresentation of the real earth in a parameterized model. Therefore there may be cases where the predictability of seismic arrivals becomes limited, not because of an unrealistic model but because of the algorithm applied to it.

[20] During ray tracing, a resolution matrix is calculated which shows how well each model parameter is resolved, based on the number of rays and the uncertainties for each model node [Zelt and Smith, 1992]. Node errors are, however, to be considered as minima, the trade-off between velocity and depth adjustment is not taken into account [Zelt and Smith, 1992]. Also neglected are possible phase misidentifications (including picking of later arrival branches of one phase), modeling 3-D structures in two dimensions, and deviations of the station geometry from a straight line for lateral velocity variations.

[21] Although the inversion algorithm by Zelt and Smith [1992] is able to invert the entire model simultaneously, modeling was aimed to be carried out from top to bottom, in a “layer-stripping” manner. The reason for this approach is based upon previous studies of the velocity structure in central South Island, all of which exhibit a degree of complexity [Smith et al., 1995; Kleffmann, 1999; Eberhart-Phillips and Bannister, 2002]. This becomes important when creating a starting model for the inversion. A top-to-bottom modeling scheme enables the modeler to control the model, especially where large models comprise complicated structures and a large amount of data.

[22] Layer stripping assumes that the velocity and shape of one layer only depends on the refractions and reflections in that particular layer. Shallower phases do not usually penetrate deeper layers and do not have to be taken into account. However, deeper phases may constrain shallower...
parts, and there could exist a potential pitfall in smearing wrongly modeled shallow misfits into deeper sections by not taking into account deeper phases that constrain shallow parts. This problem can only be solved by assessing each possible case individually.

Different modeling strategies (inversion styles and goals) are discussed by Zelt [1999], and the strategy chosen here is an initial “fine-grid parameterization and prior-structure model” that converts to a “minimum-parameter/prior-structure model.”

5. Final Modeling

5.1. Shallow Structures

Offshore, shallow structures were determined from fitting two-way travel time (TWT) arrivals of strong reflectors in the stacked MCS data simultaneously with the shallow sediment refractions and reflections from OBS/H data. On land, only refractions from sediments could be modeled. Automatic inversion of these data quickly converged toward reasonable and well fitting shallow structures (Figure 3) with evenly distributed data misfits. Statistics for the data fit of the shallow structures offshore are given as of in Table 2, where $T_{\text{RMS}}$ is the RMS value of the travel time residuals, and $\chi^2$ is the normalized sum of the squares of all travel time residuals divided by their corresponding picking uncertainties.

The role of modeling sedimentary structures lay in determining accurate velocity depth structures for calculating takeoff angles and time shifts for all deeper penetrating phases. Layer boundaries within the sediments therefore mark velocity contrasts rather than age boundaries, and this has to be taken into account when interpreting the shallow parts of the velocity model. For example, in a recent

![Figure 4. Complete final proposed model of line 2 from structural inversion of wide-angle data using Zelt and Smith's [1992] method. Circled numbers mark locations of key features as described in text. Thick lines are modeled reflectors. Velocity contours are in km/s.](image-url)
interpretation of west coast sediments high amplitude, discontinuous reflections within the Plio-Pleistocene sediments were observed but not picked [Harrison, 1999]. These discontinuities have been included as layer boundaries in the model.

Overall, though, the depth to basement and to other horizons that mark changes in both age and velocity correlate well. At the location of the only drill hole on the line, Clipper-1 70 km off the east coast, located between OBS A2 and A3 (Figure 1), depths to major reflectors were correlated with zero-offset TWT to determine layer velocities, thus matching model and borehole data. At all other locations offshore, no direct depth control is available from drill hole information but from interpreted, depth converted seismic stacks.

5.2. Deep Structures

For modeling the middle to lower crustal section, the automatic inversion did not work well. Inversion attempts created some cases of layer crossing, strong velocity reversals, or extreme and unreasonable velocities. Often after inversions new rays could not be traced between shot and receivers where arrivals are observed. This unstable inversion behavior is attributed to the complexity of the structure. Therefore a large proportion of forward modeling was applied to determine carefully middle and lower crustal structures. In addition, several cases of multiple arrival time branches of the same phases occurred, and inverting the structure for those was impossible using Zelt and Smith’s [1992] algorithm. Instead, forward modeling helped in these instances. In total, five top-to-bottom modeling approaches were carried out to derive a final crustal model.

