# LITHOSPHERIC DEFORMATION AT THE SOUTH ISLAND OBLIQUE COLLISION, NEW ZEALAND

by

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#### Abstract

Lithospheric deformation is investigated within the Southern Alps oblique collision zone of the Australian and Pacific plate boundary. Seismological methods and gravity modelling are used to estimate seismic anisotropy, wave-speed anomalies and mass anomalies in the uppermost mantle. While seismic anisotropy is generally interpreted to result from Cenozoic mantle shear, wave-speed and mass anomalies can be explained solely by thermal contraction of mantle rocks that results from the downward deflection of isotherms during mantle shortening.

Along the eastern Southern Alps foothills and ~15° clockwise from their axis, earthquake Pn waves propagate at  $8.54 \pm 0.20$  km/s. This high wave speed is attributed to a high average Pn speed ( $8.3 \pm 0.3$  km/s) and Pn anisotropy (7–13 %) in the mantle lid beneath central South Island. Two-dimensional ray-tracing suggests that the crustal thickness is  $48 \pm 4$  km beneath the Southern Alps' southern extent near Wanaka (western Otago). Such a thickness represents an  $18 \pm 4$  km thick crustal root that is thicker than necessary to isostatically sustain the ~1000 m topographic load of this region. A mass excess is proposed in the mantle below the region of over-thickened crust to compensate for the crustal root mass deficit. Assuming that the crustal root represents a -300 kg/m<sup>3</sup> density contrast with the mantle lid, this mantle mass excess requires a minimum density contrast of  $35 \pm 5$  kg/m<sup>3</sup>,  $110 \pm 20$  km width and  $70 \pm 20$  km thickness that will impart a downward pull on the overlying crust. Teleseismic P, S and SKS receiver functions are calculated for Geonet stations RPZ, JCZ, WKZ and EAZ. At Canterbury station RPZ, delay times of conversions and their free-surface multiples imply a low Vp/Vs ratio of 1.60 for the upper crust. At western Otago stations EAZ and WKZ, delay times suggest crustal thicknesses of  $32 \pm 3$  km and  $39 \pm 4$  km, respectively, inconsistent with the gravity and other crustal thickness estimates. Therefore, conversions may be interpreted to arise from a lower crustal boundary. This interpretation supposes a low seismic contrast at the Moho and possibly partial eclogitisation in the lower crust. A mantle discontinuity is interpreted 15–30 km below the Moho from conversions with delay times of 7–9 s that display a move-out similar to that of direct conversions. Modelling suggests a rotation of the anisotropy symmetry axis and/or a minimum 0.3 km/s wave speed increase in association with the discontinuity.

Beneath the central Southern Alps, teleseismic P arrivals display relative travel-time advances of 0.3–1.8 s relative to those predicted from the Southern Alps crustal structure alone. The smallest and largest time advances are for rays arriving at azimuths perpendicular and sub-parallel to the plate boundary, respectively. The average time advance is consistent with a 0.5–0.6 km/s wave-speed anomaly within the surrounding mantle that is a body sub-parallel to the plate boundary, sub-vertical, 100–130 km wide and centred at about 110 km depth. However, smaller and larger wave-speed anomalies of ~0.3 km/s and ~1.1 km/s are necessary to explain the smallest and largest time advances, respectively. The difference is attributed to a minimum ~8 % seismic anisotropy in the shortened mantle.

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### Chapter 1

## Introduction

At convergent active continental margins, oceanic mantle lithosphere is shortened in the simple and efficient process of subduction. At continental collision zones, however, the process of mantle shortening and deformation are less well understood. Mantle shortening is, however, important because it represents a large and deforming mass anomaly within the earth.

The isostatic gravity anomaly map of New Zealand (Fig. 1.1) is one illustration of the action of the mantle lithosphere on the crust within an active plate boundary zone. Isostatic gravity anomalies indicate departure from the Airy isostatic equilibrium (Watts, 2001), where topographic loading is imperfectly balanced by the crustal thickness variations. Positive isostatic anomalies often represent regions of dynamic support of the lithosphere, while negative isostatic anomalies may represent zones of lithospheric downwarping. The Southern Alps oblique collision, for instance, is associated with a weak -30 mGal isostatic anomaly of regional dimension (Fig. 1.1). This negative anomaly indicates that the Southern Alps region is not in Airy isostatic equilibrium, and is consistent with a thickened mantle lithosphere applying a downward pull on the overlying crust (Scherwath *et al.*, 2006). The origin of this pull is, however, uncertain.



Figure 1.1: This map shows the New Zealand isostatic gravity anomaly onshore North and South Island (Reilly *et al.*, 1977) and the free-air gravity anomaly offshore (Seasat, 1999). Coordinates are in meters of New Zealand Map Grid. The Southern Alps oblique collision is associated with a weak  $\sim$ -30 mGal negative isostatic anomaly.

This research complements previous studies by addressing the following question: what is the mode of lithospheric deformation beneath the Southern Alps oblique collision ?

Regional and teleseismic earthquake data are used as well as gravity anomalies to quantify properties of the lithosphere beneath the Southern Alps. These properties are:

- seismic wave speeds,

- their variations across and along the plate boundary,

- their variations with azimuth, i.e. seismic anisotropy,
- crustal thicknesses,
- mass anomalies.

#### 1.1 New Zealand's tectonic setting

North and South Island of New Zealand are the emergent portions of a much greater New Zealand microcontinent that is bisected by the Australian-Pacific plate boundary (Fig. 1.2). West of North and South Island, the submerged continental crust is composed of the Challenger Plateau (Fig. 1.2). East of South Island and from north to south, the submerged crust is composed by the Hikurangi Plateau, the Chatham Rise and the Campbell Plateau (Fig. 1.2). These are pieces of continental crust that have been stretched and subsided below sea-level during Gondwana break-up (e.g. Bradshaw, 1989; Deckert et al., 2002; Kamp and Hegarty, 1989). The plate boundary is divided into three distinct tectonic zones. From north to south, these zones are: (1) the Hikurangi Trench off-shore east of the North Island, where the Pacific oceanic plate subducts beneath the Australian plate; (2) the Southern Alps oblique collision, where continental crusts of the Chatham Rise and Campbell Plateau have been juxtaposed next to and collide with that of the Challenger Plateau; and (3) the Puysegur Trench, where young Australian oceanic crust ( $\sim 45-30$  Ma; Wood *et al.*, 1996) subducts northeastward and below the southwestern South Island. Thus, the plate boundary is composed of two subduction zones of opposite polarities with a continental collision zone in between. The two subduction zones are linked by the Alpine Fault continental transform (Fig. 1.2; e.g. Kearey and Vine, 1996). This link contrasts to the well-known San Andreas transform, which links two spreading ridges (Atwater and Stock, 1998a).

#### **1.2** Tectonic history

New Zealand's origins are traced back to the Early Paleozoic (Cambrian-Ordovician). Metasedimentary sequences in Western Fiordland show a link with the Lachlan and the older Delamerian fold belts found in Australia and Antarctica (Gibson and Ireland,



Figure 1.2: New Zealand microcontinent, bathymetry (Smith and Sandwell, 1997) and the Australian-Pacific plate boundary delineated by the black curve. The *black arrow* indicates the relative plate motion vector (DeMets *et al.*, 1994).

1996), indicating that New Zealand was once part of Gondwana.

In Late Paleozoic, oblique subduction was occurring along the eastern Gondwana margin, where seafloor was being accreted (Mortimer, 2003). This paleo-subduction marks the boundary between two provinces of New Zealand's basement termed the Western and Eastern Provinces (Fig. 1.3; Bradshaw, 1993). The Western Province designates a Paleozoic Gondwana fragment that is separated from the Eastern Province by the Median Tectonic Zone (MTZ in Fig. 1.3). The Eastern Province, in contrast, is composed of Late Paleozoic and Mesozoic rocks that accreted along the Gondwana margin. These rocks are mainly greywackes and schists that were accreted in a deep water trench adjacent to Gondwana.

It is proposed that subduction and collision stalled as the mid-ocean spreading ridge reached the subduction trench, which is interpreted as the reason for a major change in the tectonic regime from compressional to extensional at ca. 105 Ma (Bradshaw, 1989). Continental extension from ca. 100 Ma and sea-floor spreading from 85 Ma to 55 Ma, i.e. from the late Cretaceous into the early Tertiary (Paleocene), followed subduction and collision. Sea-floor spreading led to the opening of the Tasman Sea and the separation of the New Zealand fragment from Gondwana (Hayes and Ringis, 1973; Weissel and Hayes, 1972). As a consequence, the early Tertiary geological history of New Zealand is dominated by extensional tectonics and subsidence of the landmass.

Since the Cenozoic at 45 Ma, the Australian-Pacific Euler pole has continuously migrated southward (King, 2000; Walcott, 1998). Rotation around the Euler pole has led to ca.  $850 \pm 100$  km dextral shear between the Australian and Pacific plates, 460 km of which have been accommodated by strike-slip on the Alpine Fault (Molnar *et al.*, 1999; Sutherland, 1999; Wellman, 1952) at an estimated rate of ~35 mm/yr (Cande and Stock, 2004; Walcott, 1998) since its inception. It is thought that the inception



from Bradshaw (1993)

Figure 1.3: Restored pre-late Cenozoic configuration of the Median Tectonic Zone (MTZ) between the Western Province and the arc Brook Street terrane (BS) and forearc Muhuriku terrane (Mu) of the Eastern Province (Bradshaw, 1993). The top right insert represents the current configuration of the Median Tectonic Zone.

of the Alpine Fault at  $\sim 23$  Ma was controlled by an Eocene passive margin that separated Paleozoic continental lithosphere of the Challenger Plateau from younger Eocene oceanic lithosphere (Sutherland *et al.*, 2000).

As the Euler pole passed south of South Island, the Australian-Pacific relative displacement evolved from purely translational to transpressional (e.g. Walcott, 1998). The convergence rate has progressively increased for the past 20 Myr with a possible speed up at 6.4 Ma (Walcott, 1998). Plate reconstructions estimate the total amount of convergence between 75 km (Cande and Stock, 2004) and 115 km (Walcott, 1998). Cande and Stock (2004) estimated a 75 km total convergence based on convergence continuously increasing and a rate of  $6.5 \pm 1.8$  mm/yr over the past 6 Myr, consistent with  $9.1 \pm 1.5$  mm/yr (Beavan *et al.*, 2002) determined from contemporary GPS velocities. Shortening has been accommodated by crustal thickening in southern to central South Island, giving rise to the Southern Alps (e.g. Wellman, 1979). In contrast, subduction was initiated at the southwestern end of South Island (Lebrun *et al.*, 2003), where progressive translation juxtaposed Australian oceanic crust with Pacific continental crust.

In conclusion, convergence in South Island is only recent (in the past 45 Myr) and has been minor compared to translation. As a result, both collision and subduction are largely oblique in South Island.

#### **1.3** Modes of lithospheric shortening at collision zones

In continental collision zones, how the crust thickens can be observed from reverse faulting and mountain growth. What is not so obvious is how shortening occurs in the mantle: as intra-continental subduction (Mattauer, 1986) or continuous thickening of the entire lithosphere (England and Houseman, 1986; Molnar, 1992). In intracontinental subduction, the strength of the delaminating uppermost mantle provides resistance to bending, and deformation by simple shear is localised at the interface with the subducting plate. In continuous thickening in contrast, the mantle lithosphere is a continuum, whose strength provides the resistance to deformation. Deformation is continuously distributed and accommodated as pure shear (called vertical lengthening by Sanderson and Marchini, 1984).



after Pysklywec et al. (2002)

Figure 1.4: Down-warp of isotherms for intra-continental subduction of mantle lithosphere after 9.6 Myr and 150 km convergence (after Pysklywec *et al.*, 2002, their model EX3 with lower crust and mantle dry olivine rheology and viscosity parameter A=4.85 x  $10^{-17}$  Pa<sup>-n</sup>.s<sup>-1</sup>, n=3.5). The *dashed line* denotes detachment of the lower crust.

Seismic anisotropy, which results from shear strain, should therefore be more localised in intra-continental subduction than in distributed thickening. Furthermore, both modes of shortening cause downwarp of isotherms in the upper mantle by pushing lithospheric material downward and into the warmer asthenosphere. This isotherm deflection produces a temperature anomaly in the mantle (Fig. 1.4, 1.5). Thermal contraction (Anderson *et al.*, 1992) of the lower temperature material results in a small density increase and, thus, a zone of faster wave speeds than in the surrounding region. The geometry of this zone should differ between both modes, being either symmetric (continuous thickening) or dipping and asymmetric (intra-continental subduction).



Figure 1.5: Hypothetical downwarp of isotherms and temperature anomalies for thickened lithosphere with 100 km of shortening smoothly spread over a zone 200 km wide. The negative temperature anomaly produces a high wave-speed anomaly (Stern *et al.*, 2000).

Both modes of lithospheric shortening have been proposed for the transition between the Hikurangi and Puysegur subduction zones of opposite polarities situated north and south of South Island, respectively. Early models of convergence suggested intra-continental subduction of Pacific mantle lithosphere (Fig. 1.4; Beaumont *et al.*, 1996; Beavan *et al.*, 1999; Waschbusch *et al.*, 1998; Wellman, 1979). More recently, continuous thickening has been proposed for South Island (Fig. 1.5) based on teleseismic P travel-time delays in the central Southern Alps that suggest the presence of a "fast" mantle zone with symmetric geometry (Molnar *et al.*, 1999; Stern *et al.*, 2000).

#### 1.4 Characteristics of the Southern Alps

#### 1.4.1 The crust

#### **Crustal thickness**

Crustal thickening is asymmetric and predominantly located east of the Alpine Fault trace (Scherwath *et al.*, 2003; Van Avendonk *et al.*, 2004). In central South Island, elevations rapidly increase eastwards from near sea level at the fault trace to a mean elevation of ~1,500 m (Koons, 1993) and a maximum elevation of 3,754 m at Mount Cook (Fig. 1.6). At mantle depth, the presence of a thick crustal root is indicated by a strong Bouguer gravity minimum of -95 mGal (Fig. 1.6; Reilly, 1962; Reilly and Whiteford, 1979; Woodward, 1979). The gravity low, however, is ~70 km offset from the maximum elevations to the south of Mount Cook and suggests that the thickest crust is similarly offset from the maximum elevations. Moreover, the gravity anomaly is oriented ~15° counter-clockwise from the strike of the mountain range. Extrusion of ductile lower crustal material perpendicular to the relative plate motion vector has been proposed as an explanation for the offset between the thickest crust and the maximum elevations (Gerbault *et al.*, 2002).

Maximum crustal thicknesses of  $\sim 37$  km (Van Avendonk *et al.*, 2004) and  $44 \pm 2$  km (Scherwath *et al.*, 2003) were estimated from inversion of reflection-refraction data along SIGHT (South Island Geophysical Transect) Transect 1 and Transect 2, respectively (T1 and T2 in Fig. 1.6). These respectively represent  $\sim 10$  km and  $\sim 17$  km crustal thickening relative to a 27 km thickness at the coast (Melhuish *et al.*, 2005; Reyners and Cowan, 1993; Scherwath *et al.*, 2003; Van Avendonk *et al.*, 2004). The 7.8 km/s wave-speed contours from 3D inversion of earthquake and active source seismic data (Eberhart-Phillips and Bannister, 2002) suggest a maximum 43 km thickness ca. 50 km south of Mt Cook, and also at the southern end of the gravity low.



Figure 1.6: South Island topography and approximate locations of the Otago, Fiordland and Canterbury regions. *Grey shaded regions* represent elevations greater than 800 m. The *black outlined vector* denotes the Australian-Pacific relative plate motion (DeMets *et al.*, 1994). *Colored contours* represent the Bouguer gravity anomaly (Reilly and Whiteford, 1979) in 40 mGal intervals. The *red star* indicates the location of Mt Cook. T1 and T2 denote the SIGHT (South Island Geophysical Transect) Transect 1 and Transect 2 (Okaya *et al.*, 2002). *Double arrows* show the central and southern portions of the Southern Alps and Alpine Fault (*thick grey line*) as used in the text.

#### Elastic strength

Flexural modelling suggests vanishing elastic strength for the central Southern Alps lithosphere assuming a continuous plate (Stern *et al.*, 2002). Stern *et al.* interpreted the low strength to be a result of ductile flow in the mid to lower crust and the upper mantle, along with upper crustal faulting.

#### The Alpine Fault

The Alpine Fault is a  $\sim 460$  km long dextral continental transform striking N55°E (Fig. 1.6; Wellman, 1953). Mylonitic foliations confined to a 1-2 km wide strip in the hanging wall of the central Alpine Fault (Fig. 1.6) suggest that the Alpine Fault is dipping at ca.  $33^{\circ}$  to the southeast of the surface trace (Sibson *et al.*, 1979). Seismic reflections from 20–30 km depth suggest a dip angle of  $\sim 40^{\circ}$ -55° when projected onto the Alpine Fault surface trace (Davey et al., 1995; Kleffmann et al., 1998; Stern et al., 2007). Away from its central portion (Fig. 1.6), however, the Alpine Fault is inferred to steepen from decreasing dip-slip rates to the northeast and southwest of this central portion and shortening being distributed further to the east of the Alpine Fault in Otago (Fig. 1.6; Norris and Cooper, 2001). Dip-slip reaches a maximum rate of 8– 12 mm/yr (Norris and Cooper, 2001) in the central portion of the Alpine Fault but decreases to ca. 6 mm/yr to the northeast and ca. 0 mm/yr to the southwest (Norris and Cooper, 2001), where deformation is more distributed. In contrast, strike-slip motion is relatively uniform with rates within  $27 \pm 5$  mm/yr (Norris and Cooper, 2001) along the entire fault. The Alpine Fault provides a ramp, along which crustal rocks are being uplifted (Little et al., 2002a; Wellman, 1953). High metamorphic grade rocks, the Alpine Schist, are found southeast and within  ${\sim}20~{\rm km}$  of the Alpine Fault. The Alpine Schist displays metamophic grades gradually increasing towards the fault up into the amphibolite facies (Grapes, 1995; Grapes and Watanabe, 1994, 1992).

#### Mid-crust

A zone of low seismic wave speeds with Vp ~ 5.5 km/s, i.e. a speed reduction of 6–10 %, was inferred for the mid-crust along SIGHT Transect 2 from travel-time delays of wideangle reflections and teleseismic P waves (Smith *et al.*, 1995; Stern *et al.*, 2001). This low-velocity zone (LVZ) was interpreted to result from high pore fluid pressure from water released during prograde metamorphism of the mid-crustal rocks (Stern *et al.*, 2001). The low wave speeds are associated with a low Vp/Vs ratio of 1.56 (Pulford, 2002) or 1.65 (Kleffmann, 1999), either of are consistent with the presence of pore fluids under high pressure (Marquis and Hyndman, 1992). Furthermore, the low wave-speed zone was linked to the low electric resistivity of mid-crustal rocks detected through magnetotelluric methods along the SIGHT Transect 1, which suggests the presence of interconnected pore fluids (Wannamaker *et al.*, 2002).

#### 1.4.2 The mantle lid

#### Seismicity

The low rate of intermediate-depth seismicity of the Southern Alps region provides a marked contrast to the well-defined Wadati-Benioff zones of seismicity of the neighbouring Hikurangi and Puysegur subduction zones (Anderson and Webb, 1994). This contrast points to a first-order difference in rheology and the manner in which relative plate motion is accommodated in the mantle. Only small  $M_L < 4$ , 30–97 km deep earthquakes have been recorded beneath the Southern Alps (Kohler and Eberhart-Phillips, 2003; Reyners, 1987). These earthquakes are proposed to represent brittle failure of 2 km<sup>2</sup> small patches in a viscously deforming mantle lithosphere and correlate with zones of combined high shear strain gradients and depressed geotherms (Kohler and Eberhart-Phillips, 2003) derived from strain modelling.



Figure 1.7: South Island mantle wave speeds. Small bold values and dotted lines denote Haines' (1979) Vp estimates and orientations from the measurement of regional earthquake travel times between station pairs. Italic values and double arrows represent Smith and Davey's (1984) Vp estimates and approximate orientations from travel-time inversion. The grey shaded zone is their interpreted partition of upper-mantle wave speeds. Black lines annoted T1, T2, 3W and 4E are SIGHT transects and corresponding wave speeds (Baldock and Stern, 2009, in prep.; Godfrey et al., 2001; Melhuish et al., 2005; Scherwath et al., 2003; Van Avendonk et al., 2004). The blue ellipse represents the ~120 km deep mantle wave-speed anomaly inferred from three teleseisms located northwest of New Zealand (Stern et al., 2000). Note that the ellipse is a schematic representation of the wave-speed anomaly, whose extent is not known.

#### P-wave speed

The southern half of South Island is characterised by relatively high mantle wave speeds compared to the worldwide average of  $\sim 8.05$  km/s for sub-Moho wave speeds (e.g. IASP91, Kennett and Engdahl, 1991). Estimates of  $8.3 \pm 0.1$  km/s (Haines, 1979) and  $\sim 8.4$  km/s (Smith and Davey, 1984) were found from regional earthquakes studies (Fig. 1.7).

Below the Southern Alps, the presence of a ca. 100 km wide high wave-speed body was inferred by Stern *et al.* (2000) from the advance of P arrivals relative to predicted travel times (IASP91, Kennett and Engdahl, 1991) of three teleseisms from the western Pacific (blue ellipse in Fig. 1.7). These three teleseisms were recorded by the 1996 SIGHT deployment (Okaya *et al.*, 2002) and have back azimuths in the range  $-60^{\circ}$  to  $-25^{\circ}$  from north. This analysis showed that a symmetric quasi-vertical high-speed body with a 0.5–0.6 km/s wave-speed anomaly, i.e. a 7–8 % wave-speed perturbation, offered a better fit to measured travel-time delays of this event set than an asymmetric (either west- or east-dipping) body did (Stern *et al.*, 2000). However, the three teleseisms constrain this model from the northwest only. Teleseismic delay-time tomography (Kohler and Eberhart-Phillips, 2002) also indicates high mantle wave speeds of 2–4 % perturbations (Kohler and Eberhart-Phillips, 2002) in this region. These correspond to wave-speed perturbations of 5–10 % if resolution tests, which indicate recovery of 40 % of the wave-speed perturbation, are taken into account. The tests also show that the resolution is too low to constrain the dip.

#### Seismic anisotropy

SKS splitting (Cochran, 1999; Duclos *et al.*, 2005; Klosko *et al.*, 1999) and Pn anisotropy (Scherwath *et al.*, 2002; Smith and Ekström, 1999) both indicate strong mantle seismic anisotropy beneath most of the New Zealand region (see Figure 2.4 of Section 2.5 that further discusses Pn anisotropy). SKS fast polarisations north of central South Island

are oriented N50°E and sub-parallel to the plate boundary (Klosko *et al.*, 1999). In the southern South Island, however, these are rotated counter-clockwise and oriented ca. N20°E (Klosko *et al.*, 1999). It is debated whether the general SKS fast polarisation alignment in New Zealand is due to mantle shear, mantle flow in the absolute plate motion reference frame, or frozen fabric in the mantle lid of the Pacific plate (Duclos *et al.*, 2005; Molnar *et al.*, 1999; Savage *et al.*, 2007b).

#### Mass anomalies

A ~100 km wide density contrast of 40–50 kg/m<sup>3</sup> with respect to the surrounding mantle was inferred beneath the central Southern Alps (Stern *et al.*, 2000) from comparison of the predicted Bouguer gravity anomaly for the SIGHT Transect 2 crustal structure with the observed one (Reilly and Whiteford, 1979). This density anomaly has been interpreted to represent continuously thickened lithosphere (Molnar *et al.*, 1999; Stern *et al.*, 2000). Three-dimensional inversion for this density anomaly, assuming a 40 kg/m<sup>3</sup> density contrast in the mantle lid and crustal thicknesses from previous studies, suggests that this anomaly is widening to the south, and may represent southward mantle creep (Scherwath *et al.*, 2006).

#### 1.5 Thesis outline

This introduction is followed by following chapters.

Chapter 2 reports on an earthquake refraction analysis along the Southern Alps foothills. The Pn speed and the dip of the Moho are estimated along a temporary deployment from Pn arrivals of regional earthquakes. Comparison with the Pn speed and crustal thickness along SIGHT Transect 2, enables the calculation of the Pn anisotropy at the intersection, and an estimate to be made of the maximum crustal thickness along the profile line. A minimum density contrast in the mantle lithosphere below is found from the misfit between the gravity effect predicted for the crustal root thickness with the Bouguer gravity anomaly.

Chapter 3 investigates the effects of anisotropy and dipping layers on S receiver functions via synthetics.

Chapter 4 uses freely available broadband data from the Geonet seismic network to measure conversions from teleseismic phases, so called receiver functions. The receiver functions are used to derive a lithospheric profile across the southern extent of the Southern Alps. The structure is estimated by forward modelling of P, S and SKS receiver functions.

Chapter 5 presents teleseismic P travel-time delays measured across the central Southern Alps from a wide range of back azimuths. A 3D forward modelling approach is utilised to constrain the size, amplitude and geometry of the mantle wave-speed anomaly beneath the Southern Alps.

Chapter 6 includes a summary of this thesis' findings, and a discussion of lithospheric deformation beneath central and southern South Island.

Two papers are included in the appendix, which have been published on the contents of Chapter 2. These papers are:

 Crust and mantle thickening beneath the southern portion of the Southern Alps, New Zealand (2007), Bourguignon, S., Stern, T. A. and Savage, M. K., *Geophys. Jour. Int.*, Vol. 168, pp. 681-690;
## 1.5. THESIS OUTLINE

 Crustal thickness and Pn anisotropy beneath the Southern Alps oblique collision, New Zealand (2007), Bourguignon, S., Savage, M. K. and Stern, T. A., in A Continental Plate Boundary: Tectonics at South Island, New Zealand, D. Okaya, T. A. Stern and F. J. Davey (eds), Geophysical Monograph Series, Vol. 175, American Geophysical Union, pp. 115–122.