Mantle refractions \((Pn)\) were generally inverted together with Moho reflections \((PmP)\) for determining the depth of the Moho. Only for the last inversion was the Moho depth fixed and \(Pn\) arrivals inverted to determine all mantle velocities. Ray tracing for the \(Pn\) arrivals was sometimes achieved by tracing them as turning waves and sometimes as head waves. Turning waves are a feature of constant velocity layers, which is unlikely in reality. In some cases, however, the rays of turning waves did not bend around undulations in the Moho, possibly due to too low a velocity gradient for the mantle in the model, or the inadequacy of ray tracing to simulate finite frequency waves. In these cases \(Pn\) phases were traced as head waves.

The final velocity depth section as shown in Figure 4 is a smoothed version of the final output using Zelt and Smith’s [1992] model format. It consists of one water layer (offshore), three sedimentary layers offshore, and two onshore, an extensive midcrust with three nonreflective interfaces for changes in velocity gradients with depth, a lower crust, and an underlying mantle.

6. Model Resolution, Coverage, and Data Fits

Data fits for the shallow structures are discussed above, and therefore the sediments are not included here. Figure 5 shows the model resolution of the crust and upper mantle in the final model. Resolution values vary between 1.0 (perfectly resolved node) and 0.0 (unresolved node). Zelt and Smith [1992] state that a resolution greater than 0.5 can mostly be considered as “well” resolved.

Resolution values were calculated by non-two-point ray tracing, meaning that for each gather a suite of rays was traced for all crustal and mantle phases, and where rays emerged near observed arrivals, these rays were used to calculate the model resolution. In total, about 24,000 rays matched arrival time picks. However, where one phase produces multiple arrivals at one offset (triplication, bowtie effects) the program cannot determine which one is correct, and the resultant model resolution will be apparently higher than with two-point ray tracing where each arrival time pick is only used once. (Unlike the model resolution, the data fit decreases in cases of multiple arrivals of one phase at a single offset.)

Consequently, the seemingly good model resolution for the depth nodes in the center of the crustal root is...
probably less than automatically calculated. Single arrival time picks might have been used several times for calculating the model resolution. This is one of the main reasons why the model has been determined with controlled forward modeling, especially in this part of the model. However, the thick black marked reflectors in Figure 5 are the actual bounce points of rays used for determining the model, therefore higher resolution is still expected to occur here. For the remaining depth nodes, the lower crust and Moho reflectors seem well resolved, but the nonreflective interfaces without velocity contrasts (dashed lines in Figure 5) are poorly resolved. This is because these latter interfaces are just for defining velocity gradient changes within one layer. Only a change in the velocity gradient constrains the depth of such interfaces, and this can primarily only be forward modeled.

Resolution of the velocity nodes is good, especially in the mid crust (>0.7 for more than half the nodes). Some upper crustal velocities are of lower resolution, however, resulting from a lack of close offset arrivals near the nodes. This occurs offshore where the station spacing is sparse and also onshore near the west coast where the geometric distortion of the line prevented the use of many near offset arrivals. Lower resolution in velocity also occurs at the depth of such interfaces, and this can primarily only be forward modeled.

Ray coverage of all Pg, PlP, PmP and Pn phases in the crust and upper mantle can be seen in Figure 6. Instead of simple contouring of the number of rays used to calculate any change to a grid node (so-called hit counts per grid node), Figure 6 shows the dimensionless values of DWS (derivative weight sum) as relative measure of the ray distribution [Toomey and Foulger, 1989]. DWS values weight each ray according to its spatial separation from the grid node, which is superior to the unweighted hit count of data per node. The software of Zelt and Smith [1992] does not calculate DWS values, and therefore other seismic reflection tomography software by van Avendonk et al. [1998] was used to present the ray coverage here. The main differences between the two programs lie in the ray-tracing algorithm (forward ray tracing [Zelt and Smith, 1992] versus two-point ray tracing [van Avendonk et al., 1998]) and in the model definition (irregular, possibly coarse model node distribution [Zelt and Smith, 1992] versus regular fine grid parameter mesh [van Avendonk et al., 1998]). Figure 6, although not exact, should nevertheless be representative. The larger the DWS the better the data coverage.

A dense ray coverage exists for the model with an expected decrease toward the edges and the shadow zone in the root (Figure 6). The largest contribution of the coverage comes from the onshore-offshore data, seen as a large fan of high ray density from the center outward. Strongest focusing is observed in the upper mantle west of the root, where onshore-offshore Pn in the west merge as they turn at a shallow level. The deepest part of the crustal root is covered by only one (yet strong) onshore shot record, hence the low ray density here.

Selected data fits to the wide-angle data are shown in Figures 7–11. An overall fit for crust and upper mantle phases, as determined from Zelt and Smith’s [1992] method, is given in Table 3. The large misfit of PmP and the moderate misfit of PlP are attributed mainly to the triplication of the phases in the root. Several branches of a single reflection occur at the same offset (Figure 7). Model arrivals that do not match the data have misfits of several hundreds
of milliseconds, thus severely distorting the travel time residuals. Smoothing the model prevents this effect and improves the model fit, as shown by the bracketed values in Table 3.

7. Model Uniqueness and Uncertainties

[37] Given the high data coverage and generally good resolution of the model, together with the reasonable fit to the data expressed in travel time residuals ($T_{RMS}$) and their variance $\chi^2$ as defined above (see also Tables 2 and 3), it appears that the model represents the data well. However, the final solution for the model was largely based on manual and automated forward modeling mainly due to the apparent heterogeneity in the crust. Therefore the final model can be biased by the modeler and the modeling strategy chosen and hence is nonunique.

[38] Nevertheless, many of the features found in this study were also derived by others in the same localities; similarities and differences between the different studies are discussed below. Distinctive parts of the model are tested for uniqueness in section 8.

[39] Calculating or estimating model uncertainties needs to include error propagation. Zelt and Smith [1992] suggest to perturb individual model nodes until inversions cannot “heal” these perturbations. This essentially propagates the error, however it also requires the final model to be derived through inversions, in which case several alternative ways of uncertainty estimations can also be applied, such as checkerboard tests, inversions from

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**Figure 7.** Example of data fits for onshore shot data. (top) Ray paths corresponding to picked arrivals; only every second to tenth ray path is plotted; thick black lines are bounce points of deep reflections used in the modeling. (bottom) A full set of predicted $P$ wave arrival branches overlaying the seismic data. This high-quality shot gather was essential for defining the shape of the crustal root, in particular its western edge. See color version of this figure at back of this issue.
Figure 8. As Figure 7, but for ocean bottom data offshore west. A relatively shallow Moho is required in the west to fit the deep reflections observed around ~40 km offset. See color version of this figure at back of this issue.

Figure 9. As Figure 7, but for ocean bottom data offshore east. PIP and PmP arrivals indicate thickening of lower crust around model km 475. See color version of this figure at back of this issue.
Figure 10. As Figure 7, but for onshore-offshore west data. This receiver station (3008) is located halfway between SP25 and SP26 (Figure 1). \(PmP\) arrivals such as the ones shown here were used to limit the western extent of the root. See color version of this figure at back of this issue.

Figure 11. As Figure 7, but for onshore-offshore east data. This receiver station (2420) is located halfway between SP23 and SP24 (Figure 1). \(P\)P arrivals like the ones here exhibit the thickness variation of the lower crust eastward. See color version of this figure at back of this issue.
Table 3. Inversion Statistics for Deeper Structuresa

<table>
<thead>
<tr>
<th>Phase</th>
<th>Number of Data Points</th>
<th>TRMS, s</th>
<th>χ²</th>
</tr>
</thead>
<tbody>
<tr>
<td>Pg</td>
<td>8,060</td>
<td>0.100</td>
<td>1.066</td>
</tr>
<tr>
<td>PIP</td>
<td>7,170</td>
<td>0.217 (0.161)</td>
<td>1.441 (0.904)</td>
</tr>
<tr>
<td>PnP</td>
<td>4,958</td>
<td>0.302 (0.223)</td>
<td>5.013 (2.089)</td>
</tr>
<tr>
<td>Pt</td>
<td>3,817</td>
<td>0.139 (0.150)</td>
<td>0.898 (1.026)</td>
</tr>
<tr>
<td>Total</td>
<td>24,005 (16,558)</td>
<td>0.198 (0.179)</td>
<td>1.966 (1.282)</td>
</tr>
</tbody>
</table>

aValues in parentheses are from smoothing the model stronger (one depth node every 6 km instead of every 1 km). Pg could not be traced through the smoother model.