# Chapter 2

# Earthquake refraction

# 2.1 Abstract

Over-thickened crust and fast, anisotropic mantle material are interpreted beneath South Island, New Zealand, from an earthquake refraction study along the Southern Alps foothills. An  $8.54 \pm 0.20$  km/s Pn speed is estimated along the N60°E striking refraction profile. Comparison with the Pn-speed estimate along the intersecting SIGHT Transect 2 at the northern end of the profile near Lake Tekapo suggests a high isotropic (i.e. average) Pn speed and anisotropy arising from finite strain of the mantle lid rocks. The Pn anisotropy is estimated to be a minimum of  $6.5 \pm 3.5$  %. A maximum Pn anisotropy of 7–13 % and an isotropic Pn speed of ~8.3 km/s are predicted by adopting the fast polarisation orientation from previous SKS splitting measurements at the profile intersection. The Pn speed of 8.3 km/s is consistent with findings from previous studies showing high average Pn speeds below the southern half of South Island and the presence of cold, dense mantle lithosphere.

A maximum crustal thickness of  $48 \pm 4$  km is inferred near Wanaka township, at the southern end of the profile. The crustal thickness represents an 18 km thick crustal root relative to a 30 km coastal average. Thus, the root is 2–3 times thicker than expected for Airy isostatic compensation of the mean ~1000 m Southern Alps topographic load. The thick crustal root suggests that the underlying mantle plays an active role in depressing topography. The mantle load is similar to that beneath the central Southern Alps, despite the predicted convergence across the Alpine Fault there being nearly twice that at Wanaka. Gravity modelling of the crustal structure along a profile through Wanaka suggests that this mantle load has a minimum density contrast of  $35 \pm 5$  kg/m<sup>3</sup> between thickened mantle and asthenosphere, assuming an across-Moho density contrast of -300 kg/m<sup>3</sup>.

A model is proposed, in which the nearby subducted Australian plate at the southwestern corner of South Island is the cause of enhanced lithospheric thickening beneath Wanaka. In this model, the subducted slab is a rigid backstop, onto which Pacific mantle collides at  $\sim 26$  mm/yr, or ca. 3/4 the full plate speed.

# 2.2 Introduction

Shortly after the  $M_W$  7.2 Fiordland earthquake (21<sup>st</sup> of August 2003), seven threecomponent seismographs were deployed along the Southern Alps eastern foothills and in-line with the RPZ (Rata Peak) Geonet station located northeast of the profile (Fig. 2.1). The profile line (herein called the Fiordland Cheviot profile) strikes at N60°E, i.e. 5° clockwise from the strike of the Alpine Fault and ~20° clockwise from the trends of the Bouguer gravity anomaly and supposed crustal root (see blue contours in Fig. 2.1). By striking obliquely and through the middle of the supposed crustal root, the profile minimises potential first arrival refractions off the sides of the crustal root.

The Fiordland earthquake aftershocks and one earthquake off-coast Cheviot (Fig. 2.1, Tab. A.2) at the northeast end of the Fiordland Cheviot profile allows the analysis of refraction travel times along the Southern Alps crustal root and to thereby estimate



Figure 2.1: This map depicts the Fiordland-Cheviot refraction profile, South Island fault lines (*light lines*), lake contours (*black lines*) and the Bouguer gravity anomaly in 40 mGal intervals (*coloured contours*; Reilly and Whiteford, 1979). Symbols feature the Australian-Pacific relative plate motion (*open arrow; after DeMets et al.*, 1994), deployed seismographs (*red triangles*), the RPZ Geonet permanent station (*black-contoured red triangle*), earthquakes used in this study (*red stars*), previous SIGHT seismic lines T1, T2, 3W and 4E (Okaya *et al.*, 2002) and this study's Fiordland-Cheviot refraction profile (*thin lines*).



Figure 2.2: Top: Arrivals from the Fiordland aftershock are bandpass filtered at cut-off and corner frequencies of 0.5-1-5-10 Hz. First-break Pn are indicated by the bottom pair of arrows and single arrows in blow-up on the right. Dashed curves denote predicted Pg and Pn travel-times (see model Fig. 2.3c). Pn arrivals are followed ca. 1.5 s later by arrivals (~1.5-s peg-leg indicated with top pair of arrows) with much larger amplitude. These second arrivals have the same apparent wave speed as the Pn and are interpreted as an internal reflection ~5 km near the source. The Pn-speed estimate and corresponding 95 % confidence interval (right-hand side of graph) is the mean of single regression slopes weighted with their respective standard deviations. Pn arrivals are followed ca. 1.5 s later by arrivals (~1.5-s peg-leg indicated with top pair of arrows) with much larger amplitude. These second arrivals have the same apparent wave speed as the Pn and are interpreted as an internal reflection ~5 km near the source of arrows) with much larger amplitude. These second arrivals (~1.5-s peg-leg indicated with top pair of arrows) with much larger amplitude. These second arrivals have the same apparent wave speed as the Pn and are interpreted as an internal reflection ~5 km near the source. The Pn-speed estimate and corresponding 95 % confidence interval (right-hand side of graph) is the mean of single regression slopes weighted with their respective standard deviations.

Bottom: Arrivals from the  $M_L$  4.1 off-shore Cheviot event are bandpass filtered at cut-off and corner frequencies of 0.5–1–3–5 Hz. Note the offset axis is in the opposite direction to that of the top figure. First-break Pn are indicated by the pair of arrows and the predicted Pn travel-time curve by a dashed curve (see P-wave speed model of Fig. 2.3c). The Pn-speed estimate is the result of a single linear regression and is given with corresponding 95 % confidence interval (right-hand side of graph). In both graphs the trace of the third station from the left was shifted by 3.5 s to correct a timing error. However, the pick wasn't included in Pn-speed calculations because of uncertainty in the timing error. Note also the two overlapping traces recorded at two close stations MCV and BAP (Fig. 2.1). seismic properties, density and temperature distributions in the mantle lid. The Pnwave speed is determined along the profile line. The Pn anisotropy and thickness of the crustal root are estimated beneath the central Southern Alps and their southern extent, respectively. The gravity effect of the crustal root is modelled in order to define how much of the crustal root thickness is due to topographic loading, and how much can be ascribed to the positive load of a mass excess in the subjacent shortened mantle lithosphere.

# 2.3 Refraction analysis

In the two-week deployment period, five Fiordland aftershocks of  $M_w \ge 5$  occurred at the southwest end of the profile and one  $M_L$  4.1 earthquake occurred off-coast Cheviot at the northeast end of the profile line (Tab. A.2). Maximum epicentral distances of 490 km enable the picking of Pg and Pn first arrivals and analysis of refraction travel times along the root of the Southern Alps. Pg and Pn apparent speeds are determined for single events from the inverse of regression slopes on the first break picks (Fig. 2.2). A single speed value is obtained by weighting the individual speeds with their corresponding inverse standard deviations (Bevington, 1969, Appendix A). The true Pn speed and apparent dip are calculated by assuming a uniform dipping Moho along the refraction profile (e.g. Stein and Wysession, 2003). Error bars presented are 95 % confidence intervals.

#### Crustal wave speeds

Crustal phases are seen in the Fiordland aftershock records (top of Fig. 2.2) but not in the record from Cheviot (bottom of Fig. 2.2). The determined Pg speeds of the Fiordland records are relatively high and show spatial variations with approximately 6.8 km/s for events 1, 4 and 6, and approximately 6.4 km/s for events 2 and 3 (Fig. 2.1). These speed values and offsets of 100–250 km suggest that the lower crust is being sampled by raypaths. These Pg speeds are slightly smaller, but consistent with values of 6.7–6.9 km/s at 4–8 km depth and 7.1–7.4 km/s from 8 km depth as determined from seismic refraction profiles in the exhumed Fiordland crustal block (Davey and Broadbent, 1980). They are also consistent with 6.25–7.5 km/s from 4 to 62.5 km depth from 3D inversion of local earthquakes (Eberhart-Phillips and Reyners, 2001). The 6.4–6.8 km/s wave speeds are not representative of the lower 6.0 to 6.2 km/s average P-wave speed in the Southern Alps mid-crust but with the ~6.8 km/s in the lower crust (Eberhart-Phillips and Bannister, 2002; Scherwath *et al.*, 2003; Van Avendonk *et al.*, 2004).

#### Pn-wave speed

The apparent Pn speeds determined from the off-shore Cheviot event and the reverse events in Fiordland are  $8.21 \pm 0.22$  km/s and  $8.92 \pm 0.18$  km/s, respectively (Fig. 2.2). Taking a 6.0–6.2 km/s mid-crustal wave speed (more representative of the Southern Alps crustal wave speed than the Fiordland 6.8 km/s), i.e. a 6.1–6.23 km/s average for the entire crust, results in an average Pn speed of  $8.54 \pm 0.20$  km/s and an apparent  $2.5^{\circ} \pm 1.3^{\circ}$  SW dipping Moho. The dip is a conservative value compared to an apparent dip value of  $2.6^{\circ}$ – $3.1^{\circ}$  SW for a 6.4–6.8 km/s crustal wave speed, but is also a much lower value than the apparent  $\sim 8^{\circ}$  SW dip calculated between where Transect 1 (Van Avendonk *et al.*, 2004) and Transect 2 (Scherwath *et al.*, 2003) intersect with the Fiordland-Cheviot profile. The inconsistency may result from the assumption of a uniform dipping Moho.



Figure 2.3: a) The contour map features the Bouguer gravity anomaly (Reilly and Whiteford, 1979) in 50 mGal intervals and the locations of Mt Cook (*triangle*) and of the main divide (*dashed line*) and Fiordland-Cheviot (*solid line*) profiles of the graph below. b) The *thin curve* is the mean topography in a 10 km wide swath along the Fiordland-Cheviot profile; the *thick dashed curve* is that along the Main Divide; and the *thick curve* is the Bouguer anomaly (Reilly and Whiteford, 1979) along the Fiordland-Cheviot profile (Fig. 2.1). c) 2D velocity model based on results from: Davey and Broadbent (1980); Eberhart-Phillips and Bannister (2002); Eberhart-Phillips and Reyners (2001); Reyners and Cowan (1993); Scherwath *et al.* (2003); Van Avendonk *et al.* (2004). Deployment (*triangles*); intersections with SIGHT previous crustal studies (T1/T2); ray-tracing for events 5 and 6 (predicted travel-time curves on Fig. 2.2); constrained portion of the Moho between Wanaka and Tekapo (*thick dashed line*); unconstrained interfaces (*question marks*).

## 2.4 Velocity model and crustal thickness

Previous crustal studies in South Island (Tab. A.4) are included to constrain the crustal structure (Fig. 2.3). The Fiordland-Cheviot profile intersects SIGHT Transect 2  $\sim$ 25 km east of the maximum crustal thickness of 44  $\pm$  1.4 km (Scherwath *et al.*, 2003). There, the Moho is  $\sim$ 42 km deep (Scherwath *et al.*, 2003) and accordingly fixed to 42 km depth. Similarly, the Moho depth is fixed to 33 km at the intersection with SIGHT Transect 1 (Van Avendonk et al., 2004). South of the crossing with SIGHT Transect 2, the dip of the Moho is set to  $2.5^{\circ}$  and the upper mantle wave speed is set to 8.54 km/s, as determined above. Two-dimensional ray-tracing (MacRay Luetgert, 1992) is applied to determine the Moho ray coverage. Rays propagating from both source locations, Fiordland and off-shore Cheviot, indicate that an approximately 150 km Moho portion, extending from Wanaka (southern South Island) to Tekapo (central South Island), is constrained. A maximum Moho depth of  $48 \pm 4$  km is estimated near Wanaka at the southwestern tip of this zone (Fig. 2.3). Hence, the crustal root is  $18 \pm 4$  km thick at Wanaka (relative to average coastal values of 30 km in South Island; Godfrey et al., 2001; Melhuish et al., 2005) and  $\sim 4$  km thicker than imaged along SIGHT Transect 2 near Mount Cook (Scherwath et al., 2003) suggesting thickening of the crustal root from north to south along the Southern Alps.

# 2.5 Pn anisotropy

#### 2.5.1 Method

In a weakly anisotropic mantle, small perturbations of the isotropic Pn-wave speed,  $\alpha_0$ , can be approximated with a fourth-degree trigonometric polynomial in  $\varphi$ , the Pn-wave propagation azimuth. The Fourier series of this polynomial is (Backus, 1965):

$$\alpha^{2}(\varphi) - \alpha_{0}^{2} = A + C\cos(2\varphi) + D\sin(2\varphi) + E\cos(4\varphi) + F\sin(4\varphi), \qquad (2.1)$$

where  $\alpha(\varphi)$  is the Pn speed for the propagation azimuth  $\varphi$ . A minimum of five wavespeed measurements along non-colinear profiles is necessary to solve for the five parameters A to F (Backus, 1965). Based on the  $2\varphi$  dependence displayed by real wavespeed measurements, however, Smith and Ekström (1999) proposed an approximation of Equation (2.1):

$$\alpha(\varphi) = \alpha_0 + C\cos(2\varphi) + D\sin(2\varphi), \qquad (2.2)$$

Solving for the three unknown parameters,  $\alpha_0$ , the isotropic Pn speed, and both constants, C and D of Equation (2.2), requires only three known wave speeds along intersecting profiles. This study's and the intersecting SIGHT Transect 2's (Scherwath *et al.*, 2003) Pn-speed estimates,  $\alpha_1$  and  $\alpha_2$ , respectively, provide two equations. A third equation,  $\frac{d\alpha(\varphi)}{d\varphi}|_{\varphi=\Phi} = 0$ , is found by assuming  $\alpha(\varphi)$  is maximum for the fast propagation azimuth,  $\Phi$ , from a nearby SKS-splitting fast polarisation orientation. The resulting system of three equations is:

$$\alpha_{1} = \alpha_{0} + C\cos(2\varphi_{1}) + D\sin(2\varphi_{1})$$

$$\alpha_{2} = \alpha_{0} + C\cos(2\varphi_{2}) + D\sin(2\varphi_{2})$$

$$0 = C\sin(2\Phi) - D\cos(2\Phi)$$
(2.3)

with solutions:

$$\alpha_{0} = \alpha_{1} - Ccos(2\varphi_{1}) - Dsin(\varphi_{1})$$

$$C = \frac{\alpha_{1} - \alpha_{2}}{G}cos(2\Phi)$$

$$D = \frac{\alpha_{1} - \alpha_{2}}{G}sin(2\Phi)$$
(2.4)

where

$$G = \cos(2\Phi) \left[ \cos(2\varphi_1) - \cos(2\varphi_2) \right] + \sin(2\Phi) \left[ \sin(2\varphi_1) - \sin(2\varphi_2) \right].$$

### 2.5.2 Pn anisotropy at the intersection with SIGHT Transect 2

Taking this study's Pn speed of 8.54  $\pm$  0.20 km/s and that of 8.0  $\pm$  0.2 km/s on the nearly perpendicular profile SIGHT Transect 2 (Scherwath *et al.*, 2003), implies 6.5  $\pm$  3.5 % apparent anisotropy (T in Fig. 2.4). A maximum possible anisotropy can be estimated using a nearby SKS fast polarisation measurement and Equations (2.5). A nearby SKS fast polarisation orientation (Klosko *et al.*, 1999) is, however, located at the transition between two domains of anisotropy approximately coinciding with SIGHT Transect 2: 1) a central South Island domain, where SKS fast polarisation orientations,  $\Phi$ , are sub-parallel to the Alpine Fault; 2) a southern South Island domain, where these are consistently ~35° counter-clockwise from the Alpine fault, i.e. the orientation of simple shear in the mantle deforming in a ductile fashion (Fig. 2.4; Klosko *et al.*, 1999; Molnar *et al.*, 1999). This Pn anisotropy measurement may, thus, be an average resulting from the overlap of Fresnel zones over the two domains. Therefore, two possible fast orientations need to be considered. SKS fast polarisation orientations in the central and southern South Island (Fig. 2.4; Klosko *et al.*, 1999) respectively average to 56°  $\pm$  2° and  $\Phi$  of 21°  $\pm$  1°. Taking the central South Island  $\Phi$  of 56°  $\pm$  2° implies a maximum



Figure 2.4: This map summarises anisotropy measurements in South Island. *Thin lines* denote previous SIGHT seismic profiles T1, T2, 3W and 4E (Okaya *et al.*, 2002) and this study's Fiordland-Cheviot refraction profile. The *open arrow* indicates the Australian-Pacific relative plate motion (after DeMets *et al.*, 1994). Existing anisotropy measurements include SKS-splitting (*orange bars* with lengths proportional to delay times and orientations parallel to fast polarisation orientations; Duclos *et al.*, 2005; Klosko *et al.*, 1999) and Pn anisotropy (*double arrows*; apparent values in italic and absolute values are in lower right corner). E1, E2 (*in green*) are absolute Pn anisotropy measurements from a global earthquake study (Smith and Ekström, 1999). The double arrows are orientated along the fast propagation orientation.

Pn anisotropy  $\delta P$  of  $7 \pm 3.5$  % and an isotropic Pn speed  $\alpha_0$  of  $8.25 \pm 0.22$  km/s, while taking southern South Island mean  $\Phi$  of  $21^{\circ} \pm 1^{\circ}$  implies a  $\delta P$  of  $13.3 \pm 3.5$  % and an  $\alpha_0$  of  $8.42 \pm 0.22$  km/s (Tab. 2.1). A mean  $\alpha_0$  of  $8.3 \pm 0.3$  km/s fits both results and is consistent with Haines' (1979) average  $8.3 \pm 0.1$  km/s for southern South Island and wave-speed perturbations of  $\sim 2$  % relative to a world-wide average of 8.1 km/s (IASP91; Kennett and Engdahl, 1991) imaged by inversion of teleseismic travel times (Kohler and Eberhart-Phillips, 2002). The 8.3 km/s Pn speed is 2–3 % more than 8.1 km/s and suggests cold and dense upper mantle material.

If the 7 % or 13 % anisotropy at the intersection with SIGHT Transect 2 is constant throughout the mantle lid, then an anisotropic layer of about 100 km or 50 km thickness, respectively, would account for the observed SKS-delay time of 1.76 s (Klosko *et al.*, 1999), assuming a P- to S-anisotropy ratio of 1.4 and a 4.7 km/s average S-wave speed in the uppermost mantle (Ben Ismael and Mainprice, 1998).

Dynamic slip alone can not explain in situ anisotropy greater than a theoretical maximum of 10 % (Ribe, 1992) as calculated for southern South Island, but requires additional dynamic recrystallization by subgrain rotation and grain-boundary migration (Karato, 1988; Nicolas *et al.*, 1973), additional pure shear or infinite strain. All these processes have the effect of rotating fast propagation orientations parallel to the shear orientation, i.e. reducing the obliquity of fast orientations to that of shear. How-

Figure 2.4 continued: S (Scherwath *et al.*, 2002), B1 and B2 (*in blue*) (Baldock and Stern, 2009, in prep.) and T (this study as *crossing red arrows with outline*) are apparent anisotropy measurements at intersecting profiles. Intersecting double arrows indicate the apparent fast and slow propagation orientations. The arrow length in the fast orientation is scaled to the percentage of anisotropy.

ever, SKS fast polarisations of southern South Island are  $\sim 28^{\circ}$  oblique to the Alpine Fault and the shear orientation, suggesting that olivine a-axes haven't aligned with the shear orientation yet, and that anisotropy should be less than the theoretical maximum of  $\sim 10$  %. Hence, a  $\sim 13$  % Pn anisotropy as calculated for the southern South Island fast orientation seems incompatible with the obliquity of fast polarisation orientations. Conversely, for central South Island, where fast polarisations are oriented parallel to the shear orientation and suggest infinite strain, pure shear and/or dynamic recrystallisation, the Pn anisotropy may be expected to be greater than 10 % and to be more than the 7 % calculated in this study. An amount of anisotropy intermediate to 7 % and 13 % or rotation of material independent of strain would resolve this paradox.

#### 2.5.3 Comparison with previous Pn anisotropy measurements

Three other Pn anisotropy measurements have been made on crossing refraction lines (Baldock and Stern, 2009, in prep.; Scherwath *et al.*, 2002). The Pn anisotropy is  $11.5 \pm 2.0 \%$  on the Australian side, 30 km west of the surface trace of the Alpine Fault (S in Fig. 1; Scherwath *et al.*, 2002). If the dip of the Alpine Fault is taken as  $40^{\circ}$  SE (Kleffmann *et al.*, 1998), then at the Moho, the measurement on the Australian side is at a similar distance to the Alpine Fault as this study's measurement of 7–13 % anisotropy on the Pacific side. Offshore and 230 km east of the Alpine Fault, two null Pn anisotropy measurements on crossing lines SIGHT Transect 1 and Transect 3 and SIGHT Transect 2 and Transect 3 (B1 and B2 in Fig. 2.4) show that upper mantle anisotropy does not extend 50 km east of South Island (Baldock and Stern, 2009, in prep.). The Pn speed is  $8.1 \pm 0.1$  km/s in both transect azimuths and can be assumed as the isotropic Pn speed. Beneath the Canterbury plains (east of the Southern Alps) into the offshore, however, northwest-southeast raypaths define a broad region of 7.8  $\pm$  0.1 km/s Pn speed (Baldock, 2004a). Assuming 7.8 km/s and 8.1 km/s are the

minimum and isotropic Pn speeds, respectively, the Pn anisotropy beneath the Canterbury plains is  $7.5 \pm 3.0$  % (Baldock and Stern, 2009, in prep.).

These measurements suggest that the Pn anisotropy is strong up to a 70–80 km distance from the Alpine Fault at depth. Scherwath *et al.* (2002) noted that the  $11.5 \pm 2.0$  % Pn anisotropy (S in Fig. 2.4) is slightly greater than the theoretical maximum of ~10 % for strain-induced anisotropy (Ribe, 1992). They suggested dynamic recrystallization, a pure shear component and/or infinite strain as possible mechanisms to explain the high observed anisotropy. In the east of South Island, the Pn anisotropy is less strong, and possibly extends as far as the east coast, about 150 km east from the Alpine Fault at Moho depth.

Table 2.1: Pn anisotropy and parameters  $\alpha_0$ , C and D of Equation (2.2) are estimated at the intersection of the Fiordland-Cheviot refraction profile with SIGHT lines T1 and T2. The Pn speed is 7.9  $\pm$  0.2 km/s and 8.0  $\pm$  0.2 km/s along SIGHT T1 and T2, respectively (Scherwath *et al.*, 2003; Van Avendonk *et al.*, 2004). Expressions for  $\sigma_{\alpha_0}$ ,  $\sigma_C$  and  $\sigma_D$  are detailed in Appendix B.

$\perp$	$\Phi$	$lpha_0$	$\sigma_{lpha_0}$	C	$\sigma_C$	D	$\sigma_D$	$\delta P$	$\sigma(\delta P)$
	$(^{\circ})$	$(\rm km/s)$	$(\rm km/s)$	$(\rm km/s)$	$(\rm km/s)$	$(\rm km/s)$	$(\rm km/s)$	(%)	(%)
Τ2	56	8.25	0.22	-0.11	0.05	0.27	0.12	7.0	3.5
T2	21	8.42	0.22	0.42	0.19	0.38	0.17	13.3	3.5
T1	56	8.21	0.24	-0.12	0.07	0.31	0.15	8.1	3.5

## 2.6 Gravity modelling

Intra-continental subduction and continuous thickening both involve displacement of asthenosphere with colder mantle lithosphere and, therefore, a downwarp of the isotherms. Thus, in both models, the negative temperature contrasts and resultant thermal contraction within the upper mantle produce positive density contrasts, which in turn produce positive gravity anomalies.

The Southern Alps region exhibits a negative Bouguer gravity anomaly (Fig. 2.5b), as is usually observed above crustal roots that sustain the load of mountain ranges (Airy isostasy). However, a closer look at the Southern Alps shows that the topography (Fig. 2.5a) and the Bouguer anomaly (Fig. 2.5b) do not correlate well (Woodward, 1979) and trend at different angles with a  $\sim 15^{\circ}$  difference. Moreover, the mean elevations of ca. 1000 m, as seen in the Wanaka region (Fig. 2.3), should only require the support of a ca. 6–9 km thick crustal root beneath, if Airy load compensation (e.g. Watts, 2001) and a -300 to -400 kg/m<sup>3</sup> density contrast between crustal root and mantle are assumed. However, as discussed in Section 2.4 the crustal root is  $18 \pm 4$  km at Wanaka (relative to a 30 km coastal average in South Island). Therefore, at Wanaka the crustal root is at least two times thicker than needed to support the topography. The anomalous gravity effect of the over-thickened crust, i.e. the deviation from Airy load compensation, is visible in the negative isostatic anomaly of the Southern Alps region (Fig. 2.5c). Below, a mass excess is assumed to exist in the mantle that maintains isostatic equilibrium by balancing part of the mass deficit of the crustal root and pulls the crustal root down. The positive gravity effect of such a mantle body, herein called the mantle residual anomaly, is obscured by the large negative anomaly of the crustal root. As a result, the observed Bouguer anomaly low is -80 mGal near Wanaka township (Fig. 2.3, 2.5b; Reilly and Whiteford, 1979) and less than that expected for an 18 km thick crustal root alone.



Figure 2.5: a) Topography of South Island, SIGHT T2 and Jackson Bay-Dunedin profile (JB-D). Highest elevations (Mt Cook region) appear in yellow. b) Bouguer gravity anomaly (adapted from Scherwath, 2002). *Rectangles*: domain and axes of the gravity model used in  $2^{\frac{3}{4}}$ D modelling and extent of model cross-sections taken along SIGHT T2 and profile JB-D (Fig. 2.6a); *hatched rectangle*: mantle body (model 1 of Tab. 2.2). c) The isostatic gravity anomaly onshore (DSIR; Reilly *et al.*, 1977) is superposed on the free air anomaly offshore (Seasat data, 1999).

The present modelling aims at defining the minimum density contrast and the lateral and vertical extent of the mantle body that fits the Bouguer gravity onshore (Fig. 2.5b). Although available offshore, the free-air gravity isn't included, because the edge effect of the continental shelf there is predominant over the long wavelength effect of the putative mantle body (Fig. 2.5c).