Table 4. Final Estimated Uncertainties in Depths and Velocities

<table>
<thead>
<tr>
<th>Layer</th>
<th>Z, m</th>
<th>ΔZ, m</th>
<th>V, m/s</th>
<th>ΔV, m/s</th>
</tr>
</thead>
<tbody>
<tr>
<td>Water</td>
<td>730</td>
<td>35</td>
<td>1500</td>
<td>0</td>
</tr>
<tr>
<td>Sediment 1</td>
<td>1,110</td>
<td>65</td>
<td>1980</td>
<td>140</td>
</tr>
<tr>
<td>Sediment 2</td>
<td>1,760</td>
<td>200</td>
<td>2480</td>
<td>500</td>
</tr>
<tr>
<td>Sediment 3</td>
<td>2,940</td>
<td>300</td>
<td>3630</td>
<td>250</td>
</tr>
<tr>
<td>Basement</td>
<td>22,490</td>
<td>900</td>
<td>6060</td>
<td>230</td>
</tr>
<tr>
<td>Lower crust</td>
<td>26,140</td>
<td>1400</td>
<td>6950</td>
<td>400</td>
</tr>
<tr>
<td>Mantle</td>
<td>8150</td>
<td>250</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

8. Model Features

8.1. Low-Velocity Zone Offshore West, Feature 1

[45] Reduced upper crustal and midercrustal velocities occur over a 25 km wide zone around 20 to 30 km west of the west coast. The amount of velocity decrease is between 3.3% and 4.5%, although these values are entirely dependent on the width of the zone of velocity reduction. A narrower zone would require a larger reduction in velocity. A 25 km wide zone was chosen as it creates a smoother model which produces more stable ray tracing. An even wider zone contradicts observed Pg arrivals to either side. Therefore the width of 25 km is a maximum and the amount of velocity reduction, up to 4.5% here, needs to be regarded as a minimum.

8.2. Low-Velocity Zone at Alpine Fault, Feature 2

[47] Another LVZ appears close to the SE dipping Alpine fault, extending roughly between the surface traces of the Alpine fault and the Irishman Creek fault in the final model. The maximum velocity reduction is 8% (at about 15 km depth); however, similar to the west coast LVZ, a relatively smooth final model was preferred over a sharp one, which means that a narrower LVZ with more velocity reduction is possible at the cost of accurate ray tracing.

8.3. Low-Velocity Zone at Alpine Fault, Feature 3

[48] Figure 13 shows the effect of not including this LVZ on onshore-offshore west Pg arrivals. A clear delay (200–300 ms) is needed to fit the data. A similar delay in arrival times can be observed in the onshore shot gather, where both Pg refractions and lower crustal reflection are affected [Kleffmann, 1999], and also in teleseismic P wave delays [Stern et al., 2000]. Kleffmann [1999] and Stern et al.
[2000, 2001] show delays of 300 ms in \(Pg\) and 800 ms in \(PmP\) arrivals travelling through this Alpine fault zone, indicating the depth extent of this LVZ.

8.3. Schist Velocities East, Feature 3

Toward the eastern part of the model a “wedge” of 6.3 km/s fast crustal material between 10 and 20 km depth is surrounded by lower velocities above and below and terminates in the west by the onset of the LVZ associated with the Alpine fault described above. This velocity is typical for New Zealand schists, but faster than previously modeled [Kleffmann, 1999]. For the first time, onshore-offshore data are used, therefore longer offset \(Pg\) arrivals were included into the modeling, and a 1.5% increase in midcrustal velocities from Kleffmann’s [1999] 6.22 to 6.3 km/s improves the fit as shown in Figure 14.

8.4. Velocity Reversal Around East Coast, Feature 4

A velocity reversal is modeled to occur beneath the zone of 6.3 km/s crustal velocities noted above. It is most apparent from where the crust thickens toward the root, i.e., from around 50 km off the east coast. Here, the lower midcrustal velocities (above the actual lower crust) decrease by about 3%. This velocity reversal is not required to fit \(Pg\) phases as it occurs below the maximum penetration depth of \(Pg\) phases. It is instead identified by the moveout of lower crust reflections (\(PlP\)). Figure 15 shows how observed \(PlP\) indicates lower average velocities than those produced by a downward continuation of \(Pg\)-derived midcrustal velocities. Since a lower moveout velocity is observed in both updip and down-dip \(PlP\) reflections (Figure 15), a change in reflector dip cannot produce a better fit.

8.5. Crustal Thinning at Western End, Feature 5

Coincident with the deepening of the sediments at the western edge of the model, a shallowing of the Moho by about 5 km is observed. This observation is based on two ocean bottom recordings (one of which is shown in Figure 8), where \(PmP\) phases were interpreted to arrive earlier than elsewhere. These arrivals could, however, stem from the lower crust (\(PlP\)) which was not picked in this region as only a single reflection branch was observed. Preference was given for this reflection to represent \(PmP\) for the stronger impedance contrast at the Moho rather than at the lower crust. This then resulted in a thin (~500 m) lower crust at the western end of the model.

8.6. Steep Western Flank of Crustal Root, Feature 6

Underneath the surface trace of the Alpine fault the Moho dips SE by about 45° in the model. Direct evidence for this feature comes from the shot record SP27 (Figure 7) where an observed delayed \(PmP\) reflection is interpreted to bounce off the steeply dipping western flank of the crustal root. Furthermore, the western end of the root is well

![Figure 12. Predicted \(Pg\) arrival times overlying a detailed section onshore-offshore west data of station near SP27 (Figure 1), from final model with (black) and without (gray) LVZ offshore west. Note that without LVZ, arrivals are predicted up to 200 ms earlier than observed. Delayed arrivals around –100 km offset travel through the half graben shown in Figure 3 at model km 115.](image)

![Figure 13. Predicted \(Pg\) arrival times for onshore-offshore west data of station near SP25 (Figure 1), from final model with (black) and without (gray) LVZ immediately east of the Alpine fault. Without LVZ, arrivals are predicted up to 300 ms earlier than observed. The shallow half graben structure (Figure 3) causes the delay around –165 km offset.](image)
defined by onshore-offshore west data (Figure 10), and reflections bouncing off the deeper part of the crustal root up to model km 225 (Figure 7). Interpolation between the defined parts of the Moho (thick lines in Figure 4) yields a minimum dip of $45^\circ$ for the western flank of the root.

8.7. Thickening of Lower Crust in Root, Feature 7

In the deepest part of line 2, where a 44(±1.4) km deep crustal root has been modeled, the midcrust doubles and the lower crust more than triples in thickness compared to outside the root. Whereas the top of the lower crust is defined by $P_{IP}$ phases from onshore shot gathers, the deepest part and therefore the actual shape of the Moho in the crustal root (between model km 225 and 275) is only observed on shot gather SP27 (Figure 7).

8.8. Thin Lower Crust at East Coast, Feature 8

A particularly thin part of the lower crustal layer exists beneath the east coast, at model km 375. Here, only the top of the lower crust produced observed reflections; the Moho being undefined by reflections and the depth to the Moho here is solely determined by $P_n$ refractions (for an example, see Figure 11). A thicker lower crust would delay $P_n$ arrivals, and higher mantle velocities would be necessary to compensate for this. Since the mantle velocity is already high (8.2 ± 0.25 km/s, Figure 4) in this region, a 1 km thin lower crust is a more likely solution.

8.9. Thickening of Eastern Lower Crust, Feature 9

Thickening of the lower crust is also observed at model km 475. The top of the lower crust is defined here by both far offset $P_{IP}$ recorded on onshore-offshore data east (Figure 11) and $P_{IP}$ phases recorded at OBS/H data east (Figure 9). The Moho is only defined by $P_{mP}$ from OBS/H data east. $P_{mP}$ from onshore-offshore data east appears to terminate due the larger than critical angles of incidence on the lower crust.

8.10. Low Upper Mantle Velocities Beneath Alpine Fault, Feature 10

Relatively low-velocity upper mantle material appears to exist underneath the surface trace of the Alpine fault and extending about 100 km west. This is analyzed in more detail and discussed as sampling of the slow orientation of seismic anisotropy by Scherwath et al. [2002]. The eastward extent of this zone is hidden due to a lack of $P_n$ beneath the root but may also extend up to 100 km where eastern $P_n$ observations commence.