#### 2.6.1 Model domain

The mantle residual anomaly is estimated along a profile crossing the South Island at Wanaka, herein called Jackson Bay-Dunedin profile (JB-D in Fig. 2.5b and 5). A  $2^{\frac{3}{4}}$ D gravity modelling software (GM-SYS<sup>TM</sup>) is used that allows bodies of finite extent and non-orthogonal strike in the dimension perpendicular (Y-axis) to the calculated gravity profile (X-axis, Fig. 2.5b–c). The chosen gravity model is ca. 400 km long in the orientation parallel to the JB-D interpretation profile (X-axis in Fig. 2.5b) and extends 200 km northeast and 100 km southwest from profile JB-D (Y-axis in Fig. 2.5b). The JB-D line is well to the north of the Australian slab subducting beneath Fiordland. The gravity effect of the subducting slab is, therefore, neglected in the modelling that follows.

#### 2.6.2 Crustal model

Modelling is done relative to a reference crust of 30 km thickness (Godfrey *et al.*, 2001; Melhuish *et al.*, 2005) above a mantle of 3300 kg/m<sup>3</sup> density. An average density contrast of  $-300 \text{ kg/m}^3$  is adopted for the crustal root. This value is less than the  $-450 \text{ kg/m}^3$  (Stern *et al.*, 2000) density contrast estimated at SIGHT Transect 2. As discussed below, using a density contrast of  $-450 \text{ kg/m}^3$  in this study results in



(a) Jackson Bay-Dunedin

Figure 2.6: Cross-sections through gravity model 1 (Tab. 2.2). The thin oceanic crust, the crustal root and the mantle body are represented with their density contrasts determined relative to a reference crust of 30 km thickness and an average mid-lower crustal density of 3000 kg/m<sup>3</sup> above a mantle of 3300 kg/m<sup>3</sup> density. a) Mean topography in a 10 km wide band, Bouguer gravity anomaly (Reilly and Whiteford, 1979) and X-cross section (Y=0 km) are taken along the Jackson Bay-Dunedin profile (JB-D). b) The Y-cross section (X=100 km in Fig. 2.5b-c) is taken perpendicular to profile JB-D. JB-D and T2 denote the intersections with crossing profiles.

an unreasonably large density contrast within the upper mantle. The SIGHT Transect 2 mostly traverses greywacke, apart from a ~60 km wide strip of Alpine Schist directly south east of the Alpine Fault. In the mid and lower crust, rocks are inferred to be greywacke/schist (Scherwath *et al.*, 2003) and oceanic crust (Kleffmann, 1999), respectively. In contrast, the JB-D line is substantially within the Otago Haast Schist (greenschist facies). These schists represent the deeply exhumed part of a Mesozoic accretionary prism on the margins of Gondwana (e.g. Mortimer, 2004). The JB-D line strikes parallel to the axis of an antiform that corresponds to the largest amount of exhumation within the Otago Schist possibly reaching as much as 10–25 km. Thus, it is possible that along the line JB-D rocks at a present depth of 30 km were once possibly ~50 km deep. At these depths and temperatures, continental crust starts to transform to eclogite (e.g. Wyllie, 1992) of ~3550 kg/m<sup>3</sup> density (Hacker and Abers, 2004). If the lower crust were to be partially transformed to eclogite, the across-Moho density contrast would be low or even absent (see also Section 4.3 on Otago Schists).

The crustal root is 14 km thick at SIGHT Transect 2 (Y=-140 km) and thickens to 18 km ~30 km along strike to the southwest of SIGHT Transect 2 (Y=-100 km in Fig. 2.6b). The crustal structure along profile JB-D is not known in detail and is constrained by only three points of known crustal thickness. The crustal thickness is ca. 30 km off-shore Jackson Bay (Melhuish *et al.*, 2005),  $48 \pm 4$  km thick near Wanaka (this study) and ranges between 27 and 33 km off-shore from Dunedin (Godfrey *et al.*, 2001). In between these three points the shape of the crustal root and the location of its deepest point are not well constrained. The simplest hypothesis is that the crustal root is asymmetric as imaged along SIGHT Transect 1 (Van Avendonk *et al.*, 2004) and T2 (Scherwath *et al.*, 2003). However, geodetic strain-rates (Henderson, 2003) and Holocene reverse faulting show that contraction occurs as far as eastern Otago (Fig. 2.5a) in southern South Island, and is here more distributed than in central South Island (Norris and Cooper, 2001). Three-dimensional crustal structure obtained from simultaneous inversion of earthquake and shot arrival times and gravity data included below 20 km (Eberhart-Phillips and Bannister, 2002, their Fig. 12) also suggests a wider crustal root in south than in north. Hence, the crustal root may be distributed further south east from the Alpine Fault along profile JB-D than it is in the north (SIGHT Transect 2). The crustal root is therefore assumed symmetric with respect to the gravity minimum.

The maximum crustal thickness is fixed to 48 km at the intersection with the Fiordland-Cheviot profile (X=100 km of profile JB-D) and the crustal root is symmetric with respect to the gravity minimum (X = 123 km in Fig. 2.5b and 5a).

Due to the strong trade-off between shape of the crustal root and symmetry of the mantle body, the mantle residual anomaly, i.e. the difference between the Bouguer anomaly and the modelled crustal root gravity effect, is symmetric (Fig. 2.7) and requires the presence of a symmetric positive mantle mass excess. In contrast, an asymmetric crustal root would require an asymmetric, i.e. dipping, mantle body. In other words, reasonable constraints can be placed on the mass excess of the mantle body, but not its shape.

#### 2.6.3 Models for the mantle body

Assuming the mantle mass excess is a cylindrical-type source, a first-order maximum depth of 90–100 km is estimated for the centre of mass from the half-maximum of the mantle gravity anomaly (Nettleton, 1976). The mass per unit length of strike of the mantle body is ca.  $2.7 \ge 10^{11} \text{ kg/m}$  as determined by the mass balance between topography and crustal root.



Figure 2.7: Bouguer gravity anomaly (Reilly and Whiteford, 1979) and gravity anomalies calculated for model 1 (Tab. 2.2, Fig. 2.6) using GM-SYS<sup>TM</sup>, a  $2^{\frac{3}{4}}D$  gravity modelling software. The curves represent: the anomaly for a symmetric crustal root alone; the anomaly for the entire model (crustal root + mantle body); the mantle residual anomaly (Bouguer – crustal root); and the total misfit (Bouguer anomaly – entire model).

Thus, although the density contrast, width and depth extent of the mantle body are free parameters, the mass excess is required to attain the above mass per unit length of 2.7 x  $10^{11}$  kg/m<sup>3</sup> within 10 % (Tab. 2.2). In addition, the Moho depth, i.e. the minimum depth that the top of the mass excess can reach, and the first-order depth estimate of the centre of mass, provide bounds to the vertical extent of the mantle body. A minimum density contrast is found, for which the body's vertical dimension is maximum but contained within the vertical bounds mentioned above, and the resulting gravity effect satisfies the amplitude and wavelength of the Bouguer anomaly. For density contrasts smaller than this minimum, vertical stretch of the mantle body is necessary in order to fit the maximum amplitude of the gravity anomaly. However, because the gravity effect is proportional to  $r^{-2}$ , deeper mass is less effective in producing a gravity effect and more mass needs to be added than required to attain mass balance. Hence, there is no body found with density contrast below this minimum that can fulfill the mass balance requirement as well.

A minimum density contrast,  $\Delta \rho$ , of  $35 \pm 5 \text{ kg/m}^3$  is required for a mantle body centred at 90–100 km depth in order to satisfy the mass balance and the wavelength of the gravity anomaly with a misfit of the order of 10 mGal west of kilometer 230 (Fig. 2.7; model 1 of Tab. 2.2 and Fig. 2.6). The misfit increases east from kilometer 230, possibly where the edge effect of the continental shelf starts to be seen. The lateral and vertical dimensions of this body are  $110 \pm 20$  km and  $70 \pm 20$  km respectively. The minimum density contrast ( $\Delta \rho$ ) is 20 ± 5 kg/m<sup>3</sup> for an across-Moho density contrast of  $-250 \text{ kg/m}^3$  (model 2 of Tab. 2.2) or  $55 \pm 5 \text{ kg/m}^3$  for the across-Moho density contrast of  $-350 \text{ kg/m}^3$  (model 3 of Tab. 2.2). The minimum density contrast is even larger,  $\Delta \rho \sim 120 \text{ kg/m}^3$ , for an across-Moho contrast of  $-450 \text{ kg/m}^3$ , as assumed under SIGHT Transect 2 (Stern et al., 2000). However, such a density contrast is far beyond the average of 60 kg/m<sup>3</sup> (e.g. Houseman *et al.*, 2000) for an approximate  $500^{\circ}C$  temperature contrast and a  $\sim 60$  km vertical deflection of isotherms alone that can be considered as a reasonable maximum for South Island (Fig. 1.5). Further chemical heterogeneity between mantle lithosphere and asthenosphere or the presence of eclogitic rocks would be required within the mantle if the density contrast were to be that large.

In summary, the mantle body is wider and less thick than previously inferred along the SIGHT Transect 2 line (model 0, Tab. 2.2; Stern *et al.*, 2000), but provides a similar mass excess in the case of a crustal root with  $-300 \text{ kg/m}^3$  density contrast. Table 2.2: Gravity models are derived for an 18 km thick crustal root below a 30 km deep Moho, symmetric to X=123 km and maximum between X=100 and X= 146km of profile JB-D (Fig. 2.5b-c and 5a). The crustal root is thickened at Y=-100 km from 14 km at SIGHT Transect 2 to 18 km at JB-D (Fig. 2.6b). A crustal density of 2700 kg/m<sup>3</sup> is adopted for the topography. Density contrasts,  $\Delta \rho$ , in the range -250 to -350 kg/m<sup>3</sup> are adopted for the crustal root. The minimum density contrast and dimensions of a mantle anomaly are varied with the requirement to fit the Bouguer anomaly (Reilly and Whiteford, 1979)[; Fig. 2.7] and to attain the mass balance between topography, crustal root and mantle body within 10 %:  $\delta m$ (mantle) + $\delta m$ (topography)  $\approx -\delta m$ (crustal root).  $X_1 - X_2$  is the lateral extent of the mantle body,  $Z_1 - Z_2$ , the depth range (Fig. 2.6a),  $A = (X_2 - X_1)(Z_2 - Z_1)$ , the cross-section and  $\delta m = A \cdot \Delta \rho$ , the linear mass excess of the mantle body. The temperature contrast,  $\Delta T$ , is calculated by assuming that the density contrast  $\Delta \rho$  is solely due to thermal contrast and taking  $\Delta \rho = -\rho \alpha \Delta T$  with  $\alpha = 3.5 \times 10^{-5}$  (Anderson *et al.*, 1992) the coefficient of thermal expansion and  $\rho = 3300 \text{ kg/m}^3$  the uppermost mantle density.

model	topography	root		mantle body							
	$\delta m$	$\Delta \rho$	$\delta m$	$\Delta \rho$	$X_{1m} - X_{2m}$	$Z_1 - Z_2$	A	$\delta m$	$\Delta T$		
	(kg/m)	$(kg/m^3)$	(kg/m)	$(kg/m^3)$	$(\mathrm{km})$	(km)	$(m^2)$	(kg/m)	$(^{\circ}C)$		
1	$2.4 \ge 10^{11}$	-300	$-5.1 \ge 10^{11}$	+35	60 - 175	60 - 130	$8.0 \ge 10^9$	$2.8 \ge 10^{11}$	305		
2	$2.4 \ge 10^{11}$	-250	$-4.1 \ \mathrm{x} \ 10^{11}$	+20	50 - 200	75 - 125	$7.5 \ge 10^9$	$1.5 \ge 10^{11}$	170		
3	$2.4 \ge 10^{11}$	-350	$-5.9 \ge 10^{11}$	+55	60–170	60 - 125	$7.1 \ge 10^9$	$3.9 \ge 10^{11}$	480		
mantle body derived at SIGHT Transect 2 (Stern <i>et al.</i> , 2000)											
0	$3.4 \ge 10^{11}$	-450	$-5.5 \ \mathrm{x} \ 10^{11}$	+30	70–160	70 - 170	$9.0 \ge 10^9$	$2.7 \ge 10^{11}$	260		

# 2.7 Discussion

The  $8.54 \pm 0.20$  km/s Pn speed estimated at 5° clockwise from the strike of the Alpine Fault is interpreted to be the result of seismic anisotropy (Scherwath, 2002) and a high average wave speed in the mantle lid. Seismic anisotropy is recognised to be mostly the product of 850 km shear between the Pacific and the Australian plates in the past 45 Myr (Baldock and Stern, 2009, in prep.; Little *et al.*, 2002b; Molnar *et al.*, 1999; Savage *et al.*, 2004). Pn speeds higher than the worldwide average upper mantle wave speed of 8.1 km/s (IASP91; Kennett and Engdahl, 1991) are indicative of shortened and cold mantle beneath the Southern Alps after 20 Myr of convergence (Cande and Stock, 2004; Walcott, 1998), as discussed below.

The average Pn speed of ~8.3 km/s is a 2–3 % perturbation relative to the upper mantle wave speed of 8.1 km/s (Kennett and Engdahl, 1991). Taking  $\delta V_P/\delta T = 5 \ge 10^{-4} \text{ km/s}/^{\circ}\text{C}$ (Anderson and Isaak, 1995), the relationship between lateral wave speed variation and temperature, this wave-speed perturbation could be explained by a ~400 °C negative temperature contrast with the surrounding mantle. Similarly, the temperature contrast,  $\Delta T$ , caused by the deflection of isotherms can be estimated from the density contrast,  $\Delta \rho$ , with  $\Delta \rho = -\rho \alpha \Delta T$ . Taking  $\alpha = 3.5 \ge 10^{-5}$  (Anderson *et al.*, 1992) as the coefficient of thermal expansion and  $\rho = 3300 \text{ kg/m}^3$  for the uppermost mantle density, the equivalent average temperature contrast ranges from -170 °C to -480 °C in the case of a 20–55 kg/m<sup>3</sup> density contrast in the mantle lid, as suggested by the gravity modelling described above.

The bulk of the inferred mantle body (Fig. 2.6) compares well with a zone of fast wave speed below the Southern Alps imaged by 3D inversion of teleseismic travel-time residuals (Kohler and Eberhart-Phillips, 2002, their Fig. 7). Along profile JB-D, their inversion results display a zone of anomalous mantle with P-wave speed perturbations of 1.5–3 % relative to 8.1 km/s existing in a ca. 100 km wide zone located ca. 50 km offset east of the Alpine Fault. Similar to the results of their 3D inversion, the present modelling suggests that the anomalous mantle extends deeper north (Mt Cook region) than south (Wanaka region) (compare models 1 and 0 of Tab. 2.2).

Crustal roots of  $14 \pm 2$  km thickness in central South Island (Scherwath *et al.*, 2003) and  $18 \pm 4$  km thickness in southern South Island (relative to a coastal average of 30 km) as well as mantle mass of similar excess beneath both regions seem, at first, counter-intuitive with the total convergence across the Alpine Fault being ca. 40 km less (Cande and Stock, 2004) and elevations ca. 500 m less across southern than central South Island. Lower crustal extrusion (Bird, 1991) in an oblique convergent setting has been suggested as a possible mechanism for maximum crustal thickening south east of the Southern Alps topographic maximum and at ca.  $15^{\circ}$  counter-clockwise from the Alpine Fault (Gerbault *et al.*, 2002). Although lower crustal extrusion is a possible explanation for the large crustal thickness beneath Wanaka, a further process is required that thickens the mantle lithosphere and provides the mass excess to fit the gravity anomaly beneath the Wanaka region.

Two observations suggest that the nearby Puysegur margin may contribute to thickening of the Pacific lithosphere of the southwestern South Island. First, hypocentres image north-eastward steepening of the Benioff zone in the Australian slab beneath Fiordland (white line in Fig. 2.8 after Reyners *et al.*, 2002; Smith and Davey, 1984). Smith and Davey (1984) proposed a model, in which steepening of the slab results from the northeast section of the Australian slab breaking while being obliquely subducted beneath the Pacific lithosphere. In this model, the broken section has rotated about a horizontal axis into a more upward position and now resembles a ploughshare.



Figure 2.8: *Green arrows* represent the projections of the 34 mm/yr Australian-Pacific relative plate motion vector (*blue arrows*) perpendicular to the top of the Australian slab (*white line as interpreted by* Reyners *et al.*, 2002) and the Alpine Fault. These overlay a 100 km depth slice of P-wave speeds beneath Fiordland (Eberhart-Phillips and Reyners, 2001, their Fig. 6d). The black line denotes the Fiordland Cheviot refraction profile.

The raised top edge of the rotated slab section provides the explanation for the dynamic uplift of the overriding Fiordland block. Second, 3D inversion of local-earthquake data (Eberhart-Phillips and Reyners, 2001) indicates a zone of high mantle  $V_P$  ( $V_P >$ 8.5 km/s, i.e. to 2–3 % faster  $V_P$  than the surrounding) beneath Fiordland that is east of and adjacent to the Australian slab (Fig. 2.8). This high  $V_P$  zone shallows from 90 km depth beneath Fiordland to 60 km depth beneath the Southern Alps (Eberhart-Phillips and Reyners, 2001), while the eastern extent is unresolved. This zone of high wave speed could, thus, represent the southernmost expression of the thickened Pacific



Figure 2.9: Interpretative figure of the lithospheric structure along the Fiordland-Cheviot profile. *Top*: The *thin curve* is the mean topography in a 10 km wide swath along the Fiordland-Cheviot profile; and the *thick curve* is the Bouguer anomaly (Reilly and Whiteford, 1979) (Fig. 2.1). *Bottom*: real scale 2D model of the Fiordland-Cheviot profile, seismicity within a 10 km wide swath and interpretation.

mantle lithosphere within the South Island plate boundary. Here, the Australian slab may act as a rigid backstop, called a buttress by Malservisi *et al.* (2003), which contributes to thickening of the Pacific mantle lithosphere from the southwest. Projecting the  $\sim$ 34 mm/yr relative plate motion at the latitude of Fiordland (DeMets *et al.*, 1994) onto the Australian slab (Reyners *et al.*, 2002) results in a convergence rate as large as  $\sim$ 26 mm/yr, i.e. 6 times greater than the convergence rate perpendicular to the Alpine Fault at Jackson Bay (compare blue arrows in Fig. 2.8; also Fig. 2.9). Alternatively, Malservisi *et al.* (2003) interpret a backstop wider than the slab inferred from seismicity (Reyners *et al.*, 2002) onto which the Pacific mantle collides at the almost full plate speed of  $\sim$ 34 mm/yr. Assuming that the Pacific mantle has been converging for 10–20 Myr at a rate of  $\sim$ 26 mm/yr onto the Australian slab, then the total shortening across the margin would be as large as 250-480 km. The length of the Australian slab Benioff zone implies that at least 150 km of the total shortening must have been accommodated as subduction, while the 100-330 km remainder may have been accommodated by thickening of the Pacific mantle lithosphere. Hence, in the southern South Island, shortening of the mantle may occur both at a slow convergence rate oriented perpendicular to the Alpine Fault that resulted in ~40 km of total convergence (Cande and Stock, 2004), but also at a faster rate oriented perpendicular to the Australian slab. As a result, the thickened mantle lithosphere is an effective load that pulls down and thickens the overlying crust.

# 2.8 Conclusions

This earthquake refraction study offers new constraints on the uppermost mantle properties beneath the Southern Alps, in a direction almost parallel to the Australian-Pacific plate boundary and perpendicular to former crustal studies across the Southern Alps.

- The average Pn speed along the N60°E profile is 8.54 ± 0.20 km/s and the Moho is dipping at an apparent angle of 2.5° ± 1.3° SW. A 48 ± 4 km crustal thickness is estimated near Wanaka. At 80 km east of the Alpine Fault but 40–55 km east at Moho depth, the Pn anisotropy is in the range 7–13 % and the isotropic Pn speed is 8.3 ± 0.3 km/s.
- 2) The Southern Alps crustal root near Wanaka is  $18 \pm 4$  km thick (relative to a coastal average of 30 km in South Island) and is twice that required by Airy isostasy for a crustal root of  $-300 \text{ kg/m}^3$  density contrast.
- 3) Mass balance predicts the presence of a mantle mass excess per unit strike length of 2.7 x 10<sup>11</sup> kg/m beneath the southern Southern Alps (Jackson Bay-Dunedin profile), for an 18 km thick crustal root of assumed -300 kg/m<sup>3</sup> density contrast

with the lithospheric mantle. This mantle mass excess is approximately the same across central South Island (SIGHT T2; Stern *et al.*, 2000). The contrast would, however, be greater for larger across-Moho density contrasts, e.g. 40 % greater for a crustal root of  $-350 \text{ kg/m}^3$ .

- 4) The mantle body has a positive density contrast of  $35 \pm 5 \text{ kg/m}^3$  minimum with  $110 \pm 20 \text{ km}$  width and  $70 \pm 20 \text{ km}$  thickness for an across-Moho density contrast of  $-300 \text{ kg/m}^3$ .
- 5) For crustal roots with density contrasts of -400 kg/m<sup>3</sup> and more, the minimum density contrast required for the mantle body is larger than can be explained by the downwarp of isotherms alone.
- 6) The Puysegur margin, located southwest of the Southern Alps collision zone, may contribute to thickening of the Pacific mantle lithosphere beneath southern South Island by its subducted slab acting as a rigid backstop. This thickened mantle lithosphere is an effective load that pulls down the overlying crust.
- 7) The present gravity modelling is limited by the lack of constraints on the crustal structure of southern South Island. Here crustal and mantle investigations are needed in order to model the gravity effect of the crustal root more precisely, and constrain the bulk and geometry of the mantle body.

# Chapter 3

# Synthetic P and S receiver functions

# 3.1 Introduction

While in P receiver functions crustal reverberations can provide a constraint on crustal thickness and the Vp/Vs ratio, they present the disadvantage of interfering with direct Ps conversions. These reverberations may be strong, especially in tectonically active regions such as the Southern Alps. Reverberations can, thus, render the interpretation of P receiver functions difficult. Such interference, in contrast, does not occur in S and SKS receiver functions. Advantage can be taken of this property to discriminate reverberations from direct conversions within the P receiver functions. S and SKS receiver functions are directly comparable with P receiver functions in the case of flat isotropic layers. However, in the presence of dipping layers and anisotropy, especially, polarities of S and P receiver functions differ. Effects on S receiver functions of dipping and shallow or deep anisotropic layers with fast or slow symmetry axis are examined using synthetic receiver functions.

#### 3.1.1 P receiver functions

Receiver functions produce an image of the subsurface in terms of P-to-S conversions. They appear as a series of pulses, whose delay times increase with increasing conversion depth, and whose amplitudes and polarities depend on the conversion from P to S phases at the corresponding boundaries.

As the incidence of a teleseismic P wave is sub-vertical beneath a station, nearly all P particle motion is contained in the vertical component, Z. In contrast, the particle motion of the P-to-S (Ps) converted energy is mainly horizontal. Hence, the Z component,  $s_Z(r_i, t)$ , can be approximated as the convolution of the earthquake source function,  $S(r_0, t)$ , with the instrument response,  $R(r_i, t)$ . The horizontal component  $s_H(r_i, t)$ , in addition, contains all conversion information. Thus,  $s_H(r_i, t)$  is defined as the convolution of  $S(r_0, t)$  with  $R(r_i, t)$  and the Green's function,  $G(r_o, r_i, t)$ , i.e. the Earth's response to an impulse:

$$s_H(r_i, t) = S(r_0, t) * G(r_o, r_i, t) * R(r_i, t)$$

The Earth's response is isolated by deconvolving  $S(r_0, t)$  and  $R(r_i, t)$ , i.e.  $s_Z(r_i, t)$ , from  $s_H(r_i, t)$ . The remainder is a series of transmission and reflection coefficients that characterise discontinuities encountered by the P phase along the raypath.

The amount of transmission and conversion of a phase at a discontinuity is directly dependent on the phase incidence angle, the contrast in seismic properties across the discontinuity, called the impedance contrast (Zoeppritz, 1919), and the velocity gradient. In P receiver functions, the Ps pulse is typically positive as the incident P phase is transmitted from a faster to a slower medium, i.e. wave speed increasing with depth (Fig. 3.1).

In the case of flat isotropic layers, all P-to-SV converted energy remains in the source-receiver plane and is seen in the radial receiver function. In the presence of an interface dipping away from the source-receiver plane, however, energy is deflected away and onto the transverse component (Cassidy, 1992; Savage, 1998). Similarly, in the presence of anisotropy the P particle motion is converted into a quasi-P with linear particle motion not quite parallel to the propagation direction and the S wave splits into a quasi-S1 and a quasi-S2 with particle motions parallel and perpendicular to the fast polarisation orientation for the propagation direction in question (Savage, 1998). As a result, each quasi-S particle motion has a radial as well as a transverse component that produce pulses on both radial and transverse components of the receiver functions.

#### 3.1.2 S receiver functions

Similar to P waves converting to S waves, S and SKS phases experience conversions into P. In contrast to P and Ps, the S and Sp particle motions are mostly horizontal and vertical, respectively. Hence, Sp conversions and information on the subsurface can be isolated by deconvolving horizontal components from the vertical (Bock and Kind, 1991; Faber and Mueller, 1980; Farra and Vinnik, 2000). Similar to a direct P, the SKS phase is polarised in the source-receiver plane and energy on the transverse SKS receiver function results from anisotropy and dipping layers. In contrast, the polarisation of the direct S wave depends on the focal mechanism and has variable proportions of radial and transverse components. Hence, the resulting receiver function will vary from earthquake to earthquake. As a result, S receiver functions appear less coherent, and energy on the transverse component can not be attributed to anisotropy or dipping layers only.

Free-surface multiples of the direct S, like those of the P, arrive after the direct S (see Figure 3.2). In contrast to Ps conversions that follow the direct P however,



Figure 3.1: Top: Amplitude (solid curve) and phase (dashed curve) of transmitted P and SV in the case of an incident P (left) or SV (right) propagating from a fast into a slower medium (see model below). Amplitude and phase are calculated using Zoeppritz's equations (Zoeppritz, 1919).  $i_{Pr}$  and  $i_{Pt}$  are the reflected P and transmitted P critical angles for an incident SV, respectively. At incidence angles below the reflected P critical angle,  $i_{inc} < i_{Pr}$ , the transmitted P is delayed by 180°. It is advanced by 180° for  $i_{Pr} < i_{inc} < i_{Pt}$ . The P transmission coefficient is, hence, negative for SV incidence angles smaller than the transmitted P critical angle,  $i_{inc} < i_{Pt}$ . The transmission coefficient equals zero for  $i_{inc} > i_{Pt}$ . Bottom: Vp and Vs model used to calculate above P and SV transmission coefficients. P and S slownesses of 8  $s/^{\circ}$  and 12  $s/^{\circ}$  are typical of epicentral distances of  $\sim 45^{\circ}$  and  $\sim 65^{\circ}$ .
Sp conversions precede the direct S wave. This difference is a clear advantage of S receiver functions relative to P receiver functions as multiples do not interfere with Sp conversions.



Figure 3.2: Synthetic P and S receiver functions (*right*) for an isotropic 30-km thick crust (*left*). Note that polarities of the S receiver function are shown reversed from the true ones to make them resemble P receiver functions. Annotations P and S indicate the direct wave pulses. Ps and Sp denote conversions from the Moho. PpPs, PsPs+PpSs, SpPp and SsPp are the Moho free-surface multiples.

On the other hand, S receiver functions are noisier than P receiver functions because of interference with the P coda. Furthermore, interference of S and SKS phases within epicentral distances of  $85^{\circ}-90^{\circ}$ , restricts the use of S and SKS receiver functions to  $\Delta < 85^{\circ}$  and  $90^{\circ} < \Delta < 210^{\circ}$ , respectively (Yuan *et al.*, 2006). Corresponding S and SKS slownesses are within 10–15.5 s/° and 5–6.5 s/°, respectively. P slownesses range from 5 s/° to 9 s/° and, thus, overlap with the slowness range of SKS waves. Because of the larger S slowness, Sp conversions of S receiver functions have greater move-outs and amplitudes (Fig. 3.1) than conversions in P and SKS receiver functions (Fig. 3.3). Finally, S and SKS are lower frequency than P phases and, thus, provide less resolution than P receiver functions. The resolution is half a wavelength in transmission.

As illustrated in Figure 3.1, the Sp phase has a phase shift of  $-180^{\circ}$  at incidence angles less than the reflected P critical angle,  $i_{inc} < i_{Pr}$ . This phase shift is  $180^{\circ}$  at incidence angles within the reflected and transmitted P critical angles,  $i_{Pr} < i_{inc} < i_{Pt}$ , where there is no reflected P phase. Finally, at incidence greater than  $i_{Pt}$ , there is no transmission as a P phase. As a result, the Sp transmission coefficient is negative at incidence angles less than the transmitted P critical angle,  $i_{inc} < i_{Pt}$ , and zero above. Thus, polarities of S receiver functions are opposite to those of P receiver functions.

In order to enable comparison of time delays,  $\tau$ , and polarities of S receiver functions with those of P receiver functions, S and SKS receiver functions of the present chapter are displayed with amplitude and time scales reverse. Pulse polarities will be presented as shown in figures, i.e. reversed to the real polarity. Direct comparison of Ps and Sp polarities is possible in the case of flat isotropic layers only as shown in Figure 3.3. The effects of dipping layers and anisotropy are explored in Section 3.3.

# 3.2 Method

Synthetic SV and SH receiver functions are produced in two steps. Firstly, threecomponent synthetic seismograms are computed with a code of Frederiksen and Bostock (2000) code for an incident SV or SH phase and a 2 s pulse. This code allows one to specify models with dipping layers and anisotropy with hexagonal symmetry and a plunging symmetry axis. Secondly, the synthetic seismograms are used as input for a program of Park and Levin (2000) that computes the receiver functions. The receiver



Figure 3.3: Left: Vp and Vs distribution with depth for a standard lithosphere (Kennett and Engdahl, 1991). Right: corresponding synthetic SKS, P and S radial receiver functions as a function of slowness. Annotations C, M and L denote conversions from the top of the lower crust, the Moho and the lithosphere-asthenosphere boundary, respectively. Arrows on the right-hand side indicate the slowness range of the incident SKS, P and S waves. Positive (blue) and negative (red) pulses in the time and slowness range of 8–16 s and 6.5–9 s/° are C and M lower crust and Moho free-surface multiples PpPs (positive) and PsPs+PpSs (negative). Note the increasing move-out with increasing slowness and increasing depth to the discontinuity. Interference occurs in the P receiver functions between the negatively polarised lower crust multiples PsPs+PpSs and lithosphere-asthenosphere conversion (L). Low amplitude peaks in the S receiver functions at large slowness and at twice the C and M delay times correspond to numerical ringing.

functions are compiled in the ZRT reference frame with radial (R) and transverse (T) components in the right-handed convention, in which the transverse lags the radial by  $90^{\circ}$ .

Park and Levin's 2000 code implements the multiple-taper spectral correlation method (Kuo *et al.*, 1990; Vernon *et al.*, 1991) before calculating the deconvolution. Seismograms are multiplied with a set of orthogonal slepian tapers (Fig. 3.4) then Fourier transformed into so called spectral estimates.

Slepian tapers are designed to minimise spectral leakage, that is to maximise the ratio of spectral amplitudes between the central lobe and the sidelobes (Walden et al., 1995, e.g.). The width of the central lobe can be estimated from the minimum distance between two uncorrelated spectral elements (Walden *et al.*, 1995, e.g.). It is defined by two times the so-called bandwidth, W. The bandwidth is therefore a measure of the spectral resolution, e.g. the smaller the bandwidth the higher the spectral resolution. In slepian tapers, however, spectral resolution trades off with energy maximisation, e.g. the higher the spectral resolution the lower the energy maximisation and the stronger the spectral leakage. Such effect can be compensated by increasing the length, T, of the time series, which yields lower spectral leakage than the shorter time series. The value of the time-bandwidth product, p=TW, determines the spectral resolution and energy maximisation that can be achieved for a certain combination of T and W. For Slepian tapers, the length of the selected time window, T, and the time-bandwidth product, p, also determine the time, t=T/2p, to which a signal (e.g. Ps conversion) energy is retrieved effectively. Finally, the time-bandwidth product determines the maximum taper order beyond which maximisation of energy doesn't hold. For slepian tapers, K = 2p - 1 tapers of orders 0,... 2p - 2 should be used at the most (e.g. Walden *et al.*, 1995).

The code estimates the receiver functions in the frequency domain from the sum

of the cross-correlations of the horizontal and vertical component spectral estimates, normalised by the sum of the auto-correlations of the vertical component spectral estimates and a pre-event noise spectrum estimate for frequency-dependent damping (Eq. 3 in Park and Levin, 2000). Such multi-taper correlation has the advantage of avoiding numerical instabilities that result from zero division in the spectral domain (Ammon



Figure 3.4: *Top*: synthetic waveform (blue) and slepian tapers (red) used in the multitaper correlation (MTC) with time-bandwidth product p=2.5 and number of windows K=3. *Bottom*: tapered waveform before Fourier transformation.

*et al.*, 1990). In addition, the code is supplemented by an uncertainty estimate for the frequency-domain multi-taper receiver function (Park and Levin, 2000) that is used to weight the single receiver functions for stacking. This weighted stacking renders the code insensitive to single noisy receiver functions.

Finally, spectra are low-pass filtered within the code up to a user-defined cut-off frequency,  $f_c$ , with a cosine-squared function,  $\cos^2(\pi f/2f_c)$  (Park and Levin, 2000), to further minimise ringing in the spectrum.

Park and Levin's 2000 code was modified here for the purpose of calculating S receiver functions and stacking receiver functions as a function of slowness (herein expressed as ray parameter, p).

In the following synthetic examples, a time-bandwidth product p=2.5 and K=3 tapers have been adopted as recommended by Park and Levin (2000). Pre-event and post-event windows are 100 s and 20 s long, respectively, in order to retrieve energy up to 20 s. A cut-off frequency of 0.5 Hz is used that effectively allows energy up to 0.33 Hz. Single receiver functions are stacked in back azimuth and epicentral distance bins of  $10^{\circ}$  with increments of 5°, so that each receiver function participates in two neighbouring bins. Rays are incident from north in the epicentral distance stacks. Rays have a slowness of 13.9 s/°, equivalent to an epicentral distance of 50° or an incidence angle of ~35°, in the back azimuth stacks.

# 3.3 Synthetic SV/SH receiver functions

Effects of dipping layers and anisotropy on P receiver functions have been compared in detail by Cassidy (1992) and Savage (1998). Below, such effects are investigated for SV and SH receiver functions.

#### 3.3.1 Effect of a dipping layer

Continuity of the incident SV or SH particle displacement at the boundary but a new propagation direction will result in the transmitted S polarisation having both radial and transverse components. The radial component of particle motion allows conversion of S energy into P.

Figure 3.5: Model DIP8 (Tab. 3.1) represents an interface dipping  $8^{\circ}$  to the north. The dipping interface has the effect of deflecting the incident energy away from the original propagation plane, the vertical plane passing through source and receiver.



SV and SH receiver functions of Figure 3.6 and 3.7, respectively, are described below:

- (a) SV-to-P conversions are absent where the angle between dipping plane and incident S is greater than the transmitted P critical angle  $i_{inc} > i_{Pt}$  (Fig. 3.1;  $\Delta < 45^{\circ}$ in top of Figure 3.6-3.7 and back azimuths within  $\pm 20^{\circ}$  of the dip direction in the bottom of Figure 3.6-3.7). The absence of a direct S is interpreted as a numerical problem in Park and Levin's 2000 computer code;
- (b) energy at 0 s delay time of the component perpendicular to the incident polarisation – transverse component for an incident SV and radial component for an incident SH – results from the incident S energy being deflected away from the original propagation plane;
- (c) Sp energy is apparent on the component of the incident polarisation only radial (or transverse) in the case of an incident SV (or SH);



Figure 3.6: Effect on SV receiver functions of a layer dipping 8° to the north (model DIP8 of Tab. 3.1) as a function of epicentral distance (top) and back azimuth (bottom). *Top*: rays' azimuth is north. *Bottom*: rays' slowness is 13.9 s/°, which is equivalent to an S wave incidence angle of 35° at the Moho and an epicentral distance of ~50°. A 0.5 Hz low-pass filter is applied. Amplitudes and time scale are reversed as discussed in the text to make S receiver functions polarities and delay times comparable to P receiver functions.



Figure 3.7: Same as Figure 3.6 for and incident SH wave.

- (d) SV waves travelling updip produce Sp conversions with the largest amplitudes and delay times. Those travelling downdip produce Sp with the smallest amplitudes and shortest delay times;
- (e) SH waves travelling updip produce Sp conversions with the smallest amplitudes and largest delay times. Those travelling downdip produce Sp with the largest amplitudes and shortest delay times;
- (f) polarities of the radial component are all positive (or all negative) in the case of an incident SV (or SH);
- (g) polarities of the transverse are anti-symmetric about the dip direction, i.e. they reverse at 0° and 180° from this direction. As a result, the initial S pulse on the transverse is always negative for back azimuths east of the dip direction;
- (h) a third arrival at delay times twice those of the Sp conversions is an artifact produced by numerical ringing and is enhanced for waves travelling updip;

#### Similarities with P receiver functions:

- (a) Deflection of the incident wave energy towards the transverse plane results in pulses with 0 s delay time on both components (Cassidy, 1992);
- (b) Sp delay times in both SV and SH receiver functions are the largest for waves propagating updip (Cassidy, 1992);
- (c) Sp amplitudes in the SV receiver functions are the largest for waves propagating updip (Cassidy, 1992) but the smallest in the SH receiver functions;
- (d) polarities of the transverse component are anti-symmetric about the dip direction (Cassidy, 1992; Savage, 1998);
- (e) similar to P receiver functions (Savage *et al.*, 2007a), the direct S pulse has negative polarity for updip-travelling waves with SKS incidence impinging on



Figure 3.8: Schematic representation of the polarity reversal of the direct S pulse in the radial SV receiver function that can occur in the case of an updip-travelling SKS but not of a shallow incident S (modified from Savage *et al.*, 2007a, for an incident P).

a steeply-dipping interface with a strong contrast at the top (not shown in the synthetic receiver functions but in Figure 3.8).

#### Differences with P receiver functions:

- (a) absence of conversions where  $i_{inc} > i_{Pt}$ ;
- (b) an Sp pulse exists for the component that contains the incident S polarisation, only. In contrast, the Ps pulse exists for both components of the P receiver functions (Cassidy, 1992; Savage, 1998).

#### 3.3.2 Effect of crustal anisotropy

Primary causes of crustal anisotropy are rock foliation and aligned cracks (Babuska and Cara, 1991; Crampin, 1978). These are typically modelled with hexagonal symmetry and a symmetry axis along the slow propagation orientation. Figure 3.9: Model ANISO1 (Tab 3.1) consists of a top anisotropic layer overlying an isotropic halfspace. The anisotropy is defined by a slow symmetry axis dipping to the north, that is the wave speed is 5 % lower in the axis direction than in the plane perpendicular to it.



As the incident SV or SH enters the anisotropic layer, this is split into two quasi-S waves, qS1 and qS2, whose polarisations have variable amounts of radial and transverse components. This, in particular, allows conversion of an incident transversely polarised SH into qP. Both SV-to-qP and SH-to-qP conversions will be referred as  $S_qp$  in the following text and figures.

Characteristics of the SV and SH receiver functions (Fig. 3.10 and 3.11) in the presence of a shallow anisotropic layer are described below:

- (a) S wave splitting into qS1 and qS2 is apparent as a pulse at τ = 0 s of the radial receiver function, but a pulse at τ < 0 s on the transverse (Note that in the current representation of times a negative delay represents a delayed pulse). The negative delay on the transverse varies with back azimuth;
- (b) conversion of the qS1 and qS2 particle motions into qP is observed on both radial and transverse receiver functions. However, the amplitude ratio S<sub>q</sub>p/S is larger on the radial than on the transverse;
- (c) S and  $S_q p$  polarities of the transverse receiver function are anti-symmetric about the direction of the symmetry axis. They reverse at 0° and 90° (or less) and 180° from this direction in the case of an horizontal (or dipping) axis. The polarity reversal is at ±60° in the case of an axis of symmetry plunging 45° to the north.

In addition, in the case of an incident SH (Fig. 3.11):



Figure 3.10: Effect of anisotropy with slow axis of symmetry (model ANISO1 of Tab. 3.1) on SV receiver functions as a function of epicentral distance at 0° from north (top) and back azimuth at 50° epicentral distance, i.e. slowness of 13.9 s/° (*bottom*). Anisotropy is -5 %, has hexagonal symmetry and a slow axis plunging at 45° to the north (Fig. 3.9). A 0.5 Hz low-pass filter was applied. Ray parameters are same as in Figure 3.6. Amplitude and time scales are reversed as discussed in text.



Figure 3.11: Same as Figure 3.10 for an incident SH wave.

- (a) the initial qS1/qS2 and corresponding conversions into qP appear as two short pulses with opposite polarities on the transverse component;
- (b)  $S_q p$  delay time variations as a function of back azimuth are smooth in the case of a horizontal axis of symmetry (not shown) but occur as an abrupt phase shift on the radial component in the case of a dipping axis of symmetry. Phase shifts and amplitudes of the  $S_q p$  on the radial receiver function are symmetric about the symmetry axis.

#### Similarities with P receiver functions:

- (a) Splitting of the incoming S is apparent in the presence of energy on the receiver function component perpendicular to the incoming S polarisation (Savage, 1998).
  This includes pulses for the qS and corresponding S<sub>q</sub>p conversion,
- (b) polarities of the transverse receiver function are anti-symmetric about the azimuth of the symmetry axis and reverse at 0° and ≤90° from the axis direction (Savage, 1998).

#### Differences with P receiver functions:

- (a) The S pulse on the transverse S receiver function, displays a phase shift that does not exist in P receiver functions;
- (b) in the case of an incident SH, the direct S and conversions consist of two short pulses with opposite polarities;
- (c) the SH-to-qP conversion displays sudden phase shifts on the radial receiver function that are symmetric about the symmetry axis.

#### 3.3.3 Effect of mantle anisotropy

Lattice Preferred Orientation of olivine minerals is the primary cause of mantle anisotropy. In models ANISO2a–d (figure below; Tab. 3.1), the anisotropic mantle is a half-space with fast propagation along the symmetry axis. The anisotropic mantle is overlain by an isotropic crust. Models ANISO2a–c have symmetry axes that are horizontal, dipping 45° N and vertical, respectively. Finally in model ANISO3 (Tab. 3.1), the fast axis of model ANISO2a is replaced by a slow one.



Figure 3.12: Models ANISO2a-c of Table 3.1

Figure 3.13: Model ANISO2d (Tab. 3.1) is similar to model ANISO2a but has a lower wavespeed contrast at the top of the anisotropic halfspace than model ANISO2a has.



The incident SV or SH is split in the anisotropic half-space into quasi-S waves, qS1 and qS2. Both qS1 and qS2 are converted into S waves (i.e. S1 and S2) at the top of the anisotropic medium and propagate independently up to the surface. In addition, the portion of the qS particle motion that is parallel to the propagation direction is

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Figure 3.14: Effect of anisotropy with horizontal fast axis of symmetry on SV receiver functions (model ANISO2a of Tab. 3.1) as a function of epicentral distance at 0° back azimuth (*top*) and back azimuth at 50° epicentral distance, i.e. slowness of 13.9 s/° (*bottom*). The anisotropy is of  $\delta P = 6$  % and  $\delta S = 4$  % and has hexagonal symmetry. The symmetry axis is horizontal and oriented north.



Figure 3.15: Effect of a plunging symmetry axis on the direct S pulse polarity (compare with top of Figure 3.14). Synthetic SV receiver functions are presented as a function of epicentral distance at  $0^{\circ}$  back azimuth. These are for model ANISO2b with a symmetry axis plunging  $45^{\circ}$  to the north (*top*) and model ANISO2c with a vertical symmetry axis (*bottom*). Other model parameters are same as in model ANISO2a (Tab. 3.1).



Figure 3.16: Effect of the velocity contrast at the top of the anisotropic half-space on the direct S pulse polarity. Synthetic SV receiver functions are presented as a function of epicentral distance at 0° back azimuth. These are calculated for model ANISO2a with a 10 % Vp increase with depth from 6.8 km/s (*top*) and model ANISO2d (Tab. 3.1) with 18 % Vp increase (*bottom*).

converted into P at the top of the half space. In the case of overlying isotropic layers only, S1 and S2 polarisations are conserved up to the surface. In the other case, their polarisations are altered.

The anisotropy has the following effects on SV receiver functions (Fig. 3.14-3.16):

- (a) shear-wave splitting of the incident SV or SH is apparent as a pulse at 0 s time delay on both radial and transverse components (no phase shift on the transverse receiver function as in contrast to shallow anisotropy);
- (b) the polarity of the initial S pulse of both radial and transverse periodically switches with the back azimuth and the epicentral distance (Fig. 3.14). On the radial and for close epicentral distance of  $\sim 30^{\circ}$  (S incident angle  $i_{inc} \sim 40^{\circ}$ ; not shown), the direct SV pulse polarity is negative at all back azimuths. At epicentral distances up to  $45^{\circ}$  ( $i_{inc} \geq 35^{\circ}$ ), however, the SV pulse polarity is negative for back azimuths within  $45^{\circ}$  of the fast polarisation orientation (bottom of Figure 3.14). The critical epicentral distance, up to which a polarity reversal occurs, decreases as the symmetry axis steepens (compare top of Figure 3.14 and 3.15; models ANISO2a-c) or as the velocity contrast at the top of the anisotropic medium decreases (Fig. 3.16; model ANISO3);
- (c) on the transverse, the initial S pulse polarity reverses every 45° from the symmetry axis (Fig. 3.14; model ANISO2a);
- (d) Sp conversions exist on the radial receiver function and are positive where the initial S pulse is positive but are absent where the initial S pulse is negative;
- (e) Conversely, Sp conversions are strong on the transverse component where the initial S pulse is absent on the radial.



Figure 3.17: Same as Figure 3.14 for an incident SH wave.

SH receiver functions (Fig. 3.17) display similar characteristics to the SV receiver functions (Fig. 3.14) with a few exceptions:

- (a) the negatively polarised S pulse occurs on the transverse only, that is the S pulse is always positive on the radial but is negative on the transverse for back azimuths within 45° east of the symmetry axis and of the perpendicular direction;
- (b) the Sp conversion on the transverse receiver function has same polarity as the direct S-wave pulse.

Figure 3.18: In model ANISO3 (Tab. 3.1), the fast axis of symmetry of model ANISO2a has been replaced by a slow axis. Although receiver functions (Fig. 3.19) present characteristics similar to receiver functions of model ANISO2a (Fig. 3.14), polarity switches of the initial S pulse occur more often, circa every  $30^{\circ}$  on the radial.



#### Similarities with P receiver functions:

Anisotropy produces receiver functions with a strong transverse component (Levin *et al.*, 2002; Savage, 1998).

#### Differences with P receiver functions:

(a) The split S results in energy at τ = 0 s on both radial and transverse receiver functions. This is a major difference with P receiver functions, where in the case of deep anisotropy, there is no direct P pulse on the transverse receiver function (Savage, 1998);



Figure 3.19: Effect of deep anisotropy with horizontal slow axis of symmetry on SV receiver functions (model ANISO3 of Tab. 3.1) as a function of epicentral distance at 0° back azimuth (*top*) and back azimuth at 50° epicentral distance, i.e. slowness of 13.9 s/° (*bottom*). The anisotropy is of  $\delta P = -6$  % and  $\delta S = -4$  % and has hexagonal symmetry. The symmetry axis is oriented north.

(b) at large S incidence angles (large slowness) and large impedance contrasts at the top of the anisotropic medium, the initial S pulse has negative polarity. This is interpreted as the result of the elliptical particle motion that arises from the interference between the incident S wave and its free surface Sp conversion, which occurs at S wave incidence angles greater than the S critical angle, i.e. incidence angles greater than allowed by the shear wave window (Booth and Crampin, 1985). A possible explanation for the unusually large S wave incidence at the free surface is given by the anomalous refraction angle, which occurs when a phase is converted at an interface separating an anisotropic medium from an isotropic one (e.g. Slawinski *et al.*, 2000).

# 3.4 Conclusions

In the case of flat isotropic layers SV receiver functions are directly comparable with P receiver functions. In the presence of dipping layers or anisotropy, however, polarities in the S and P receiver functions differ appreciably. These differences are inherent to the S and P wave particle motions.

For instance, in P receiver functions, the presence of energy at  $\tau \sim 0$  s on the transverse component can be attributed entirely to a dipping layer or shallow crustal anisotropy (Savage, 1998). In S receiver functions, in contrast, energy at  $\tau \sim 0$  s is no longer a criterion as such energy is present in most cases: dipping boundary, shallow and deep anisotropy.

Furthermore with P receiver functions, dipping boundaries produce Ps conversions that appear on both radial and transverse receiver functions. In the case of an incident S wave, however, the Sp conversion from a dipping boundary appears on the receiver function component parallel to the polarity of the incident S only, i.e. on the radial receiver function for an incident SV but on the transverse receiver function for an incident SH.

Polarity reversals on the transverse receiver function may be a further criterion to distinguish between a dipping boundary, and shallow or deep anisotropy. In the case of a dipping boundary, polarity reversals on the transverse component are similar to those of P receiver functions – they occur at 180° back azimuth intervals (Savage, 1998). In the case of shallow anisotropy, the polarity reversals occur at  $\leq 90^{\circ}$  intervals as in P receiver functions (Savage, 1998) but, unlike in that case, they occur more frequently in the presence of deep anisotropy. However, these polarity reversals may be difficult

model	Η	ρ	$\mathbf{V}_P$	$\mathbf{V}_S$	$\delta P$	$\delta S$	trend	plunge	$\operatorname{strike}$	dip	$V_P/V_S$
	(km)	$(\mathrm{kg}/\mathrm{m}^3)$	(m/s)	(m/s)	(%)	(%)	$(^{\circ})$	$(^{\circ})$	$(^{\circ})$	$(^{\circ})$	
DIP8	40	2700	6800	3885	-	-	-	-	-	-	1.75
	20	3300	8000	4571	-	-	-	-	-90	8	1.75
ANISO1	5	2700	6800	3885	-	-	-	-	-	-	1.75
	35	2700	6800	3885	-5	-5	0	45	-	-	1.75
	20	3300	8000	4571	-	-	-	-	-	-	1.75
ANISO2a	40	2700	6800	3885	-	-	-	-	-	-	1.75
	20	3300	8000	4571	6	4	0	0	-	-	1.75
ANISO2b	40	2700	6800	3885	-	-	-	-	-	-	1.75
	20	3300	8000	4571	6	4	0	45	-	-	1.75
ANISO2c	40	2700	6800	3885	-	-	-	-	-	-	1.75
	20	3300	8000	4571	6	4	0	90	-	-	1.75
ANISO2d	40	2700	6800	3885	-	-	-	-	-	-	1.75
	20	3300	7500	4286	6	4	0	0	-	-	1.75
ANISO3	40	2700	6800	3885	-	-	_	-	-	-	1.75
	20	3300	8000	4571	-6	-4	0	0	-	-	1.75

Table 3.1: Models of synthetic SV	and SH receiver functions of Figure 3.6–3.17.
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to distinguish in the presence of complex crustal structure where dipping boundaries and anisotropy are both present.

Shear-wave splitting of an incident S wave in the presence of anisotropy is one reason for differences between P and S receiver functions. In the case of an incident P wave traversing an anisotropic medium, the particle motion is slightly rotated away from the propagation direction into a quasi-P. The quasi-P particle motion, however, is reset to a true radially polarised P in a subsequent isotropic layer. Thus, later Ps conversions of P receiver functions are unaffected by a deep anisotropic layer.