9. Discussion

9.1. Shallow Structures

Offshore to the west, the concept of a foreland basin system has been adopted for the South Westland basin [Sircombe and Kamp, 1998; Harrison, 1999]. The zones of foredeep, forebulge and back deep are shown in Figure 3. The back deep zone, reaching over 3 km depth, is underlain by a 2 km deep basin of pre-Miocene sediments. Below the forebulge there is a 1.5 km deep half graben structure. Toward the coast, the up to 4.2 km deep South Westland basin comprises almost entirely post-Miocene sediments,

Figure 15. Predicted $P_{IP}$ arrivals for land shot data SP22 (down dip) and SP25 ( updip), reflecting off the lower crust between model km 280 and 330. Arrivals predicted earlier (gray) are from a model with midcrustal velocities of 6.3 km/s continued downward to the lower crust reflector. The hyperbolic moveout curvature in both updip and down dip reflections suggest a velocity reversal below the eastern midcrust to lower the average velocity of the crust and thus fit the data.
forming the sedimentary load deposited after plate collision commenced. Within these post-Miocene sediments, the velocities increase toward the coast, presumably correlating with increasing compaction toward the South Westland fault zone. A strong increase in shallow reflection amplitudes occurs in this zone [Harrison, 1999].

[55] On land, the West Coast Platform (Figure 3) comprises a sedimentary sequence with a constant thickness of about 1 km. This thickness is adopted from Kleffmann [1999] and not remodeled as the refractions determining this structure lie too far outside the model plane. It is assumed here that the sedimentary thickness does not change SW of the model plane where the relevant refractions were recorded. Across the Alpine fault, the basement refractions steps up by ~500 m, and the sedimentary cover thins gradually toward the Main Divide. East of the Southern Alps, the maximum thickness of sedimentary deposits in the areas of Lake Pukaki, Mary Burn, and Mackenzie Basin are about 1 km, 250 m, and 1.5 km, respectively (Figure 3). The Irishman Creek fault has been marked as a 55° SE dipping fault on the basis of gravity data [Chetwin, 1998], although from shallow seismic sections of medium quality other interpretations also exist [Long et al., 2003]. Around the Irishman Creek fault the upper basement velocities are slower than elsewhere. Sediments within the Cannington Basin are about 1.5 km thick, some 500 m of this being less than 2.0 km/s fast, presumed Quaternary gravel, which corresponds well with Langdale and Stern [1998]. The rest of the south Canterbury Basin was modeled with a steadily increasing sedimentary thickness (with the exception of where Oligocene limestone outcrops, Figure 3), of up to 2.4 km thick near the east coast.

[59] Offshore to the east, the sediments reach their maximum depth of 6 km in the Clipper Basin (Figure 3), with 3.5 km of sediments below what has been interpreted as the Oligocene reflector. These bottom sediments are presumably due to a late Cretaceous subsidence. A succession of highs and lows in the basement is observed, matching those described by Field et al. [1989] and interpreted to have formed by Cretaceous extensional faulting and volcanism. Throughout offshore Canterbury, Oligocene limestone occurs as a planar surface (less than 200 m deviation from a perfect plane along line 2), with a minor tilt SW [Field et al., 1989], <0.3° on line 2 (Figure 3). This is remarkable as the load imposed by post-Miocene sediments, forming the Endeavour High (Figure 3), appears not to depress the Oligocene reflector. The post-Miocene sedimentation rate is ~300 m/Ma, at least 3 times higher than older rates [Field et al., 1989]. This load must be compensated by the lithosphere, and its effective elastic thickness (T_e) can be roughly estimated to be of at least 25 km [Scherwath, 2002], more than the T_e of 15 ± 5 km estimated for the Australian plate off the west coast [Harrison, 1999]. It appears that after the Paleocene deformation, subsidence of the Oligocene layer occurred without significant deformation of the Pacific lithosphere.