In contrast, the incident S wave is split while traversing the anisotropic medium, and remains split in subsequent isotropic layers (or is split further in overlying anisotropic layers). Hence, the polarity of an Sp conversion at a particular interface depends on the splitting of the S wave, which occurred in previously traversed and deeper anisotropic layers below the converting interface. Conversely, the quasi-P particle motion of the Sp conversion traversing an anisotropic layer above that interface will be reset to a true radially polarised P particle motion in a subsequent isotropic layer. Thus, in the presence of multiple anisotropic layers, Sp conversions from deeper interfaces should present fewer polarity reversals with back azimuth and stack more constructively than Sp conversions from shallow interfaces.

A direct S pulse with negative polarity on the radial receiver function occurs in two distinct cases: 1) an SKS phase incident onto a dipping boundary with strong velocity contrast at the top; and 2) an S phase traversing an anisotropic layer with a strong impedance contrast at the top.

1) At SKS incidence, similar to incident P (Savage *et al.*, 2007a), such a reverse polarity occurs in the case of a dipping interface with a strong velocity contrast at the

top, and when the transmission angle  $\theta$  of the updip-propagating S is smaller than the dip  $\alpha$  of the interface, i.e.  $\theta < \alpha$  (Fig. 3.8).

2) For the shallow incident S, however, an anisotropic layer with strong velocity contrast at the top is necessary in order to reverse the polarity of the direct S pulse. The polarity reversal depends on the angle between the incident S and the symmetry axis of the anisotropic medium, and therefore, depends on both back azimuth and epicentral distance. On the radial receiver function and at epicentral distances of  $\sim 30^{\circ}$ , the polarity reversal occurs at all back azimuths. In the case of anisotropy with a horizontal fast symmetry axis, the reversal persists at epicentral distances up to  $45^{\circ}$ for back azimuths within  $45^{\circ}$  of the symmetry axis. The maximum epicentral distance to which the reversal persists, i.e. the critical epicentral distance, decreases with decreasing velocity contrast and increasing dip of the symmetry axis. Such a velocity contrast may be encountered at the Moho or the top of the lower crust. However, this explanation is non unique. A combination of dipping boundaries and/or anisotropic layers can sufficiently alter the S polarisation to produce a negative initial S pulse from elsewhere in the lithosphere.

These synthetic receiver functions have sampled only a few idealised cases. The lithosphere contains more complexity than in the one-layer models tested here, and modelling is probably needed for each special case.

# Chapter 4

# P and S receiver function analysis

## 4.1 Abstract

The lithospheric structure is investigated beneath four South Island locations using teleseismic receiver functions compiled from three-component seismograms recorded at Geonet broad-band stations: RPZ (Rata Peak), JCZ (Jackson Bay), WKZ (Wanaka) and EAZ (Earnscleugh). Crustal thickness, Vp, Vs, anisotropy and dipping layers of the subsurface are estimated by matching the radial and transverse components of P, S and SKS receiver functions with synthetic receiver functions. S and SKS receiver functions with synthetic receiver functions. S and SKS receiver functions as discussed in the previous chapter.

Receiver functions at station RPZ imply a low Vp/Vs ratio of  $\sim 1.60$  for the crust of central South Island, consistent with earlier studies of (Kleffmann, 1999; Pulford, 2002).

Across the southern South Island, receiver functions reveal the deepening of a discontinuity from  $34 \pm 5$  km (JCZ) and  $32 \pm 3$  km (EAZ) on either side of the Southern Alps to  $39 \pm 4$  km depth beneath the highest topography (ca. 50 km NW of WKZ) assuming Vp/Vs ratios of 1.65–1.75. The interpretation of this discontinuity as the Moho is inconsistent with Bouguer and isostatic gravity anomalies of -80 mGal and -10 mGal at station WKZ. Such Moho depths are also inconsistent with crustal thickness estimates of 43–45 km and  $48 \pm 4$  km from 3D inversion of travel times (Eberhart-Phillips and Bannister, 2002) and earthquake refraction (Section 2.4, p. 29), respectively. A low crustal Vp/Vs ratio of 1.55 is necessary to reconcile conversion delay times with a ~48 km deep Moho. Alternatively, the discontinuity may represent a lower crustal boundary, implying a weak contrast at the Moho.

A mantle discontinuity is interpreted to exist at 50–70 km depth, i.e. 15–30 km below the presumed Moho. This discontinuity may represent a rotation of the fast symmetry axis azimuth and/or an additional Vp increase of  $\sim 0.3$  km/s respectively suggested by S and P receiver functions. The Vp increase may be interpreted as the result of a phase change such as partial eclogitisation or the transition from spinel to garnet peridotite, called Hales discontinuity (Hales, 1969). A rotation of the anisotropy fast axis of symmetry could be explained by mechanical decoupling or transitional plasticity allowed by the volume change associated with phase transition (e.g. Bostock, 1997; Levin and Park, 2000; Sammis and Dein, 1974).

### 4.2 Introduction

While central South Island has in the past decade been the focus of extensive geophysical studies notably SIGHT, less attention has been paid to the southern South Island and western Otago. This chapter applies the receiver function method to Geonet data to provide further information on the crustal structure across this region. Geonet Canterbury station RPZ (Fig. 4.1) is selected where the crustal structure is known (Van Avendonk *et al.*, 2004) and provides comparison with crustal models derived with the present method. Three Geonet stations JCZ (Jackson Bay), WKZ (Wanaka) and EAZ (Earnscleugh) are selected that align perpendicular to the Alpine Fault (Fig. 4.1) and approximately parallel to the Jackson-Dunedin gravity profile of Section 2.6 of



Figure 4.1: Haast Schist textural zones of the southern South Island. Chlorite zones II, III and IV are known as the Otago Schist. The higher metamorphic grades of the biotite and garnet zones are known as Alpine Schist. *Red triangles* denote Geonet stations RPZ, JCZ, WKZ and EAZ used in this study. *Thick dark lines* represent metamorphic boundaries; *thin lines* are coast and lake shores.

Chapter 2.

Crustal thickness estimates for the southern South Island exist from this study's earthquake refraction (Section 2.4) with a  $48 \pm 4$  km near the Wanaka township , and from the joint-inversion for hypocentres, Vp and Vp/Vs models from active and passive source travel times (Eberhart-Phillips and Bannister, 2002). The 7.8 km/s isovel suggests an average crustal thickness of 43 km in western Otago with a maximum of 45 km directly southwest of Wanaka (Eberhart-Phillips and Bannister, 2002). Although the inverted wave speeds are well resolved where most hypocentres lie, i.e. above 25–30 km depth and up to 100 km from the Alpine Fault in the east, the resolution is less at Moho depth.

# 4.3 Otago and Alpine Schists

Stations JCZ, WKZ and EAZ are located on the Alpine and Otago Schists, both part of the Haast Schist group (Fig. 4.1; e.g. Suggate, 1961). The Otago Schist represents an accretionary prism that formed in Late Paleozoic–Mesozoic during subduction along the south Gondwana margin (Mortimer, 2000). Late Cretaceous extension on low-angle shear zones (Deckert *et al.*, 2002), erosion and possibly isostatic uplift as a consequence of underplating (Grapes and Watanabe, 1994) have combined to exhume the schist from deep within the accretionary prism. The Schist forms a ca. 150 km wide twosided metamorphic arch (Mortimer, 2000) with its axis oriented northwest-southeast, perpendicular to the present plate boundary. The Alpine Schist (Fig. 4.1), which is the highest grade schist, has been further exhumed by rapid Cenozoic uplift and erosion along the Alpine Fault. Thus, the Alpine Schist exhibits rocks with recent Cenozoic metamorphism and a deeper metamorphic record than the Otago Schist. As a result, the metamorphic grade not only increases towards the Otago Schist axis but also towards the Alpine Fault.

Thermobarometry (Grapes, 1995; Grapes and Watanabe, 1994, 1992) has constrained maximum pressures and temperatures experienced by the schists. Two metamorphic events "D1" and "D2" define a P/T path interpreted as prograde metamorphism in a low-heat flow, subduction setting (D1), followed by retrograde metamorphism during warming back to normal crustal temperatures (D2; Grapes and Watanabe, 1992). The first event "D1" is mainly preserved in chlorite and biotite-albite zone rocks (Otago Schist) that are characterised by high P/T ratios and maximum pressures of 430–750 MPa, which scale to 15–25 km depth. The later event "D2" is preserved in the garnet and oligoclase zones (Alpine Schist) and is characterised by a lower P/T ratio but higher maximum pressures of 520–920 MPa (equivalent to 20–32 km depth) than the first event (Grapes and Watanabe, 1992).

The maximum pressures recorded in "D1" suggest that schists at the surface were once buried at 15–25 km depth, and that rocks once buried to depths greater than 35 km would have been within the eclogite stability field. Eclogitic xenoliths found 80 km north of Dunedin (southeast South Island) and estimated to originate from 60 km depth (Mason, 1968) provide evidence for partial eclogitisation directly northeast of Otago. Such rocks could exist beneath Otago at a similar depth in the mantle or even shallower, i.e. in the lower crust, where exhumation was largest.

#### 4.4 Data set

P, S and SKS (SKSac branch) phases are extracted from three-component seismograms from the Geonet broad-band sites RPZ, JCZ, WKZ and EAZ (Tab. 4.1) for teleseisms



Figure 4.2: Geonet network (*grey triangles*) and estimated Ps, Sp and SKSp piercing points beneath stations RPZ, JCZ, WKZ and EAZ (Tab. 4.1) for an assumed conversion depth of 30 km and crustal Vp and Vs of 6.2 km/s and 3.54 km/s, respectively. *Blue lines* are Bouguer gravity anomaly contours at 30 mGal intervals (Reilly and Whiteford, 1979). *Dark lines* are geophysical transects SIGHT T1 and T2 (Okaya *et al.*, 2002) and the southwest-northeast Fiordland-Cheviot earthquake refraction profile of Chapter 2.

produced by earthquakes of  $M_W \ge 6.5$  and epicentral distances of 20°-100°, 30°-85° and 90°-130°, respectively. Approximately 60, 30 and 25 good P-, S and SKS receiver functions, respectively, are derived for stations JCZ, WKZ and EAZ and 74, 41 and 33 for the longer operating RPZ station (Tab. 4.1). While most P and S phases arise from earthquakes in the western Pacific, half of the SKS phases originate in the southeastern Pacific with back azimuths of 100°-140° (Fig. 4.2, 4.3). The uneven earthquake distribution is expected to produce a trade-off between back azimuth and epicentral distance.



Figure 4.3: Earthquake distribution for SKS, P and S phases recorded at station RPZ as a function of back azimuth and epicentral distance. These earthquake distributions are representative of those at stations JCZ, WKZ and EAZ.

Table 4.1: Geonet broad-band stations Earnscleugh (EAZ), Jackson Bay (JCZ), Rata Peak (RPZ) and Wanaka (WKZ) locations, installation dates and number of selected P, S and SKS phases.

station	latitude	longitude	height	available	number of events			
	(°)	(°)	(m)	since	Р	$\mathbf{S}$	SKS	
EAZ	-45.23268	169.3082	350	Nov. 2004	55	30	26	
JCZ	-44.07487	168.7854	1062	Jun. 2004	61	32	26	
RPZ	-43.71627	171.0538	453	Jun. 2001	74	39	33	
WKZ	-44.82866	169.0175	558	Jun. 2004	62	32	27	
# 4.5 Method

## 4.5.1 Receiver function preparation

P, S and SKS receiver functions are compiled from three-component seismograms in the Z-R-T domain, i.e. north and east components are rotated into radial and transverse directions in the right-handed convention. After rotation, seismograms are input into Park and Levin's code (2000) to compute P receiver functions, and into its modified version to compute S and SKS receiver functions (Section 3.2). The selected pre-event and post-event windows are 20 s and 150 s long for P receiver functions, and enable the retrieval of Ps conversions of up to 30 s. These windows are 110 s and 40 s long for the S receiver functions in order to retrieve Sp conversions up to 22 s (Section 3.2, p. 58). In both P and S receiver functions, the pre-event noise window is taken 10 s before the P onset. Low-pass cut-off frequencies of 0.5 Hz or 1 Hz are used, effectively allowing energy beneath 0.33 Hz and 0.65 Hz, respectively. Single receiver functions are stacked in slowness bins of  $0.4 \text{ s/}^{\circ}$  and back azimuth and epicentral distance bins of  $10^{\circ}$  with increments of  $0.2 \text{ s/}^{\circ}$  and  $5^{\circ}$ , respectively, so that each receiver function contributes in two neighbouring bins.

# 4.5.2 Forward modelling

In a first stage, the layer thickness, H, and the Vp/Vs ratio are estimated by fitting Ps-Sp conversion and free-surface multiple move-outs and delay times,  $\tau_{Ps}$ ,  $\tau_{PpPs}$  and  $\tau_{PpSs+PsPs}$ , from the slowness stack of radial receiver functions. The thickness is given by:

$$H = \frac{\tau_{Ps}}{\sqrt{\frac{1}{V_s^2} - p^2} - \sqrt{\frac{1}{V_p^2} - p^2}},$$

with p, the ray-parameter of the incident wave (herein called slowness). H is highly dependent on the Vp/Vs ratio, and a change in Vp/Vs ratio of 0.1 yields a change of

4 km in H (Zhu and Kanamori, 2000). P receiver function free-surface multiples PpPs, PsPs and PpSs can help to considerably reduce the error in H and Vp/Vs (Zhu and Kanamori, 2000) through:

$$H = \frac{\tau_{PpPs}}{\sqrt{\frac{1}{V_s^2} - p^2} + \sqrt{\frac{1}{V_p^2} - p^2}}, \qquad H = \frac{\tau_{PpSs + PsPs}}{2\sqrt{\frac{1}{V_s^2} - p^2}}.$$
(4.1)

The change from subtraction to addition in the denominator of the above equations yields opposite trends between delay times of direct conversions and multiples. For direct conversions, delay times increase with increasing slowness or decreasing distance. This relationship is called positive move-out. In contrast for multiples, delay times decrease with increasing slowness. The move-out is referred to be negative. This characteristic is used to identify and discriminate multiples from direct conversions in the P receiver functions. In the presence of anisotropy, and dipping layers especially, the move-out may appear anomalous relative to that predicted for a flat layer, in particular when looking at receiver function stacks for a restricted back azimuth range. This will lead to errors in the parameter estimates. Therefore, these estimates are only used as starting parameters in forward modelling.

Models are found by matching synthetic receiver functions with the back azimuth and epicentral distance stacks of receiver functions, and by visually assessing the match. Frederiksen and Bostock's code (2000) is used to produce synthetics as described in Section 3.2. This code allows for dipping layers and anisotropy with hexagonal symmetry and a plunging symmetry axis, which is advantageous in modelling the highly anisotropic and complex lithospheric structure of South Island. A 1 s pulse is used to fit 1 Hz low-pass filtered receiver functions and a 2 s pulse for 0.5 Hz low-pass filtered data. Trial and error is used in most cases. In some instances, however, a neighbourhood algorithm (Sambridge, 1999) is used to invert for some parameters (these will be indicated with an asterix in parameter Table 4.6). Because of the noise level in the real receiver functions, the inversion will tend to match pulses associated with both noise and conversions. Thus, thicknesses and wave speeds are fixed to match the time delay of a particular conversion and only strike/dip of dipping layers and trend/plunge of anisotropy symmetry axes are computed at one time. When inverting for anisotropy, a maximum of two layers should be inverted for (Scherrington *et al.*, 2004), because the resulting parameters will depend directly on the anisotropy in the surrounding layers.

# 4.6 Analysis

Conversion pulses at 2–3 s and 4–5 s of the receiver functions coincide with typical delay times from the top of the lower crust (C) and the Moho (M), respectively, and are labelled as such (Fig. 3.3). Additional mantle conversions with 5 s  $< \tau < 11$  s are labelled X. Unless otherwise stated, models are derived by trial and error, and uncertainties on model parameter are estimated from the range of matching models. Both 0.5 Hz and 1 Hz low-pass filters are used to interpret the receiver functions, but the match of synthetic receiver functions to real data is displayed with a 0.5 Hz low-pass filter in order to match major features in the receiver functions.

# 4.6.1 Rata Peak station (RPZ)

Lithospheric models derived from RPZ P-, S and SKS receiver functions are compared with the crustal structure known from the SIGHT Transects 1 and 2 (Scherwath *et al.*, 2003; Van Avendonk *et al.*, 2004) in order to assess the potential of the present P-, S and SKS receiver functions.

## Observations

Crustal conversions are most distinct if low-pass filtered below 1 Hz. Slowness and epicentral distance stacks are compiled from receiver functions in the back azimuth



Figure 4.4: Station RPZ slowness stack of SKS ( $p < 6.6 \text{ s/}^{\circ}$ ), P (6.8 s/ $^{\circ} \leq p \leq 9 \text{ s/}^{\circ}$ ) and S (10 s/ $^{\circ} ) radial receiver functions for back azimuths in the range 270<math>^{\circ}$ -0 $^{\circ}$ . A 1 Hz low-pass filter was used. *Blue and red wiggles* denote positive and negative amplitudes, respectively, i.e wave speed increasing (*blue*) and decreasing (*red*) with depth. *Color-coded curves* describe Ps-Sp (*solid*) and free-surface multiple (*dashed*) delay times predicted for assumed average Vp, Vp/Vs and conversion depth of: 6.1 km/s, 1.60 and 23 km (C); 6.3 km/s, 1.65 and 34 km (M); and 6.8 km/s, 1.70 and 62 km (X). *Grey bars* on the right-hand side denote the number of events per slowness bin of 0.4 s/ $^{\circ}$ .

range  $270^{\circ}-0^{\circ}$ , within which receiver functions sum up constructively (Fig. 4.4). Below, pulses of the direct P-, S- and SKS-waves and conversions C, M and X are discussed in order of appearance:



Figure 4.5: RPZ 1 Hz low-pass filtered P receiver functions (*coloured wiggles*) and model RP1M synthetic P receiver functions (*black curves*; Tab. 4.2) as a function of epicentral distance (*top*) and back azimuth (*bottom*).

- (a) The direct P-wave pulse is delayed up to ~0.5 s towards northern back azimuths (see slowness stack of Fig. 4.4, also Fig. 4.5), indicating a shallow low velocity layer.
- (b) The Ps crustal conversion at τ ~ 2 s (labelled C in Fig. 4.4) is sharp and is the strongest conversion in the P receiver functions. Maximum C delay times on the radial P receiver functions are observed at ca. -30° back azimuth (Fig. 4.5). A simultaneous polarity switch at τ ~ 2 s of the transverse receiver functions from negative at back azimuths of -120° to -30° to positive from -30° to 150° suggests that the crustal discontinuity is dipping to the northwest. A positive pulse with negative move-out at τ ~ 9-10 s of the P receiver functions is interpreted as a PpPs free-surface multiple of C (Fig. 4.4).
- (c) In contrast to C, Ps conversions at  $\tau \sim 3-4$  s (M in Fig. 4.4) have a clear Sp continuation within the SKS receiver functions.
- (d) A pulse with apparent positive move-out at τ ~ 6–7 s (X in Fig. 4.4) is interpreted as a mantle conversion. Such an Sp conversion is unclear in the SKS receiver functions, but could have a continuation in the S receiver functions (Fig. 4.4).
- (e) Initial pulses of S radial receiver functions at  $\tau \sim 0$  s are negative for slowness  $p \ge 13 \text{ s/}^{\circ}$  (Fig. 4.4) or epicentral distances  $\Delta \le 45^{\circ}$  (Fig. 4.6).
- (f) Initial pulses of SKS receiver functions are weak or negative in the northwest quadrant (Fig. 4.4).
- (g) A negative pulse at  $\tau = 14-16$  s and slowness  $\geq 13$  s/° of the S receiver functions is interpreted as numerical ringing (Fig. 4.4).



Figure 4.6: RPZ 0.5 Hz low-pass filtered S receiver functions (*coloured wiggles*) and model RS1M synthetic SV receiver functions (*black curves*; Tab. 4.2).

#### **P** receiver function synthetics

Minimum and maximum crustal thicknesses of models RP1m/-M (Fig. 4.7; Tab. 4.2) are found by trial and error and using SIGHT Transect 1 and Transect 2 minimum Vp of 6 km/s and maximum Vp of 6.2 km/s (Scherwath *et al.*, 2003; Van Avendonk *et al.*, 2004) in combination with maximum and minimum Vp/Vs ratios of 1.75 and 1.60, respectively. Indeed, a low Vp/Vs ratio of 1.60 yields a smaller delay time between the direct P wave and its Ps conversion than a high Vp/Vs ratio of 1.75. Thus, a low Vp/Vs ratio of 1.60 requires larger crustal thicknesses to match Ps conversion delay times than a large Vp/Vs ratio of 1.75. The low Vp/Vs ratio therefore yields thickness upper bounds, the high Vp/Vs ratio yields lower bounds.

A low upper crustal Vp/Vs ratio is suggested by a  $\sim 1.60$  estimated from the C conversion and corresponding free-surface multiples PpPs and PsPs/PpSs delay times (Fig. 4.4). The low value is consistent with a low average crustal Vp/Vs ratio of 1.65 (Kleffmann, 1999) in the central Southern Alps. The maximum Vp/Vs ratio of 1.75 is based on maximum crustal averages for intermediate crustal compositions (Zandt and Ammon, 1995).

In model RP1m/-M synthetic receiver functions (Fig. 4.5), the delayed P arrival (see Observations section a) is matched with a ca. 2 km thick top layer, assuming a Vp of 4 km/s.

The crustal conversion C (see b) is modelled with a 17–20 km deep discontinuity. The sharp and strong amplitudes at northwestern back azimuths, the maximum delay times on the radial component and the anti-symmetric polarities on the transverse component both with respect to  $\sim -30^{\circ}$  (see b) are reproduced with a northwest-dipping interface (strike/dip of 200–240°/20°–40°) that is interpreted as the top of the lower crust (Fig. 4.7).

The Moho conversion, M (see c), is matched with a 30–36 km deep interface dipping southwest (strike/dip of ca.  $150^{\circ}/10-15^{\circ}$ ). Note that a low Vp/Vs ratio of 1.60 for the crust but a Vp/Vs of 1.75 for the mantle reduces the contrast between mantle Vp and crustal Vs at the Moho and, hence, the Moho conversion amplitudes relative to conversion amplitudes in the data (model RP1M synthetics of Fig. 4.5).

Delay times of the mantle conversion X (see d) are matched by an interface ca. 30 km below the Moho, i.e. ca. 60-66 km depth, assuming a Vp/Vs ratio of 1.75 in the mantle lid.



Figure 4.7: Crust and mantle structure interpretation below station RPZ from S and P receiver functions (models RS1m/-M and RP1m/-M of Tab. 4.2, respectively). Note, the width of the models and the distance between them is not to scale.

### S and SKS receiver function synthetics

Models RP1m and RP1M display a gross fit (not shown) to S and SKS receiver functions of Figures 4.6 and 4.8 with the exception of the steeply dipping lower crust and Moho, which preclude the Sp conversion for the updip-travelling S wave. A maximum dip of ca. 10° W is allowed for the lower crustal boundary instead of 20°–40° modelled from the P receiver functions (Fig. 4.6; model RS1m of Tab. 4.2). Using the same minimum and maximum Vp and Vp/Vs combinations as for P synthetic receiver functions, depths to the lower crust and the Moho are 15–24 km and 30–43 km, respectively (Fig. 4.7). The Moho conversion is, however, only a weak pulse in the S receiver functions so that the depth is only weakly constrained. The Moho depth estimated from the SKS receiver functions is an approximate 31 km, for a Vp/Vs ratio of 1.70 (Fig. 4.8; model RSKS of Tab. 4.5).



Figure 4.8: Station RPZ 1 Hz low-pass filtered SKS receiver functions (*coloured wiggles*) and model RSKS synthetics (*black curves*; Tab. 4.2). Note that traces for  $-70^{\circ}$ ,  $40^{\circ}$  and  $100^{\circ}$  back azimuth are for only one event.

The negative polarities corresponding to the direct S-wave pulse (see e) are reproduced with mantle anisotropy below the Moho (see Section 3.3.3, p. 70). A fast azimuth of  $\sim 50^{\circ}$  is assumed from SKS fast polarisations (Klosko *et al.*, 1999), which offers a good match to most polarities of the direct S-wave pulse (Fig. 4.6).

Polarities of X at 7–8 s of the S receiver functions (Fig. 4.6) are modelled with a 50° counter-clockwise rotation of the fast symmetry axis at 55-73 km depth. However, reversing the order of the symmetry axes, i.e. a 50° anti-clockwise rotation of the symmetry axes, offers a similar match to the S-wave direct pulse polarities. Therefore, in the presence of multiple anisotropic layers, the combination of symmetry axes can't be constrained from the S-wave direct pulse polarity alone.

Negative polarities of the direct SKS-wave pulse of SKS receiver functions (see f) are matched with a shallow  $50^{\circ}$  NE-dipping interface (Fig. 4.8; see Section 3.3.1).

Note that where the direct S-wave pulse has negative polarity on the radial receiver function, synthetic SV receiver functions predict no Sp conversion on the radial component. This is inconsistent with the data, in which converted energy exists after the initial SV. The difference may be explained by some amount of SH energy in the receiver functions. A weighted sum of SH and SV synthetic receiver functions based on single event focal mechanisms was also calculated. The resulting stack did not improve the fit to the data. Alternatively, rotation of seismograms into S parallel and S perpendicular particle motion directions as predicted by focal mechanisms was also tried. This rotation resulted in destructive stacking, however.



Figure 4.9: SIGHT T1 velocity model (Van Avendonk *et al.*, 2004). Blue, red and green error bars represent depth estimates to the crustal, Moho and mantle discontinuities, respectively, from forward modelling of P (thick bars) and S (thin bars) receiver functions (Tab. 4.2).

# Comparison with SIGHT Transect 1

A top shallow dipping layer is interpreted as a ca. 2 km thick layer of moraine/till covering rhyolite and greywacke basement (Gair, 1967). The modelled crustal thickness is 30–36 km from P receiver functions and ca. 31 km from SKS receiver functions, which is consistent with 32 km estimated from inversion of wide-angle reflection/refraction data (Fig. 4.9; Van Avendonk *et al.*, 2004).

The crustal thickness estimated from S receiver functions is 30–43 km, i.e a 4 km greater thickness and a 4 km greater range than from P receiver functions. The difference in thickness may be attributed to the uneven back azimuth distribution of events in combination with crustal structure lateral variations between Ps and Sp piercing points (Fig. 4.2). Figure 4.9 indeed suggests that the majority of Sp phases cross-cut the Moho, where the crustal root is 3–5 km thicker, than where Ps and SKSp phases do.

The lack of constraint on the lower crust and Moho depth estimates is a consequence of the noise in the S receiver functions, which could result from structural complexities associated with the crustal root.

Estimated upper and lower crustal thicknesses of 17–20 km and 15–16 km from P receiver functions and 15–24 km and 15–19 km from S receiver functions (Tab. 4.2) are respectively thinner and thicker than 27 km and 5 km modelled at SIGHT Transect 1 (Van Avendonk *et al.*, 2004). A reason could be high and low average Vp/Vs ratios used for the upper and lower crust, respectively, compared to true values.

P receiver functions suggest that the Moho is dipping  $10^{\circ}-15^{\circ}$  SW, which is what may be expected from apparent crustal thickening between SIGHT Transects 1 and 2 models (Scherwath *et al.*, 2003; Van Avendonk *et al.*, 2004) and an increase in the bulk of the negative Bouguer anomaly to the southwest (Fig. 4.2). Similarly, a lower crust dipping  $20^{\circ}-40^{\circ}$  W may be expected from a thinner lower crust at SIGHT Transect 1 than at Transect 2.

#### Model parameters uncertainties and trade-offs

The range of models obtained suggests a minimum 4 km uncertainty on the Moho depth, and uncertainties of  $30^{\circ}$  on strikes and  $10^{\circ}$  on dips of the lower crustal boundary and the Moho. The uncertainty on the trend of the anisotropy axis of symmetry is a minimum of  $10^{\circ}$ , the spacing between two bins in a back azimuth stack, and as large as  $30^{\circ}$ . Uncertainties are related to the noise level in the receiver functions, to the uneven distribution of events, which affects strike and trend estimates, and also to trade-offs with the overlying structure such as dipping boundaries, wave-speed structures

(Ammon *et al.*, 1990) and, hence, anisotropy. Below is a short description of how model parameters trade-off:

- small Vp/Vs ratios yield large conversion depths. The smaller the initially assumed Vp/Vs ratio, the smaller the resulting conversion delay time and the larger the modelled conversion depth that is necessary to match conversion delay times in the data;
- 2) small P-wave speeds yield large conversion depths;
- seismic anisotropy yields either large or small conversion depths depending on the difference between P and S anisotropy percentages and ray azimuths relative to the symmetry axis of the anisotropy;
- large dips yield large conversion depths for updip-travelling rays and small conversion depths for downdip-travelling rays;
- 5) large contrasts yield shallow dips;
- 6) seismic anisotropy yields either large or small contrasts depending on the ray azimuth relative to the symmetry axis of the anisotropy.

## 4.6.2 Earnscleugh station (EAZ)

Station EAZ is situated near the structural axis of the Otago Schist metamorphic antiform (Fig. 4.1).



Figure 4.10: Earthquake distribution for P and S phases recorded at station EAZ as a function of back azimuth and epicentral distance.

# Observations

Slowness and epicentral distance stacks (Fig. 4.12) are compiled for back azimuths in the range  $270^{\circ}-0^{\circ}$ , within which most data lie (Fig. 4.10) and receiver functions sum up constructively. SKS receiver functions at station EAZ display similar characteristics with those of station WKZ (Fig. 4.22 of Section 4.6.4) and to a lesser extent with those of station RPZ. The similarity suggests that the lithosphere is homogeneous at SKS wavelengths. Stations EAZ and WKZ SKS receiver functions will, therefore, be discussed together in a separate section below (123). P- and S receiver functions are analysed below:

(a) The direct P-wave pulse is delayed by ~0.5 s time of the 0.5 Hz low-pass filtered
 P receiver functions (Fig. 4.11), suggesting a shallow low velocity layer. On the



Figure 4.11: Station EAZ 0.5 Hz low-pass filtered P receiver functions (*coloured wiggles*) and model EP1M synthetic P receiver functions (*black curves*; Tab. 4.3).

transverse P receiver functions (Fig. 4.11), the polarity of the direct P-wave pulse switches from negative at back azimuths of  $-90^{\circ}$  to  $-10^{\circ}$  to positive east of  $-10^{\circ}$ .

- (b) Crustal and Moho conversions are interpreted at τ ~ 2 s and τ ~ 3–4 s (C and M in Fig. 4.12).
- (c) The Moho conversion at  $\tau \sim 4$  s of the P receiver functions has a corresponding transverse component at  $\tau \sim 3-4$  s of the transverse P receiver functions, whose polarities switch from negative to positive at  $-10^{\circ}$  (Fig. 4.11). In addition, a strong negative pulse with large negative move-out at  $\tau \sim 16-17$  s of the P receiver functions is interpreted as the Moho PsSs free-surface multiple. Energy on the transverse receiver functions and a strong PsSs multiple suggest a dipping Moho.
- (d) A strong and coherent positive pulse at τ ~ 19–20 s of the P receiver functions (Fig. 4.12) that has a large, negative move-out is interpreted as the PpPs/PpSs multiple of a mantle conversion at τ ~ 5–6 s (labelled X in Fig. 4.12).
- (e) The direct S-wave pulse is negative at slowness p > 12.5 s/° (Fig. 4.12) or epicentral distances Δ ≤ 60° and back azimuths east of −30° (Fig. 4.13), indicating the presence of an anisotropic layer with sufficiently large velocity contrast at the top (see Section 3.3.3, p. 70).

## Synthetic P receiver functions

Synthetics are fitted to 0.5 Hz low-pass filtered data of Figure 4.11. A minimum crustal thickness is found by using SIGHT Transect 2's minimum wave speeds of 6 km/s and 6.8 km/s for the upper and lower crust, respectively (Scherwath *et al.*, 2003), in combination with a Vp/Vs ratio of 1.75. A maximum crustal thickness is derived by using a low Vp/Vs ratio of 1.65 in combination with a Vp average of 6.3 km/s measured on



Figure 4.12: Station EAZ slowness stack of P (6.8 s/°  $\leq p \leq 9$  s/°) and S (10 s/° s/°) radial receiver functions for back azimuths in the range 270°–0°. A 1 Hz low-pass filter was applied.*Blue and red wiggles*denote positive and negative amplitudes, respectively.*Color-coded curves*are Ps-Sp (*solid*) and free-surface multiple (*dashed*) delay times predicted for assumed average Vp, Vp/Vs and conversion depth H of: 6.2 km/s, 1.70 and 20 km, respectively (C); 6.3 km/s, 1.70 and 32 km (M); and 6.9 km/s, 1.70 and 53 km (X).*Bars*on the right-hand side denote the number of events per bin of 0.4 s/°.

Haast Schist rock samples (Godfrey *et al.*, 2000, Chlorite zone IV) for the upper crust and taking 7.0 km/s for the lower crust, assuming that the lower crust may be partially eclogitised (e.g. Schulte-Pelkum *et al.*, 2005, for Tibet). Note that the same parameters



Figure 4.13: Station EAZ 0.5 Hz low-pass filtered S receiver functions (*coloured wiggles*) and model ES1M synthetic SV receiver functions (*black curves*; Tab. 4.3).



Figure 4.14: Crust and mantle structure interpretation below station EAZ from S and P receiver functions (models ES1m/-M and EP1m/-P of Tab. 4.3, respectively).

will be used for stations WKZ and JCZ.

The derived crustal thickness is 29–35 km and includes a 6–10 km thick lower crust (Fig. 4.14; models EP1m/-M of Tab. 4.3). A Moho dipping  $\sim 5^{\circ}-10^{\circ}$  W (strike of 170°–220°) is modelled in order to match the polarities of the direct P-wave pulse and those of the 4 s pulse of the transverse P receiver functions that switch from positive to negative at the back azimuth of  $-10^{\circ}$  (see Observations sections a and c). A dipping Moho also contributes to reproduce the strong PsSs multiple at  $\tau \sim 16$ –17 s of the radial P receiver functions (see c; Fig. 4.11).

Crustal anisotropy with sub-vertical slow axis of symmetry (trend/plunge of  $\sim 150^{\circ}/80^{\circ}$ ) improves the match to polarities of the transverse direct P-wave pulse (see a), and enhances the conversion amplitudes of C and M on the transverse P receiver functions (see c; Levin *et al.*, 2002). The modelled anisotropy, however, is imperfect as it tends to decrease the match to polarities of the crustal and Moho conversions on the radial receiver functions. The mantle discontinuity (X) at  $\tau \sim 6$  s in (Fig. 4.12) is modelled as a minimum 0.3 km/s velocity contrast at ca. 20 km below the Moho (50–55 km depth) assuming a Vp/Vs ratio of 1.75 in the mantle lid.

#### Synthetic S receiver functions

A crustal thickness of 30–38 km is derived from S receiver functions (Fig. 4.13; models ES1m/-M of Fig. 4.14, Tab. 4.3). In contrast to P receiver functions, a dipping Moho is not necessary to match S receiver functions.

An anisotropic layer with a high impedance contrast at the top is required to reproduce initial S pulses with negative polarity for  $\Delta \leq 60^{\circ}$  (Fig. 4.13; see e). The X pulse at  $\tau \sim 8$  s of the S receiver functions is modelled as a 45–58 km deep discontinuity. The pulse changing polarities with back azimuth and epicentral distance ( $\tau \sim 6$ –8 s in Fig. 4.13) could represent a rotation of the fast polarisation azimuth. A rotation from -20° below the Moho to 20° at the X discontinuity provides a reasonable match to the initial S pulse polarities.

However, similar to modelling of station RPZ S receiver functions, the symmetry axis combination is relatively unconstrained as the inverted combination offers a reasonable match to the S-wave pulse polarities too.

## Interpretation

Crustal thicknesses of 29–35 km and 30–38 km are estimated beneath station EAZ from P and S receiver functions, respectively. P receiver functions suggest a Moho dipping at  $5^{\circ}-10^{\circ}$  to the west (strike of  $170^{\circ}-220^{\circ}$ ) that represents a 4–9 km depth

difference between P and S piercing points. A  $\sim 5^{\circ}$  west-dipping Moho is what may be expected from the Bouguer gravity anomaly being more negative towards the northwest (Fig. 4.2). Moreover, a west-dipping Moho has the effect of increasing the apparent move-out of the Moho conversion for arrivals from the west and may be the reason for the mismatch with the predicted move-out for an assumed flat discontinuity in the slowness stack of Figure 4.12. Other viable explanations for the mismatch include: 1) a higher crustal Vp but slight thinner crust; or 2) a thicker crust but a lower Vp/Vs ratio than modelled.

Crustal and mantle anisotropy are inferred from P- and S receiver functions, respectively. P receiver functions suggest crustal anisotropy with a sub-vertical and slow symmetry axis that is interpreted as the sub-horizontal foliations of the Otago Schist.

More crustal complexity than modelled is suggested by the lack of coherence of the C and M pulses of the radial P receiver functions.

Both P and S receiver functions suggest presence of a discontinuity 15–20 km below the Moho. While P receiver functions suggest a Vp increase of 0.3 km/s at 50–55 km depth, S receiver functions suggest that this could represent a rotation of the anisotropy fast symmetry axis at 45–58 km depth.

# 4.6.3 Wanaka station (WKZ)

Similar to station EAZ, station WKZ is located near the structural axis of the Otago Schist metamorphic arch (Fig. 4.1). It is sited on the west flank of the Cardrona Valley, a structural basin bounded by two reverse faults with some dextral shear component: the northwest-dipping NW Cardrona Fault and the secondary southeast-dipping SE Cardrona Fault (Beanland and Barrow-Hurlbert, 1988).



WKZ P receiver functions

Figure 4.15: Station WKZ 1 Hz low-pass filtered P receiver functions as a function of back azimuth.

# Observations

Slowness and epicentral distance stacks are compiled for back azimuths in the range  $270^{\circ}-0^{\circ}$ , within which receiver functions sum up constructively.

The shallow structure of the Cardrona Fault has strong effects on station WKZ P receiver functions, which differ notably from those at stations EAZ and RPZ. In contrast, S and SKS receiver functions, whose wavelengths are greater and, in which multiples do not interfere with conversions, display many similarities with those at station EAZ and even RPZ (SKS receiver functions are discussed separately in Section 4.6.4). Following features are noted:

- (a) The direct P-wave pulse of the P receiver functions is negative for back azimuths west of -30° on the radial and at back azimuths from -90° to 50° on the transverse (see back azimuth stack of Fig. 4.15). In addition, the first Ps/Sp conversions (labelled C1 in Fig. 4.16) experience maximum delays of ~1 s at ca. -70° from north (Fig. 4.15). These two features are compatible with a shallow and low velocity layer dipping steeply to the west (Savage *et al.*, 2007a). Note that calculated delay times of C1 free surface multiples predict interference with conversions from the lower crust in the P receiver functions (Fig. 4.16).
- (b) A second crustal conversion (labelled C2 in Fig. 4.16), is apparent in S receiver functions as a weak pulse but not in P receiver functions. The reason for this may be interference of the conversion C2 with C1 multiples. A positive pulse with negative move-out and delay times of 10–11 s of the P receiver functions is a possible C2 PpPs free-surface multiple (Fig. 4.16).
- (c) Ps/Sp conversions at τ = 3–5 s (M in Fig. 4.16) have corresponding energy on the transverse P receiver functions (Fig. 4.15). A positive pulse with negative move-out at 14–15 s of the P receiver functions coincides with M's PpPs free-surface multiple for an assumed Vp/Vs ratio of 1.70 (Fig. 4.16).
- (d) The conversion at τ ~ 6–8 s (X in Fig. 4.16) displays a similar move-out in the S receiver functions to that identified in station EAZ S receiver functions (Fig. 4.12). In the P receiver functions for WKZ, however, X delay times coincide



Figure 4.16: Station WKZ slowness stack of P (6.8 s/°  $\leq p \leq 9$  s/°) and S (10 s/° s/°) radial receiver functions for back azimuths in the range 270°–0°. A 1 Hz low-pass filter was applied.*Blue and red wiggles*denote positive and negative amplitudes, respectively.*Color-coded curves*are Ps-Sp (*solid*) and free-surface multiple (*dashed*) delay times predicted for assumed average Vp, Vp/Vs and conversion depth H of 4.5 km/s, 1.85 and 4 km, respectively (C1), 6.2 km/s, 1.60 and 26 km (C2), 6.3 km/s, 1.70 and 36 km (M) and 6.9 km/s, 1.75 and 55 km (X).*Bars*on the right-hand side denote the number of events per bin of 0.4 s/°.

with a negative pulse for the northwest quadrant and a positive pulse for back azimuths from  $-20^{\circ}$  to  $40^{\circ}$  (Fig. 4.15). Thus, it is unclear if the negative pulse is a crustal multiple or represents a direct conversion.



Figure 4.17: Station WKZ 0.5 Hz low-pass filtered P receiver functions (*coloured wig-gles*) and model WP1M synthetic P receiver functions (*black curves*, Tab. 4.4).

- (e) Three consecutive pulses at 7–10 s of the P receiver functions (X' in Fig. 4.15) have polarities rapidly alternating with back azimuth (Fig. 4.15), suggesting that these pulses may represent some crustal reverberations. On the other hand, these pulses' strong amplitudes could be explained by a direct conversion and corresponding reverberations at the top structure. Note that the three pulses interfere into a broad single pulse in the 0.5 Hz low-pass filtered data of Figure 4.17.
- (f) The direct S-wave pulse is negative at slowness p > 14 s/° (Fig. 4.16) or epicentral distances Δ ≤ 45° and back azimuths east of −30° (Fig. 4.19), indicating the presence of an anisotropic layer with sufficiently large velocity contrast at the top.



Figure 4.18: Crust and mantle structure interpretation below station WKZ from S and P receiver functions (models WS1m/-M and WP1m/-P of Table 4.4, respectively.

## **P** receiver function synthetics

The discussed models are derived by trial and error. Direct P negative polarities west of  $-30^{\circ}$  on the radial component and Ps maximum observed delay times at  $\sim -70^{\circ}$  (see

Observations section a) are reproduced with a steeply west-dipping interface of strike  $190^{\circ} \pm 10^{\circ}$  overlain by low velocity material (models WP1m/-M of Fig. 4.18, synthetics of Fig. 4.17). A dip of  $40^{\circ} \pm 10^{\circ}$  W and a 5–7 km thickness are found for a Vp of 4–5 km/s. The uncertainty on the dip results from the trade-off between dip and impedance contrast: large contrasts yield shallow dips, small contrasts yield strong dips.

Moho delay times (see c) are reproduced with a 33–41 km thick crust, assuming an average Vp of 6.2–6.5 km/s for the entire crust in association with Vp/Vs ratio of 1.75–1.65 (models WP1m/-M of Tab. 4.4; Fig. 4.17).

Strong crustal anisotropy is suggested by large amplitudes at  $\tau \sim 4$  s of the transverse receiver function (see c) and is modelled by a sub-vertical and slow symmetry axis that is, however, weakly constrained (Fig. 4.18).

X and X' delay times are matched by 48–56 km and 68–76 km deep boundaries 15 km and 35 km below the Moho, respectively (models WP1m/-M of Fig. 4.18; Tab. 4.4). The succession of positive, negative and positive polarities for interpreted conversions M, X and X' in the northwest quadrant (see d, e and bottom of Fig. 4.17) is matched with a succession of three anisotropic layers with fast axes approximately perpendicular one to the other (synthetics of Fig. 4.17; Tab. 4.4). It is however uncertain if these symmetry axis rotations represent real structures as crustal reverberations for X and X' can not entirely be rejected. Discontinuity X' won't therefore be further commented on.

#### S receiver functions synthetics

A crustal thickness of 35–42 km is found assuming upper and lower crustal Vp of 6.0– 6.3 km/s and 6.8–7.0 km/s in combination with a Vp/Vs ratio of 1.75–1.65 (models



Figure 4.19: Station WKZ 0.5 Hz low-pass filtered S receiver functions (*coloured wig-gles*) and model WS1M synthetic SV receiver functions (*black curves*; Tab. 4.4).

WS1m/-M of Tab. 4.4, Fig. 4.19).

A shallow and steeply dipping layer (see a) as modelled from the P receiver functions poses a problem in the S receiver functions, as this precludes transmission of S energy into P for  $i_{inc} > i_{Pt}$ , which does not occur in this data set.

Similar to station EAZ S receiver functions, an anisotropic layer with strong contrast at the top is required to produce direct S-wave pulses with negative polarity at  $\Delta \leq 45^{\circ}$  (see f).

X delay times of 8–10 s (see d) are matched with a 52–62 km deep layer. Similar to modelled for station EAZ, the negative polarities on the radial component of the direct S pulse and conversion X for back azimuths of  $-90^{\circ}$  to  $0^{\circ}$  are matched with a rotation of the anisotropy symmetry axis from  $-20^{\circ}$  to  $20^{\circ}$  (models WS1m/-M synthetics of Fig. 4.19).

#### Interpretation and discussion

The shallow west-dipping interface is identified as the deep structure of the NW Cardrona Fault, whose surface trace is located < 1 km east from station WKZ. The strike of  $190^{\circ} \pm 10^{\circ}$  agrees with the NW Cardrona Fault orientation at the surface (Beanland and Barrow-Hurlbert, 1988). The modelled dip of  $40^{\circ} \pm 10^{\circ}$  is close to  $30^{\circ}$  as typically expected for an optimally oriented reverse fault. However, the depth inferred from the receiver functions is much larger than predicted from the less than 1 km distant fault trace, possibly as a result of lower wave speeds and a higher Vp/Vs ratio than used in the models or the Cardrona Fault zone extending deeper than expected.



Figure 4.20: Same as Fig. 4.16 for the assumed Vp, Vp/Vs and conversion depth of 6.2 km/s, 1.55 and 28 km (C2), 6.3 km/s, 1.55 and 49 km (M) and 6.9 km/s, 1.70 and 55 km (X). Note how the move-out of discontinuity M is better matched at large slowness than with a Vp/Vs ratio of 1.70. In addition, the large positive pulse at  $\tau \sim 14$ –15 s coincides with M's multiple PpPp delay times. However, there is no obvious pulse that coincides with PpPs delay times. Moreover, the move-out of the PpPp multiple predicted for a 48 km deep interface is stronger than that of the 14–15 s pulse, suggesting that the interface is shallower than 48 km.

Crustal thicknesses of 33–41 km and 35–42 km are estimated from P and S receiver functions, respectively. This crustal thickness is inconsistent with a –80 mGal Bouguer gravity anomaly and crustal thicknesses of 43–45 km and 48  $\pm$  4 km estimated from 3D inversion (Eberhart-Phillips and Bannister, 2002) and earthquake refraction (Section 2.4), respectively. Although the minimum crustal thickness estimate of 33 km is a reasonable value for a crust, this represents 3 km crustal thickening relative to a 30 km crustal thickness at the coast, and is inconsistent with the –80 mGal Bouguer anomaly at WKZ (Fig. 4.2). A 41 km crustal thickness, i.e. 11 km crustal thickening, produces a gravity anomaly that can match the observed Bouguer anomaly assuming an across-Moho density contrast of –450 kg/m<sup>3</sup> (Stern *et al.*, 2000). The 41 km crustal thickness is, however, insufficient for likely across-Moho density contrasts of –400 kg/m<sup>3</sup> or less (see Section 2.6).

A low Vp/Vs ratio of 1.60 or 1.55 would be necessary to reconcile delay times of the discontinuity M with a 43–45 km or 48 km crustal thickness, respectively (Fig. 4.20). Such a low Vp/Vs ratio could possibly result from anisotropy in the Otago Schist (Godfrey *et al.*, 2000). Figure 4.20 however suggests that Vp/Vs ratios  $\leq$  1.60 imply that the 14–15 s pulse corresponds with the discontinuity M PpPp multiple instead of the PpPs multiple, and that the PpPs multiple has no obvious corresponding pulse. Such a low Vp/Vs ratio therefore appears unlikely for the crust below station WKZ.

# 4.6.4 SKS receiver functions at stations EAZ and WKZ

SKS receiver functions of stations EAZ and WKZ share the same features, and are presented in a common stack (Fig. 4.22). They are also similar to RPZ's SKS receiver functions, but with a relative  $\sim 20^{\circ}$  counter-clockwise rotation of the polarity switches.



Figure 4.21: Earthquake distribution for SKS phases recorded at stations EAZ and WKZ.

- (a) Polarities of the direct SKS-wave pulse are anti-symmetric with respect to a back azimuth of about ~30° on the radial receiver functions, suggesting the presence of a steeply-dipping interface (see Section 3.3.1, p. 61);
- (b) Pulses up to ~8 s delay time are anti-symmetric with respect to ~60° from north and suggest that the cause of the polarity switches (dipping layer and/or anisotropy) could be deep seated;
- (c) Pulses at 1–2 s and 4–5 s are interpreted as crustal and Moho conversions, respectively.



Figure 4.22: Stations EAZ and WKZ 1 Hz low-pass filtered SKS receiver functions and model EWSKS (Tab. 4.5) SV receiver functions.

# Synthetics

Model EWSKS (Fig. 4.22; Tab. 4.5) is derived by trial and error and assuming an average Vp/Vs ratio of 1.70 for the crust. Polarities of the direct SKS-wave pulse on the radial and transverse components (see Observations section a) are matched with a ca.  $30^{\circ}$  E dipping and ~5 km deep interface. Crustal and Moho conversions are modelled with 22 km and 37 km deep boundaries, respectively. Positive pulses for the crustal and Moho conversions at back azimuths west of  $60^{\circ}$ , but negative in the southeast quadrant (see b), are reproduced with a Moho dipping  $10^{\circ}-15^{\circ}$  to the west. Although, the dipping Moho provides the strong move-out of Moho conversions between  $-80^{\circ}$  and  $10^{\circ}$  from north, the dip may be an overestimate. Thus, additional anisotropy

may be necessary to match the move-out.

## Interpretation

The 37 km crustal thickness derived from stations WKZ and EAZ SKS receiver functions represents an intermediate value between 29–35 km and 33–41 km derived from P receiver functions at EAZ and WKZ. A shallow east-dipping layer is a pervasive feature in SKS receiver functions models for stations RPZ, WKZ and EAZ. This feature was modelled to reproduce the negative polarities of the direct SKS-wave pulse. An east-dipping layer is, however, inconsistent with other models derived from P receiver functions. It seems unlikely that the different station locations RPZ, WKZ and EAZ are all sited on a shallow and east-dipping layer, when P receiver functions unequivocally indicate a west-dipping structure (at station WKZ for instance). Thus, the source to the common negative polarity of the direct SKS-wave pulse in SKS receiver functions could represent a more complex combination of dipping layer and anisotropy than has been explored in the synthetic tests of Chapter 3.

### 4.6.5 Jackson Bay (JCZ)

JCZ is located on the Alpine Schist, ~30 km south of the Haast river and ~4 km southeast of the Alpine Fault surface trace (Fig. 4.1). At the Haast river and south, the Alpine Fault is characterised by a small degree of fault segmentation, small uplift rates  $\leq 2.5 \text{ mm/yr}$ ,  $26 \pm 7 \text{ mm/yr}$  strike-slip (Cooper and Norris, 1995; Hull and Berryman, 1986; Sutherland and Norris, 1995) and localised deformation (Sutherland and Norris, 1995). The Alpine Fault is, hence, described as sub-vertical with mainly strike-slip movement. Furthermore, the surface trace trends at N59°E (Wellman, 1953). Seismic rays incoming from azimuths between  $-120^{\circ}$  and  $60^{\circ}$  are, therefore, expected to cross the Alpine Fault zone.

#### Observations

JCZ P receiver functions from  $\tau \geq 5$  s are dominated by reverberations (Fig. 4.23), which indicate the presence of one or more steeply dipping structures. These structures and seismic anisotropy are possibly reason for no coherent pulses in the SKS receiver functions (not shown) and polarity reversals of conversion pulses in the S receiver functions. The modelling of S and SKS receiver functions is, therefore, not attempted. Slowness and epicentral distance stacks are compiled for back azimuths within 270° and 0°.