9.2. Deep Structures

[60] An important confirmation of earlier proposed features by Smith et al. [1995] and Kleffmann [1999] is the reduction in midcrustal velocities to the east of the Alpine fault (Figure 4). More evidence for the existence of this low-velocity zone at the central part of the Alpine fault comes from teleseismic delays [Stern et al., 2000], other seismic wide-angle data [Stern et al., 2001], and from a 3-D first break tomographic inversion [Eberhart-Phillips and Bannister, 2002]. An interpretation of this low-velocity zone is that it indicates the presence of pore fluids under high pressure. Such an interpretation is supported by the exposure of schists that contain intrusions of large quartz veins close to the Main Divide in the Southern Alps, linking the metamorphism to excess water [Craw, 1997; Vry et al., 2001]. Furthermore, Wannamaker et al. [2002] concluded upon the presence of water in the crust from magnetotelluric data inversion. Such pore fluids under high pressure could reduce seismic wave velocities [Stern et al., 2001].

[61] Further corroborations for previous models exists for the asymmetric shape and maximum depth of the crustal root. The deepest part of the root, 44 km deep at about 10–20 km SE of the highest mountains (Figure 4), is imaged only by the onshore explosion data. The onshore-offshore data cover only the eastern and western edges of the root, and its asymmetric shape is not affected.

[62] Below the coasts, the crust is 27 (±1.4) km deep. As the crust is expected to thicken between the coasts due to loading from topography and mantle mass anomalies [Stern et al., 2000], this value can be regarded as the undeformed or precollisional crustal thickness. Coincidentally, 27 km is also the average Moho depth along all SIGHT lines, including the offshore regions. A “normal” crustal thickness of 27 km was also found in North Canterbury [Reyners and Cowan, 1993] and is close to the thickness of much of North Island that is distant from the plate boundary [Stern, 1987].

[63] Thus a crustal thickening of 17 km is estimated along SIGHT line 2. A root of only 8 km would be predicted by a mountain range of a height of 1500 m (average for central Southern Alps) and assuming a crust-mantle density contrast of 500 kg/m³. The remaining thickening is due to a pull of thickened mantle lithosphere directly below the crustal root [Stern et al., 2000].

[64] A departure from previous models exists for midcrustal velocities east of model km 300 (Figure 4, marked as feature 3). As stated above, a 1.5% increase in velocity compared to Kleffmann [1999] was necessary to fit onshore-offshore east data. Below about 15 km depth the velocities decrease again to match the observed slower average velocities from the PIP reflections. This velocity reversal has already been suggested [Kleffmann, 1999] and was the preferred alternative to match both observed reflection amplitudes and previous upper mantle velocities from Reyners and Cowan [1993]. This velocity reversal may be an artifact due to seismic anisotropy if the schists in this region are horizontally foliated [e.g., Grapes and Watanabe, 1992; Walcott, 1998]: The faster velocities are constrained by mainly horizontally travelling ray paths, whereas the deeper, slower velocities are constrained dominantly by vertical rays, and this corresponds to the expected orientations of the fast and slow P waves [Godfrey et al., 2000].

[65] However, some long offset reflections from the lower crust in the onshore-offshore east data do penetrate the deeper part of the crust below the east coast at shallow angles as low as 20° from the horizontal, which is already close to sampling the fast orientation [cf. Godfrey et al.,...
[70] Another new feature is the discovery of a low-velocity zone in the midcrust just off the west coast (Figure 4, marked as feature 1). Its velocity reduction is less than in the central low-velocity zone, and it appears more localized. As it coincides with the location where maximum curvature of the Top of Miocene reflector is observed [Scherwath, 2002], an origin linked to inferred bending [Harrison, 1999] and associated stresses is proposed. Some of the induced stresses may be absorbed by the opening of cracks at low confining pressures, which already lowers the seismic velocity, and the remaining extensional strains over these cracks cause a further decrease in seismic velocity [Brocher and ten Brink, 1987].

[71] Cross-line results from lines 3 [Godfrey et al., 2001] and 3W (A. Melhuish et al., personal communication, 1998) match in depth within 1 km, and in velocity around 0.1–0.2 km/s. The only significant difference is in upper mantle velocities at the west coast, where the velocity difference between lines 2 and 3W is about 8%, which is ascribed strong seismic anisotropy in the uppermost mantle [Scherwath et al., 2002].

[72] Over the entire length of almost 600 km along line 2, the crustal character remains continental; both crustal thickness and crustal velocities are in the ranges of continental environments [Meiøn, 1986; Christensen and Mooney, 1995]. Some crustal thinning is interpreted to occur at the western edge of the model where the Moho depth rises from 20 to 15 km (Figure 4, marked as feature 5). It is not accompanied by a change in water depth but a doubling in sedimentary thickness, from about 2.5 to 5 km. This area probably resembles a continental margin as it is in the vicinity of deep water (Figure 1), and so the western edge of line 2 may exhibit the transition from continental to oceanic crust.