- (a) The direct P-wave pulse experiences delays up to 0.5 s, suggesting low wave-speeds near the surface (Fig. 4.23).
- (b) Energy at τ ~ 0 s on the transverse P receiver functions with negative polarities within -130° and 40° from north, i.e. an azimuth range of 170°, suggests that polarities are anti-symmetric about an azimuth of ~60° as well as the presence of a west- to northwest-dipping discontinuity (see back azimuth stack of Fig. 4.24).


Figure 4.23: Station JCZ slowness stack of P ( $p \le 9 \text{ s/}^{\circ}$ ) and S (10 s/ $^{\circ} )$  $radial receiver functions for back azimuths in the range 270<math>^{\circ}$ -0 $^{\circ}$ . A 1 Hz low-pass filter was applied. *Blue and red wiggles* denote positive and negative amplitudes, respectively. *Color-coded curves* are Ps-Sp (*solid*) and free-surface multiple (*dashed*) delay times predicted for assumed average Vp, Vp/Vs and conversion depth H of 6.1 km/s, 1.70 and 20 km, respectively (C) and 6.2 km/s, 1.75 and 35 km (M). *Bars* on the right-hand side denote the number of events per bin of 0.4 s/ $^{\circ}$ .

(c) a Ps conversion at 2–3 s of the P receiver functions (labelled C in Fig. 4.23) has: (1) an anomalous move-out with maximum delay times for northwest back azimuths; (2) polarities reversing at ~−100° and ~20° on the radial component; and (3) anti-symmetric polarities with respect to ~−30° on the transverse com-



Figure 4.24: Station JCZ 1 Hz low-pass filtered P receiver functions and model JP1m synthetic P receiver functions (Tab. 4.6).

ponent, which all suggest a northwest-dipping layer (see P receiver function stack of Fig. 4.24). Positive pulses at 6–7 s and 8–9 s coincide with predicted  $\tau_{PpPp}$ and  $\tau_{PpPs}$  for C, respectively. The dipping layer may be the reason for the strong PpPp multiple.

- (d) A pulse at 3–4 s of the P receiver functions (M in Fig. 4.23) has same moveout, amplitudes and polarities as the C conversion, suggesting that this is not a multiple. Positive pulses at 10–11 s and 15–16 s, that display move-outs opposite to that of the M pulse correspond to PpPp and PpPs predicted delay times, respectively.
- (e) Similar to S receiver functions compiled for RPZ, EAZ and WKZ, the direct Swave pulse is negative at slowness > 13 s/° in the chosen back azimuth range of −90° to 0°, and suggests an anisotropic layer with strong impedance contrast at the top.

#### Synthetic P receiver functions

Models JP1m–JP2M (Fig. 4.25; Tab. 4.6) are derived by trial and error and also inversion with same parameters used for stations EAZ and WKZ receiver function modelling.

The direct P-wave pulse delays are modelled with a  $\sim 3$  km thick layer assuming Vp  $\sim 4$  km/s (see Observations section a).

Inversion suggests that C amplitudes and move-out (see c) and those of M are produced by an 18–23 km deep discontinuity dipping west at  $40^{\circ}-50^{\circ}$  (strike/dip of  $\sim 170^{\circ}/40^{\circ}$ ) overlying a southeast-dipping structure (strike/dip of  $\sim 50^{\circ}/45^{\circ}$ ; Fig. 4.25). Although the west-dipping discontinuity best matches C move-out, this does not reproduce the negative polarities outside the northwest quadrant and epicentral distances



Figure 4.25: Crust and mantle structure interpretation below station JCZ from P receiver functions (model JP1-2M; Tab. 4.6). The structure is projected onto the range of geometric raypaths of Ps Moho conversions from the northwest assuming a flat Moho. The *thin dashed curves* represent the extent of the Fresnel zone. The *thick and vertical dashed line* is the depth projection of a vertical Alpine Fault.

 $\leq 25^{\circ}$  (see c). More complex structures than modelled may be required. Crustal anisotropy of 7–10 % with slow symmetry axis steeply plunging to the northwest provides a fit to changing polarities at 1–2 s of the transverse receiver function (see b).

Two different models are derived by inversion, which well fit the negative polarities at  $\sim 3$  s that follow the C pulse and the positive polarities of the M conversion. In model JP1m (Tab. 4.6), the sequence of polarities is produced by anisotropic material overlying a  $\sim 28$  km deep and southeast  $\sim 45^{\circ}$ -dipping discontinuity. In model JP2M (Tab. 4.6), polarities are produced by the anisotropic material being underlain by an anisotropic layer with fast axis of symmetry dipping at  $\sim 50^{\circ}$  ca.  $20^{\circ}$  NW. The combination of a dipping discontinuity and anisotropy provides a good match to the receiver functions (synthetics of Fig. 4.26), and is the preferred model shown in Figure 4.25. The crustal thickness derived from models JP1m and JP2M is 30–39 km assuming upper and lower crustal wave speeds of 6.1–6.3 km/s and 6.8–7.1 km/s in combination



Figure 4.26: Station JCZ 1 Hz low-pass filtered P receiver functions (*coloured wiggles*) and model JP1-2M synthetic P receiver functions (*black curves*; Tab. 4.6). Model JP1-2M is the combination of models JP1 and JP2M

with Vp/Vs ratios of 1.75 and 1.65, respectively.

Note that the presence of steeply dipping layers weakens the amplitudes of reverberations in the synthetic receiver functions as the reverberations are refracted away from the station. This is contrary to the strong reverberations observed in the receiver functions (Fig. 4.24). Conversely, the presence of the top west-dipping fault is necessary in order to refract conversions from the steeply east-dipping fault back towards the station. Otherwise, the conversion is refracted away from the station, and is absent in the synthetic receiver functions.

#### Interpretation and discussion

Anisotropic material with a slow symmetry axis plunging to the northwest is interpreted to represent the southeast-dipping foliations of the Alpine Schist.

A ~40° W dipping structure at 18–23 km depth coincides with the depth projection of the Alpine Fault (Fig. 4.25), but is inconsistent with the Alpine Fault being near vertical and striking northeast (Section 1.4.1 13). Anisotropic material below the west-dipping structure may be associated with the Alpine Fault zone. A ~45° SE dipping discontinuity at ~28 km depth could represent a lower crustal boundary or be associated with the Alpine Fault.

Inconsistencies between modelled dips of boundaries and a near-vertical Alpine Fault may result from modelling in one dimension a structure, which is both laterally offset and near vertical.

model	label	depth	Н	ρ	$V_P$	$V_S$	$\delta P$	$\delta S$	trend	plunge	strike	dip	$V_P/V_S$
		(km)	(km)	$(\mathrm{kg}/\mathrm{m}^3)$	(m/s)	(m/s)	(%)	(%)	$(^{\circ})$	(°)	$(^{\circ})$	$(^{\circ})$	
RP1m		0	2	2600	4000	2286	-	-	-	-	-	-	1.75
		2	15	2700	6000	3429	-	-	-	-	235	10	1.75
	С	17	13	3000	6800	3886	-	-	-	-	200	20	1.75
	М	30	30	3300	7800	4457	-	-	-	-	150	15	1.75
	Х	60	50	3300	8300	4743	-	-	-	-	-	-	1.75
RP1M		0	20	2700	6200	3875	-	-	-	-	-	-	1.60
	С	20	16	3000	7000	4375	-	-	-	-	200	40	1.60
	М	36	30	3300	8100	4629	-	-	-	-	150	10	1.75
	Х	66	50	3300	8400	4800	-	-	-	-	-	-	1.75
RS1m		0	15	2700	6000	3529	-	-	-	-	-	-	1.70
	С	15	15	2900	6800	4000	-	-	-	-	190	10	1.70
	М	30	25	3300	7800	4457	7	5	50	0	-	-	1.75
	Х	55	65	3300	8300	4743	7	5	-20	0	-	-	1.75
RS1M		0	24	2700	6200	3875	-	-	-	-	-	-	1.60
	$\mathbf{C}$	24	19	2900	7000	4375	-	-	-	-	190	10	1.60
	М	<b>43</b>	30	3300	8300	4743	7	5	50	0	-	-	1.75
	Х	73	65	3300	8300	4743	7	5	0	0	-	-	1.75

Table 4.2: Station RPZ forward models derived from P- and S receiver functions. Numbers in bold are Moho depths. Note that mantle fast symmetry axes orientations are mostly unconstrained.

model	label	depth	Н	ρ	$V_P$	$V_S$	$\delta P$	$\delta S$	trend	plunge	strike	dip	$V_P/V_S$
		(km)	(km)	$(\mathrm{kg}/\mathrm{m}^3)$	(m/s)	(m/s)	(%)	(%)	$(^{\circ})$	$(^{\circ})$	$(^{\circ})$	$(^{\circ})$	
EP1m		0	10	2700	6000	3429	-	-	-	-	-	-	1.75
		10	13	2700	6000	3429	-10	-10	150	80	-	-	1.75
	С	23	6	3000	6800	3886	-	-	-	-	-	-	1.75
	Μ	29	20	3300	7800	4457	-	-	-	-	190	5	1.75
	Х	50	-	3600	8300	4743	-	-	-	-	-	-	1.75
EP1M		0	10	2700	6300	3818	-	-	-	-	-	-	1.65
		10	15	2700	6300	3818	-10	-10	150	80	-	-	1.65
	С	25	10	3000	7000	4243	-	-	-	-	-	-	1.65
	Μ	<b>35</b>	20	3000	8300	4743	-	-	-	-	190	10	1.75
	Х	54	-	3300	8600	4914	-	-	-	-	-	-	1.75
ES1m		0	17	2700	6000	3429	-	-	-	-	-	-	1.75
	С	17	13	3000	6800	3886	-	-	-	-	-	-	1.75
	Μ	30	15	3300	8300	4743	7	5	-20	0	-	-	1.75
	Х	45	-	3300	8300	4743	7	5	20	0	-	-	1.75
ES1M		0	23	2700	6300	3818	-	-	-	-	-	-	1.65
	С	23	15	3000	7000	4243	-	-	-	-	-	-	1.65
	Μ	38	20	3300	8300	4743	7	5	-20	0	-	-	1.75
	Х	58	-	3300	8600	4914	7	5	20	0	-	-	1.75

Table 4.3: Station EAZ forward models derived from P- and S receiver functions.

model	label	depth	Η	ρ	$V_P$	$V_S$	$\delta P$	$\delta S$	trend	plunge	strike	dip	$V_P/V_S$
		(km)	(km)	$(\mathrm{kg}/\mathrm{m}^3)$	(m/s)	(m/s)	(%)	(%)	$(^{\circ})$	$(^{\circ})$	$(^{\circ})$	$(^{\circ})$	
WP1m		0	5	2600	4000	2286	-	-	-	-	-	-	1.75
	C1	5	28	2600	6200	3543	-10	-10	-90	70	200	50	1.75
	М	33	15	3300	7800	4628	7	5	-40	0	-	-	1.75
	Х	48	20	3300	7800	4628	7	5	20	0	-	-	1.75
	Х'	68	-	3300	8300	4743	7	5	-30	0	-	-	1.75
WP1M		0	7	2600	4500	2727	-	-	-	-	-	-	1.65
	C1	7	34	2800	6500	3939	-	-	-	-	190	30	1.65
	М	41	15	3300	8300	4743	7	5	-40	20	-	-	1.75
	Х	56	20	3300	8300	4743	7	5	20	0	-	-	1.75
	Х'	76	-	3300	8600	4914	7	5	-30	0	-	-	1.75
WS1m		0	20	2700	6000	3429	-	-	-	-	-	-	1.75
	C2	20	15	3000	6800	3886	-	-	-	-	-	-	1.75
	М	<b>35</b>	17	3300	7800	4628	7	5	-20	0	-	-	1.75
	Х	52	-	3300	8300	4743	7	5	20	0	-	-	1.75
WS1M		0	25	2700	6300	3818	-	-	-	-	-	-	1.65
	C2	25	17	3000	7000	4242	-	-	-	-	-	-	1.65
	Μ	42	20	3300	8300	4743	7	5	-20	0	-	-	1.75
	Х	62	-	3300	8300	4743	7	5	20	0	-	-	1.75

Table 4.4: Station WKZ forward models derived from P- and S receiver functions.

model	label	depth	Η	ρ	$V_P$	$V_S$	$\delta P$	$\delta S$	trend	plunge	strike	dip	$V_P/V_S$
		(km)	(km)	$(\mathrm{kg}/\mathrm{m}^3)$	(m/s)	(m/s)	(%)	(%)	$(^{\circ})$	(°)	$(^{\circ})$	$(^{\circ})$	
RSKS		0	2	2600.	4000.	2424.	-	-	-	-	-	-	1.65
		2	15	2700.	6000.	3429.	-	-	-	-	330.	50.	1.65
	С	17	14	2900.	6800.	3886.	-	-	-	-	190.	10.	1.65
	М	<b>31</b>	-	3300.	8300.	4629.	7	5	40.	0	170.	20	1.75
EWSKS		0	5	2600	4000	2286	-	-	-	-	-	-	1.70
		5	17	2700	6200	3647	-	-	-	-	0	30	1.70
	С	22	15	2900	7000	4000	-	-	-	-	-	-	1.70
	М	37	20	3300	8300	4629	7	5	0	0	190	15	1.75
	Х	57	20	3300	8600	4914	7	5	40	0	-	-	1.75

Table 4.5: Stations EAZ and WKZ combined forward model derived from SKS receiver functions.

model	label	$\operatorname{depth}$	Η	ρ	$V_P$	$V_S$	$\delta P$	$\delta S$	trend	plunge	$\operatorname{strike}$	dip	$V_P/V_S$
		(km)	(km)	$(\mathrm{kg}/\mathrm{m}^3)$	(m/s)	(m/s)	(%)	(%)	$(^{\circ})$	$(^{\circ})$	$(^{\circ})$	$(^{\circ})$	
JP1m		0	3	2600	4000	2286	-	-	-	-	-	-	1.75
		3	5	2700	6100	3526	-	-	-	-	-	-	1.75
		8	10	2700	6100	3526	-10	-10	-30	70	-	-	1.75
	С	18	10	3000	6800	3886	-7	-7	-140	70	170	40	1.75
		28	2	3300	6800	3886	-	-	-	-	$51^{*}$	$43^{*}$	1.75
	Μ	30	-	3300	8300	4743	7.0	5.0	20	0	-	-	1.75
JP2M		0	3	2600	4000	2286	-	-	-	-	-	-	1.65
		3	8	2700	6300	3818	-	-	-	-	-	-	1.65
		11	12	2700	6300	3818	-10	-10	-30	70	-	-	1.65
	С	23	16	3000	7100	4303	-7	-7	-140	70	170	50	1.65
	Μ	39	-	3300	8300	4743	7	5	$19^{*}$	$53^{*}$	-	-	1.75
JP1-2M		0	3	2600	4000	2286	-	-	-	-	-	-	1.65
		3	8	2700	6300	3818	-	-	-	-	-	-	1.65
		11	12	2700	6300	3818	-10	-10	-30	70	-	-	1.65
	С	23	12	3000	7100	4303	-7	-7	-140	70	170	50	1.65
		35	4	3300	6800	3886	-	-	-	-	$51^{*}$	$43^{*}$	1.75
	Μ	39	-	3300	8300	4743	7	5	$19^{*}$	$53^{*}$	-	-	1.75

Table 4.6: Station JCZ forward models derived from P- and S receiver functions.

# 4.7 Discussion

The lithosphere beneath the South Island stations RPZ (Canterbury), EAZ and WKZ (western Otago) appears fairly uniform when imaged by S and SKS waves, but relatively heterogeneous in the light of P waves and crustal reverberations.

Beneath station RPZ, the top of the lower crust (C) is a seismic impedance boundary stronger than the Moho. A low Vp/Vs ratio of ~ 1.60 for the crust down to the top of the lower crust is suggested from crustal reverberation delay times. This Vp/Vs ratio is consistent with the value of 1.56 estimated for the upper crust from 2D modelling (Pulford, 2002) along SIGHT Transect 2 and average crustal Vp/Vs ratios of 1.65 from reflection data (Kleffmann *et al.*, 1998) and 1.67 from P receiver functions on wide-angle reflections (Galve *et al.*, 2002). Such a low Vp/Vs ratio was interpreted by Stern *et al.* (2001) as the indication of high-pore fluid pressure in the mid to lower crust.

A discontinuity is imaged at  $34 \pm 5$  km (JCZ) and  $32 \pm 3$  km (EAZ) on either side of the Southern Alps and at a depth of  $39 \pm 4$  km beneath the highest topography of the southern South Island (northwest from station WKZ). SKS receiver function modelling for stations EAZ and WKZ suggest a similar depth of ~37 km for the discontinuity. The nature of the discontinuity is uncertain. The Moho interpretation is inconsistent with the -80 mGal Bouguer and the -10 mGal isostatic anomalies as well as previous crustal thickness estimates of 43–45 km (Eberhart-Phillips and Bannister, 2002) and  $48 \pm 4$  km (Section 2.4) near station WKZ. Low Vp/Vs ratios of 1.60 and 1.55 are necessary to reconcile the observed delay times with the crustal thickness estimates of 43–45 km and 48 km, respectively (black dashed line in Fig. 4.27). Delay times of free-surface multiples and inversion of travel-time data (Eberhart-Phillips and Bannister, 2002), however, both suggest a minimum Vp/Vs ratio of 1.65 for this region



Figure 4.27: Otago cross section through stations JCZ, WKZ and EAZ (Fig. 4.2). Top: mean elevations in a 10 km wide swath and Geonet stations as triangles. Bottom: error bars denote depth ranges of major discontinuities as inferred from P- (thick bars) and S receiver functions (thin bars) at Geonet stations JCZ, WKZ and EAZ assuming crustal Vp/Vs ratios of 1.65–1.75 (Tab. 4.7). Error bar positions are mean conversion depth locations projected onto the profile. Blue bars represent crustal discontinuities (C), red bars the Moho or a lower crustal boundary (M) and green bars a mantle discontinuity (X) (Tab. 4.7). The black bar denotes the Moho depth range inferred from earthquake refraction (Section 2.4). The dashed line represents the Moho estimate for a Vp/Vs ratio of 1.55. The grey curve represents the Moho profile after Eberhart-Phillips and Bannister (2002). The arrow labelled "centre of gravity low" points to the Bouguer gravity anomaly minimum.

rendering such a low Vp/Vs ratio unlikely. Assuming that the discontinuity is located in the lower crust implies a weak or gradual contrast at the Moho. The hypothesis of a gradational Moho was tested by comparing conversion amplitudes using filters with varied low-pass corner frequencies (Owens and Zandt, 1985), but proved negative. Partial eclogitisation in the lower crust can weaken the velocity contrast at the Moho.

Conversions on S and P receiver functions suggest the presence of a mantle discontinuity (X) at 15–30 km below the discontinuity M of western Otago (green bars in Fig. 4.27). Such a discontinuity is also seen in station RPZ receiver functions, and is modelled as a 60–66 km deep boundary located 25–30 km below the Moho. A similar 45–70 km deep mantle discontinuity has been interpreted from positively polarised Ps conversions on P receiver functions in various locations of the eastern and southeastern South Island (Spasojevic and Clayton, 2008). A minimum wave-speed contrast of 0.3 km/s and/or rotation of the mantle anisotropy fast symmetry axis may be associated with the discontinuity. More events and a more complete back azimuth coverage are necessary to interpret this mantle discontinuity and to determine if the mantle discontinuities inferred to exist beneath western Otago and station RPZ are the same feature.

Mantle discontinuities in the depth range of 60–90 km have been attributed to partial eclogitisation of subducted crust (Bostock, 1998) and more often to the transition from spinel to garnet peridotite (e.g. Bostock, 1997; Hales *et al.*, 1968, 1975; Levin and Park, 2000), the so called Hales discontinuity (Hales, 1969). The Hales discontinuity has been interpreted in varied geotectonic environments in both oceanic and continental settings (e.g Levin and Park, 2000). The discontinuity is generally characterised by a ~0.3 km/s increase in P-wave speed over a 10–20 km transition at 60–90 km depth and possibly most of the time by changes in seismic anisotropy. The change in rheology and volume associated with a phase transition has been proposed as a mechanism for "transitional plasticity" (Sammis and Dein, 1974) and the development of an anisotropic layer in the mantle (Bostock, 1997; Levin and Park, 2000). Such an anisotropic layer could allow mechanical decoupling that would result in a change in the symmetry axis of the anisotropy.

Table 4.7: Depths (in km) to discontinuities C, the top of the lower crust, M, the Moho, X and X' upper mantle discontinuities as derived from P- and S receiver functions (Tab. 4.2-4.6).

Discontinuity	RPZ		JCZ		WKZ			EAZ		
	Р	$\mathbf{S}$	Р	$\mathbf{S}$	Р	$\mathbf{S}$	Р	$\mathbf{S}$		
С	17-20	15-24	18-23	-	-	20-25	23-25	17-23		
М	30-36	30-43	30-39	-	33-41	35 - 42	29 - 35	30-38		
X	60–66	55 - 73	-	-	48-56	52 - 62	50 - 54	45 - 58		
Χ'	-	-	-	-	68–76	-	-	-		

# 4.8 Conclusions

- The top of the lower crust beneath Geonet station RPZ (eastern central Southern Alps) represents a seismic impedance contrast stronger than the Moho.
- 2) A low upper-crust Vp/Vs ratio of ~1.60 beneath RPZ corroborates previous low Vp/Vs ratio estimates (Galve *et al.*, 2002; Kleffmann *et al.*, 1998; Pulford, 2002) interpreted as the effect of high-pore fluid pressure released in the mid-crust during prograde metamorphism.
- 3) A discontinuity is imaged at 34 ± 5 km (JCZ) and 32 ± 3 km (EAZ) depth on either sides of the Southern Alps and reaches 39 ± 4 km depth (ca. 50 km northwest of WKZ) beneath the highest topography, assuming crustal Vp/Vs ratios of 1.65–1.75. The interpretation of this discontinuity is uncertain. A Moho depth of 39 ± 4 km is inconsistent with crustal thickness estimates of 48 ± 4 km from earthquake refraction (Section 2.4) and ~45 km from 3D inversion of travel-time data (Eberhart-Phillips and Bannister, 2002) and a -80 mGal Bouguer anomaly at WKZ. A crustal Vp/Vs ratio of 1.55 would be necessary to reconcile the Moho interpretation with the gravity and previous crustal thickness estimates. A lower crustal boundary may represent an alternative interpretation for this discontinuity that, however, implies a low wave-speed contrast at the Moho.
- 4) A mantle discontinuity at 50–70 km depth (15–30 km below the lower crustal discontinuity or the Moho) is suggested by S and P receiver functions. This discontinuity could represent a change in seismic fast propagation orientation and possibly an additional Vp increase of ~0.3 km/s that may be interpreted as mechanical decoupling, or partial eclogitisation or the Hales discontinuity, respectively.
- 5) The modelling of P receiver functions of Geonet station WKZ (south of Wanaka)

suggests the presence of a  $40^{\circ} \pm 10^{\circ}$  west-dipping and ~6 km deep discontinuity striking at  $190^{\circ} \pm 10^{\circ}$  from north that is consistent with the NW Cardrona Fault.

6) The restricted number of events and back azimuth range in the Geonet data in combination with the presence of complex structures and anisotropy in the crust represents a strong limitation in interpreting the receiver functions. A few more years of data will be beneficial for the interpretation of receiver functions and for constraining the extent and nature of the mantle discontinuity in this region.

# Chapter 5

# P-wave travel-time delay analysis

# 5.1 Abstract

The geometry and wave-speed anomaly of an upper mantle high-speed body centred beneath the Southern Alps are tested with P travel-time delays from a wide range of back azimuths. This body possibly represents thickened (Molnar, 1992) or subducted lithosphere (Mattauer, 1986).

Two portable arrays, COOK and WCOAST, which combined into a ~80 km long profile, were deployed along the western portion of the previous SIGHT Transect 2. This chapter presents P travel-time delays measured from 35 teleseisms recorded along these two consecutive deployments. These events, along with four 1996 teleseisms recorded along SIGHT Transect 2, are used to investigate the geometry and amplitude of the mantle wave-speed anomaly.

Travel times are calculated through a three-dimensional velocity grid, based on Scherwath et al.'s (2003) two-dimensional crustal model for SIGHT Transect 2 extrapolated in the third dimension at the azimuth of the South Island plate boundary.

Depending on the ray azimuths and incidence angles, travel times show arrivals 0.3 s to 1.8 s earlier than predicted for the presence of the Southern Alps crustal structure alone. The travel-time advance appears to increase as teleseismic rays have large incidence angles at the Moho and raypaths which are close to being sub-parallel to the azimuth of the Australian-Pacific plate boundary. The negative delays and variations with back azimuth are consistent with the presence of a high-speed anomaly in the lithospheric mantle that has a horizontal cylindrical shape, whose axis is subparallel to the plate boundary. A sub-vertical, 100–130 km wide and 0.5–0.6 km/s P wave-speed anomaly centred at 100–110 km depth is a best average fitting model for the mantle body for most back azimuths.

This model, however, is insufficient to explain large travel-time advances from events located to the north with shallow and  $\sim 50^{\circ}$  incident rays. These delays can not be explained by a crustal root thinning to the northeast alone, but require a mantle body with a minimum wave-speed anomaly of ca. 1.1 km/s. In contrast, time delays for back azimuths of  $-50^{\circ}$  to  $-60^{\circ}$  from north have smaller amplitudes and require a wave-speed anomaly of ca. 0.3 km/s only. Mantle anisotropy is therefore proposed as a possible explanation for the discrepancy.

# 5.2 Introduction

Lithospheric shortening by intra-continental subduction (Mattauer, 1986) or uniform thickening (Molnar *et al.*, 1999) causes a down-warp of isotherms in the mantle lithosphere. The resulting temperature anomaly produces a positive wave-speed anomaly (see Section 1.3, p. 8). As teleseismic arrivals propagate through such a wave-speed anomaly, they accumulate time advances (negative travel-time delays) relative to arrivals that do not propagate through the anomaly. These time delays can be observed on station arrays located at the Earth's surface. Their variations with location can help to constrain the extent, amplitude and geometry of the wave-speed anomaly.