10. Summary

[73] A combination of forward modeling and inversion of P wave phases has produced a structural model of the crust and upper mantle in central South Island, New Zealand. Apparent strong heterogeneity, particularly in the midcrust, meant the ray tracing and inversion algorithm used reached its limits, and a large amount of forward modeling was applied. The final model has well resolved model parameters and a large data coverage, and it fits the data reasonably well (normalized $\chi^2$ only marginally greater than 1). The uncertainty estimates for the P wave velocity distribution are low to intermediate for the basement (±0.23 km/s) and the upper mantle (±0.25 km/s), and variable (±0.14 km/s to ±0.50 km/s) for the sediments. Depth uncertainty estimates become progressively larger for deeper structures; they are tens to hundreds of meters for the sediments, 900 m for the lower crust, and 1400 m for the Moho.

[74] The model presented in this paper provides clearly defined structural features which not only confirm several which have been proposed earlier but also many that are substantially new. They are as follows:

[75] 1. Sediments over 4 km thick have been mapped off the west coast and over 6 km thick off the east coast of South Island.

[76] 2. A remarkably flat eastern Oligocene limestone reflector below a load of postcollisional sedimentary out-
wash indicates a relatively strong Pacific crust (effective elastic thickness at least 25 km).

[77] 3. An average velocity of 6.1 km/s has been determined for the Australian crust and 6.2 km/s for the Pacific crust.

[78] 4. Reduced midcrustal velocities are proposed for a zone about 25 km wide and around 20–30 km off the west coast; the velocity reduction is about 3.3–4.5% and regarded as a minimum; this LVZ may be due to opening of cracks associated with bending stresses and localized extension.

[79] 5. Another LVZ is confirmed for the midcrust immediately east of the Alpine fault; up to 8% velocity reduction results in an LVZ around 50 km wide, which is regarded as a maximum width; this LVZ was proposed to be caused by fluid pressure approaching lithostatic.

[80] 6. A velocity reversal has been deduced to occur in the crust below about 15 km depth within the eastern part of the crustal root underling the Southern Alps; a possible explanation for this is heating due to crustal thickening, however its occurrence as an artifact due to seismic anisotropy in the schist cannot be excluded.

[81] 7. Strong crustal thickening is apparent, with the basement layer doubling in thickness and the lower crust layer roughly tripling in thickness in the crustal root, thus accommodating the crustal shortening; a total crustal thickening of 17 km is estimated from a precollisional 27 km thick crust to form an asymmetric, 44±1.4 km deep root about 10–20 km SE of the Main Divide.

[82] 8. The crustal root commences immediately below the surface trace of the Alpine fault with a steep dip of at least 45° into its deepest part; the eastern flank is less steep.

[83] 9. Average mantle velocities are around 8.1 (±0.25) km/s, with a strong reduction to 7.8 km/s beneath and up to 100 km west of the Alpine fault caused by sampling the slow orientation of seismic anisotropy; off the east coast, a zone of slightly faster upper mantle material seems to exist.

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Figure 7. Example of data fits for onshore shot data. (top) Ray paths corresponding to picked arrivals; only every second to tenth ray path is plotted; thick black lines are bounce points of deep reflections used in the modeling. (bottom) A full set of predicted $P$ wave arrival branches overlaying the seismic data. This high-quality shot gather was essential for defining the shape of the crustal root, in particular its western edge.
Figure 8. As Figure 7, but for ocean bottom data offshore west. A relatively shallow Moho is required in the west to fit the deep reflections observed around ~40 km offset.

Figure 9. As Figure 7, but for ocean bottom data offshore east. PIP and PmP arrivals indicate thickening of lower crust around model km 475.
Figure 10. As Figure 7, but for onshore-offshore west data. This receiver station (3008) is located halfway between SP25 and SP26 (Figure 1). $PmP$ arrivals such as the ones shown here were used to limit the western extent of the root.

Figure 11. As Figure 7, but for onshore-offshore east data. This receiver station (2420) is located halfway between SP23 and SP24 (Figure 1). $PmP$ and $Pn$ arrivals like the ones here exhibit the thickness variation of the lower crust eastward.