Two studies have imaged a high wave-speed anomaly beneath the Southern Alps (Kohler and Eberhart-Phillips, 2002; Stern et al., 2000). Using forward modelling of teleseismic P travel-time delays, Stern *et al.* (2000) found a sub-vertical,  $\sim 100$  km wide and  $\sim 120$  km deep wave-speed anomaly of 0.5–0.6 km/s beneath the central Southern Alps. The time delays were measured along the SIGHT Transects 1 and 2 with 2–5 km station spacing, and stemmed from three teleseisms located northwest of New Zealand only. Kohler and Eberhart-Phillips (2002) imaged a 0.2–0.4 km/s wave-speed anomaly using 3D inversion of teleseismic travel-time delays. Time delays were measured on the larger but sparser SAPSE array than SIGHT. The SAPSE station spacing was  $\sim 50$  km, which is approximately half the width of the putative wave-speed anomaly. Although the number of teleseisms and the resulting back azimuth range were both larger in the SAPSE than in the SIGHT study, most teleseisms stemmed from the northwest as well. Forward modelling on a densely spaced SIGHT array (Stern et al., 2000) could possibly resolve the full amplitude of the wave-speed anomaly for northwest back azimuths. In contrast, resolution tests showed that 3D inversion using the SAPSE array could resolve only 40 % of the anomaly (Kohler and Eberhart-Phillips, 2002). Moreover, both experiments suffered from a lack of earthquakes from the southeast, so that they could not constrain the dip of the high wave-speed mantle body. For example, the 3D inversion resolution tests showed that the uneven distribution of teleseisms resulted in a bias of  $\sim 20^{\circ}$  NW on the dip of the wave-speed anomaly (Kohler and Eberhart-Phillips, 2002).

This chapter complements these two studies by analysing teleseismic travel-time delays measured on a densely spaced array from a wider range of back azimuths than used by Stern *et al.* (2000).

#### 5.3 Data

#### 5.3.1 COOK and WCOAST deployments

The whole experiment consisted of two consecutive seismograph deployments (Fig. 5.1). During the COOK phase, six three-component short-period seismometers were deployed for a period of four months along a profile extending from Mt Cook Village (MCV) located east of the Main Divide, to Braemar station (BRM) on the east shore of Lake Pukaki. For the WCOAST phase that followed the COOK deployment, all stations but MCV were transferred to the west of the Main Divide and re-deployed for five months. An additional broad-band instrument was used, which increased the number of sites to seven. The six new sites were located between Gillespie's Beach (GIL) at the west coast and the upper Copland Valley (UCV). The seventh site at MCV was kept in both deployments in order to provide continuity between P travel-time delays measured from the COOK and WCOAST deployments. The COOK and WCOAST deployments, combined, represented a ca. 80 km long profile extending along the previous SIGHT Transect 2 and approximately perpendicular to the Southern Alps axis. The array was composed of Reftek 130 data acquisition systems connected to 1 Hz short-period L4 Mark seismometers as well as one broad-band Guralp-40T at the Karangarua Quarry (KAQ) site of the WCOAST array.

#### 5.3.2 Selected phases

Selected phases include not only first arrival P phases but also Pn, prominent pP and sP and PcP, PKP, PKiKP phases. The proximity of rivers, coast and the distance to hard rock basement render data collected at the WCOAST sites of generally lower quality than those from the COOK deployment.



Figure 5.1: Topographic map of the region of study (black rectangle on top left insert). *Red triangles*: WCOAST/COOK recording sites; *black triangles*: SIGHT Transect 2 deployment. Coordinates are in meters from the New Zealand Map Grid.



Figure 5.2: Distribution of recorded teleseisms; *red triangle*: array; *colored circles*: selected COOK and WCOAST hypocentres; *colored stars*: SIGHT hypocentres.

Travel-time measurements are attributed qualities from a scale of A to D estimated visually. Qualities from A to D denote measurements, which:

- are unequivocal and have a signal-to-noise ratio appreciated to be greater than 10 (A);

- are still unequivocal but have a signal-to-noise ratio estimated in the range 3–10 (B);

- are of similar quality to B and have one arrival pick that is ambiguous (C);

- have two or more arrival picks that are ambiguous and have a signal-to-noise ratio less than 2 (D).

Due to the short three to four months deployment period for each array, both deployments failed to record quality-A and -C phases from the southeast, i.e. off the South American coast. Nevertheless, selected events (Fig. 5.2, Tab. C.3) cover a wide range of back azimuths from  $-80^{\circ}$  to  $40^{\circ}$  from north and from the western Pacific. Moment magnitudes are  $M_W > 5.5$  and epicentral distances are in the range  $10^{\circ}-95^{\circ}$ . Seventeen phases (+ two additional Pn phases) and eighteen phases (+ one Pn phase) were selected respectively for the COOK and WCOAST deployments. These include quality-A and -C measurements as well as two quality-D measurements for the two best phases from the southeast (events 30, 35 of Tab. C.3).

#### 5.3.3 SIGHT Transect 2

Four teleseisms of the 1996 SIGHT Transect 2 deployment (Stern *et al.*, 2000) are presented to supplement the WCOAST and COOK deployments. Three are from the northwest Pacific, and one is from off the Chilean coast (Fig. 5.2, Tab. C.1). These events were recorded with  $\sim$ 100 Mark-L22 2 Hz short-period instruments. Measured travel-time delays span along a 100–160 km distance across the central Southern Alps in comparison to the only 80 km long WCOAST/COOK array. They provide necessary travel-time information east of the COOK array and have the advantage that rays sample the entire width of the Southern Alps lithospheric mantle. Some gaps exist in the data that correspond to the end of the deployment period.

## 5.4 Method

#### 5.4.1 Instrument response correction

Because the majority of seismometers on the WCOAST/COOK array were Mark-L4C 1 Hz short-period instruments with a single Guralp-40T broad-band instrument (KAQ) and because of the proximity of the coast with associated microseismic sea noise below 0.5 Hz, the Guralp's response was converted into an L4C's response. This has the advantage of avoiding the enhancement of long-period noise in the corrected short-period records. For a few events with dominant long periods in the signal, the L4C's response was converted into a Guralp's response.

# 5.4.2 Adaptive stacking

Travel-time delays were measured by an automatic correlation method, called adaptive stacking (Rawlinson and Kennett, 2004). The initial alignment of a chosen phase across the individual traces is achieved prior to correlation by correcting for the moveout shift as determined from the ak135 (Kennett *et al.*, 1995) travel-time predictions (top of Fig. 5.3–5.4).

The traces are stacked into a reference trace, the "linear stack", and its square, the "quadratic stack", in a user-specified window (the top two traces labelled zscp and zssl in Fig. 5.3–5.4). This reference trace is correlated with each individual trace in order to determine the relative time shift. Each individual trace is corrected for its respective time shift before new linear and quadratic stacks are computed. Successive iterations through the comparison of the reference trace with each individual trace and minimization of the quadratic stack result in a sharpening of the stacked waveform



asi03.282.22.aq 13.76 119.94 33.0km 9/10/2003 22:30:38.00

Figure 5.3: Quality-A event: Mw 6.2 earthquake from the Philippines that was recorded at stations MCV to BRM (event 9 of Tab. C.3). *Top*: initial alignment of the P phase after move-out correction. *Bottom*: final alignment after adaptive stacking. Traces from top to bottom represent the squared stack of single traces in a 1 s stacking window (zscp), the linear stack of single traces in the stacking window (zssl) and single traces (MCV to BRM).

œ zss MCV ø UCV station number KCJ MAI KAQ 2 GIL 0 5 10 15 20 25 time(s) asf04.003.16.aq -22.30 169.66 9.4km 3/ 1/2004 16:27:58.01 zscp œ zssl MCV ø station number UCV KCJ A MAI KAQ 2 GIL Q 5 10 15 20 25 time(s)

asi04.003.16.aq —22.30 169.66 9.4km 3/ 1/2004 16:27:58.01

Figure 5.4: Quality-B event: Mw 7.1 earthquake from the Loyalty Islands region that was recorded at stations GIL to MCV (event 21 of Tab. C.3). *Top*: initial alignment of the pP phase after move-out correction. *Bottom*: final alignment after adaptive stacking. The stacking window is 1 s long.

(compare top and bottom of Fig. 5.3 or 5.4).

The user determines the beginning and the length of the stacking window, and a maximum allowed number of iterations. A window length of 1–2 s, half to a full period, proved the best for alignment. Stability of the solution, i.e. that of the alignment and constant time shifts, is dependent upon the similarity of the waveform across the array. In the case of good-quality measurements, stability is reached after three iterations. Relatively good alignment but a non stable solution usually results from the waveform varying across the array.

Cycle skipping may occur when the travel-time delay of a single trace relative to the reference trace approaches the signal period because the adaptive-stacking method tends to correlate the most impulsive features between single traces. This often leads to a stable solution after a low number of iterations but a wrong alignment. In such a case, manual picking was used instead.

#### 5.4.3 Static corrections for the WCOAST and COOK data

Travel-time corrections, so called statics (Tab. C.5), were applied to account for elevation differences and variable geology across the array. Elevations were reduced to an elevation of 0 m using SIGHT Transect 2 shallow structure information (Kleffmann, 1999, p. 145). The static corrections were then calculated relative to that at station MCV, i.e. the static correction at MCV was subtracted.

#### 5.4.4 Processing of SIGHT Transect 2 data

SIGHT Transect 2 data were processed with the Claritas seismic package (Ravens, 2008). The first arrival or a later prominent phase that could be better correlated through the entire array was picked. Elevation statics and move-out corrections were applied using a reducing wave speed of 5.0 km/s and ak135 predicted travel times (Kennett *et al.*, 1995), respectively.

#### 5.4.5 Forward modelling

The resolution of wave-speed anomalies through tomographic inversion not only relies on the number of rays that cross-cut the model cells, but also on the back azimuth distribution of events. Uneven distributions will bias the model by distributing the wave-speed anomalies to cells where most of the ray-paths lie. New Zealand's situation is for this reason relatively inappropriate for tomographic inversions because most events that can be used are located along the western Pacific. Only a few events located east and southeast of New Zealand can be used. With a larger data set than used here, Kohler and Eberhart-Phillips (2002) showed that 3D inversion can only partially resolve the dip and amplitude of wave-speed anomalies. For this reason, forward modelling was preferred for this study.

The Southern Alps crustal structure is assumed to be two dimensional; i.e. uniform along the plate boundary. However, comparison of the crustal structures derived along the SIGHT Transect 1 (Van Avendonk *et al.*, 2004) and Transect 2 (Scherwath *et al.*, 2003) suggests that the Moho is shallowing at a  $\sim 6^{\circ}$  angle along-strike and to the northeast, based on maximum reported crustal thicknesses of 37 km and 44 km on Transect 1 and Transect 2, respectively. Below, the maximum error in calculated travel times is estimated that results from assuming a uniform crustal thickness along-strike. Rays of the closest earthquakes, which arrive from the northeast, will cross-cut the Moho at a maximum distance of 25 km from the profile line. Assuming a Moho shallowing at a  $\sim 6^{\circ}$  angle to the northeast, these rays are expected to sample a  $\sim 3$  km thinner crust than rays arriving along the profile. Assuming vertical raypaths and that mantle rocks of  $v_M \sim 8.1$  km/s replace lower crustal material with  $v_{LC} \sim 6.8$  km/s where the crust is thinner, the error amounts to  $\Delta h(v_{LC}^{-1} - v_M^{-1}) = 0.07$  s. i.e. less than 0.1 s. Thus, assuming a non-dipping Moho along-strike can lead to maximum possible error of < 0.1 s on calculated delay times, assuming that mantle rocks replace lower crustal material where the crust is thinner. Thus, the 2D assumption appears reasonable.

Forward modelling in two dimensions is sufficient as long as rays are arriving subparallel to the profile orientation, that is sub-perpendicular to the strike of the plate boundary. For rays arriving at a 20° angle from this azimuth, small differences in traveltimes are noticed. Beyond a ca.  $45^{\circ}$  angle to the profile orientation, i.e. in the case of rays arriving from an orientation sub-parallel to the strike of the Southern Alps, ray focussing and de-focussing through three-dimensional structures have major effects. For instance, because rays preferentially sample faster material, high wave-speed anomalies that stretch along the mountain axis will cause major refraction effects. Those rays, which are refracted through the wave-speed anomaly, will accumulate larger time advances than rays that propagate from the perpendicular direction. Modelling such time advances in a 2D depth section will result in mapping the wave-speed anomaly into a stronger anomaly than the real one. Moreover, the refraction angle of a phase arriving sub-perpendicular to the profile line and the 2D depth section will map into a refraction angle that is steeper (smaller) than in reality. As a consequence, the calculated raypath and travel time will be shorter than the true ones. This will result in inconsistencies between models that are derived from different event back azimuths. Because of the almost non-existent ray coverage from the southwest and southeast, 2D structures are modelled only. However, given the wide range of ray azimuths relative to the array, 3D forward modelling was use to model these 2D structures.

A 3D finite difference modelling code that implements the fast marching method (or FMM Rawlinson and Sambridge, 2005) is used to calculate synthetic travel-time delays (also called residuals) relative to ak135 predicted travel times (Kennett et al., 1995). The travel-time field is first calculated at the base of the model grid (Tab. 5.1) using the ak135 predicted travel times. The field is then calculated for the entire grid using the fast marching scheme to propagate the first-arrival wavefront. In the fast marching method, the wavefront is a narrow, one-node wide, band of grid points that separates "alive" grid points (points, through which the wavefront has already propagated) from "far" grid points (points, which are not alive yet). The FMM calculates trial travel times for all points within the narrow band, and uses a heap-sort algorithm to sort these travel times and to find the grid point with minimum travel time. The grid point with minimum travel time is made alive and is replaced by the next far point in the narrow band, allowing the narrow band to propagate a further step. This process is repeated until all grid points have become alive. Speed and stability are strengths of the method (Rawlinson *et al.*, 2006). One limitation of the program version that is used is that teleseismic rays must enter the grid through the bottom. Thus, the grid has to be made large enough to include all rays, and Moho refractions (Pn) are excluded (events 36–38 of Tab. C.3).

The three-dimensional wave-speed models are derived by rasterising the SIGHT Transect 2 model (Scherwath *et al.*, 2003; Stern *et al.*, 2000) and extrapolating this at a chosen azimuth (Fig. 5.5 and 5.6). The crustal structure is kept the same, but the mantle body is modelled by various horizontal cylindrical shapes that represent the high wave-speed anomaly caused by mantle shortening and the resulting downwarp



Figure 5.5: 3D velocity model Sed0-3km (Tab. 5.1) of the top 2.5 km of the crust taken from Kleffmann (1999) and extrapolated at an azimuth of N45°E. *Top*: map view at 0.5 km depth. *Bottom*: Depth section along the SIGHT Transect 2 with azimuth of N135°E. Distances are relative to station MCV location. Red and green triangles represent SIGHT T2 and COOK/WCOAST recording sites, respectively.



Figure 5.6: 3D velocity model SUBE (Tab. 5.2). The crust is taken from Scherwath et al. (2003) and the mantle wave-speed anomaly is modelled by a horizontal cylindrical shape. Both are extrapolated at an azimuth of N55°E. Top: map view at 110 km depth. Bottom: depth section at an azimuth of 135°. Distances are relative to station MCV location. azN represents the cylinder azimuth from north, aV is the inclination from the vertical (i.e. 90° from the dip), H is the height, z is the depth of the centre, W is the width and F is a factor between 0 and 1 that represents the fraction of the cylinder radius, to which Vp is constant. Beyond the radius fraction to the edge of the cylinder Vp decreases linearly down to that of the surrounding mantle (1=body with constant Vp+ $\delta$ Vp, 0=body with constant Vp gradient).

of mantle isotherms (e.g. Fig. 1.5). Symmetric bodies simulate distributed thickening (Molnar *et al.*, 1999; Stern *et al.*, 2000), while dipping bodies simulate intra-continental subduction (Mattauer, 1986, Fig. 5.6). The cylindrical shapes are defined by a centre (latitude, longitude and depth), a height H, a width W, a dip, an azimuth from north azN, a wave-speed anomaly  $\delta Vp$ , and a factor F between 0 and 1, that represents the fraction of the cylinder radius, to which Vp is constant. Beyond the radius fraction to the edge of the cylinder, Vp decreases linearly down to that of the surrounding mantle (1=body with constant  $Vp + \delta Vp$ , 0=body with constant Vp gradient).

In order to save computing time and space and to retain the fine structure of the top sedimentary basins, modelling is done in two stages. (1) Travel-time delays are calculated through the top 2.5 km of the crust (Fig. 5.5) using a grid of 0.3 km and 0.5 km node spacing with depth and latitude/longitude, respectively (Tab. 5.1). The velocity structure is taken from Kleffmann (1999), and is slightly modified for the western stations of the WCOAST/COOK array in order to account for different basement rocks. (2) Travel-time delays are then calculated through the crust (Scherwath *et al.*,

Table 5.1: Velocity grid origin (oz, olat, olon), spacing (dz, dlat, dlon) and number of vertices (nz, nlat, nlon) used in the FMM (Rawlinson and Sambridge, 2005) to model the crustal root (Fig. 5.5) and the mantle body (Fig. 5.6). Note that depths in the program are negative below sea level. Thus, oz positive means that the grid origin is located above sea level.

oz (km)	olat (°)	olat (°)	dz (km)	dlat (°)	dlon (°)	nz	nlat	nlon					
top 2.5 km of the crust													
0.5	-43.2	169.5	0.3	0.009000	0.012405	10	165	160					
	crust-mantle grid												
1.5	-41.9	167.8	2	0.018013	0.0.024811	109	176	176					

2003) and a range of mantle wave-speed anomalies using a grid with 2 km node spacing (Fig. 5.6; Tab. 5.1). Both sets of synthetic time delays are summed to produce the total travel-time delays. Comparison between ray-tracing and the FMM predicted travel-time delays calculated in two stages as explained above shows good agreement (Fig. 5.7) and suggests that the node spacing is sufficient.



Figure 5.7: Comparison between MacRay (*black dashes*) and FMM (*blue dashes*) predicted travel-time delays for SIGHT Transect 2 crustal structure (Scherwath *et al.*, 2003) and picks from Stern et al. (2000) (*blue error bars*) and this study (*orange error bars*). FMM predicted travel-time delays are the sum of two sets of time delays: 1) those calculated through a fine velocity grid representing the top 2.5 km of the crust and 2) those calculated through a coarser grid than the first one that represents the entire crust.
## 5.5 Relative P travel-time delays

Travel-time delays were measured along the WCOAST/COOK array relative to station MCV at the centre of the array, and along SIGHT Transect 2. These delays consistently show positive values near station MCV (kilometer 0 in Fig. 5.8a–d) and the Southern Alps highest elevations (Fig. 5.1). This is expected from the presence of a crustal root beneath the Southern Alps, through which rays are delayed.

However, as observed by Stern *et al.* (2000) for rays arriving from the northwest, travel-time delays decrease to the southeast of the array more rapidly than predicted from the presence of the crustal root. This discrepancy is apparent in Figure 5.8 for three teleseisms from the northwest (already presented by Stern *et al.*, 2000) and one teleseism from the southeast. A difference of 0.3 s (Fig. 5.8a) to 0.6 s (Fig. 5.8c) and possibly as much as 1 s (Fig. 5.8b) exists ~60 km east of MCV between the observations and the predicted effect of the crust. The mismatch is also apparent for measured travel-time delays of the events recorded on the WCOAST and COOK arrays (Fig. 5.9a–f).

A small set of COOK and WCOAST travel-time delays are selected and presented in groups of similar back azimuth and incidence angle,  $i_c$  (Fig. 5.9a-f):

- steeply incident PKPdf and PKiKP phases from the northwest (events 31 and 33);
- 2) P and pP phases from the northwest (events 5, 9, 18 and 19) that are comparable to rays from the Banda Sea earthquake recorded along the SIGHT Transect 2;
- steeply incident ScP and PcP phases from the north northwest (events 1 and 28) that have a similar back-azimuth with that of the Honshu event but a 5° steeper incidence angle;



Figure 5.8: Orange circles and error bars represent travel-time delays from the Banda (a), Irian Jaya (b), Honshu (also shown in Fig. 5.7)(c) and Chile (d) earthquakes recorded along SIGHT Transect 2 (Fig. C.1-C.4). Blue dashes are calculated delay times for SIGHT Transect 2 crustal structure (Scherwath *et al.*, 2003) only. These are time shifted to align with picks of the western (a–c) or eastern (d) stations. Note the uncertainty in the alignment in (b) as a result of noise in traces west of kilometer 0 (10–40 km in Fig. C.3). Observed delay times are corrected for elevation assuming a reducing wave speed of 5 km/s and are displayed relative to the average. Distances are relative to site MCV. The Alpine Fault surface trace passes at kilometer -25 (see a).

- 4) sP and pP phases from the north and shallow incident (events 17 and 21);
- 5) shallow incident P and sP phases from the northeast (events 22 and 32);
- 6) P and pP phases from the east and southeast (event 34 and 35) that have back azimuths within 20° and incidence angles within 8° of that of the Off-coast Chile event.

#### Effect of the crustal root orientation

Minimum and maximum possible strike directions for the crustal root of N45°E and N55°E are assumed from the orientations of the Bouguer gravity anomaly and the Alpine Fault, respectively. The closest events with north and northeast back azimuths, which are close to but not exactly the crustal root orientation, and which have the longest raypaths in the crustal root, appear as the most sensitive to the selected crustal root azimuth. For these particular events predicted travel-time delays show variations of up to 0.3 s between crustal root azimuths of N45°E to N55°E (events 17, 21 and 22, 32 of Fig. 5.9d-e).

Figure 5.9: Next page, *left*: travel-time delays and corresponding standard deviations of a few events recorded on the COOK and WCOAST stations (*circles and error bars*) are presented along with calculated time delays for SIGHT Transect 2 crustal model extrapolated into 3D (Fig. 5.5) and two possible crustal root azimuths of N45°E (*red curves*) and N55°E (*blue curves*). Time delays are shown relative to delays at station MCV and to the ak135 Earth model (Kennett *et al.*, 1995). They are presented in groups of similar back azimuth and incidence angles, and are ordered with back azimuths from west to east and small to large incidence angles.



#### Earthquakes from the northwest

Observed travel-time delays and predictions from the SIGHT Transect 2 crustal model (Kleffmann, 1999; Scherwath *et al.*, 2003) show good agreement for distances up to 20 km (Fig. 5.8a) or 0 km (Fig. 5.8b–c) east of Mount Cook Village (MCV) and distances greater than  $\sim$ 80 km east of MCV (Fig. 5.8c). For stations west of kilometer –25 and continuing westerly (Fig. 5.8c), positive and increasingly large residuals are produced by surface gravels and Tertiary sediments (Kleffmann, 1999), while near 0 km, peak time delays result from of the Southern Alps crustal root low crustal wave speeds relative to the faster wave speeds of the surrounding mantle (Fig. 5.8a–c).

However, east of  $\sim 20$  km (Banda earthquake, Fig. 5.8a) or 0 km (Honshu earthquake, Fig. 5.8c) observed arrivals are consistently earlier than predicted. This discrepancy is about -0.3 s for the Banda Sea earthquake with rays arriving from a  $-60^{\circ}$  azimuth (Fig. 5.8a) but as large as  $\sim -0.7$  s for the more distant Honshu earthquake with rays arriving from a  $-25^{\circ}$  azimuth (Fig. 5.8c). The discrepancy is possibly even greater with ca. -0.9 s for the Irian Jaya earthquake that is inline with the profile and has rays arriving from  $-45^{\circ}$  from north (Fig. 5.8b). The discrepancy could be -0.6 s, i.e. less than indicated above, because of the uncertainty of the alignment between the calculated travel-time delays and the picks (distances i 0 km of Figure 5.8b) Time delays of the Irian Jaya and Banda events appear inconsistent with each others as they show large variations for a back azimuth difference of  $15^{\circ}$  only.

Similar travel-time anomalies are seen in the COOK/WCOAST events of Figure 5.9. Note that the WCOAST/COOK array, which covers the western half of Transect 2 only, samples only the left flank of the postulated travel-time anomaly of Figure 5.8. The travel-time advance relative to delay times calculated for the crustal root alone is ca. 0.3 s for P/pP phases from the Philippines and Taïwan that are almost in line with the profile (events 5, 9, 18 and 19 of Fig. 5.9b). In contrast, the travel-time advance is large and ca. 1.8 s for pP/sP phases from the north and the close Vanuatu and Loyalty Islands (events 17 and 21 of Fig. 5.9d). Finally, the travel-time advance is moderate for earthquakes from the Fiji and Tonga Islands located northeast of New Zealand and aligned with the plate boundary (events 22 and 32 of Fig. 5.9e).

The time advance variation with back azimuth is similar to that observed from the Banda Sea and Honshu earthquakes (Fig. 5.8). The travel-time advance is largest for earthquakes located to the north (events 17 and 21), whose rays graze the plate boundary but the advance is small for rays propagating perpendicular to (Banda Sea, Chile, events 5, 9, 18 and 19) or along the plate boundary (events 22 and 32).

#### Earthquakes from the southeast

Travel-time delays from an earthquake located on the north coast of Chile with a back azimuth of N135°E (Fig. C.4, Tab. C.1) inline with the profile display a small mismatch west of the main divide, of only -0.2 s to -0.4 s, with predicted residuals from the crustal root model only (distances < -10 km of Fig. 5.8b).

Travel-time delays from the Chilean coast event (back azimuths of N135°E) that were recorded by the WCOAST array presents a contrasting difference of 0.2 s to 0.4 swith time delays predicted for a crustal root alone. These delays disagree with time delays measured from the SIGHT Chile event (compare Fig. 5.8g and Fig. 5.9f). Reasons for the discrepancy could be the low quality of this particular WCOAST measurement (quality C) or altered raypaths predicted by the crustal model only. The explanation for discrepancies greater than 0.2 s between observed travel-time delays and predictions for the SIGHT Transect 2 crustal structure (Scherwath *et al.*, 2003) is assumed to lie within the mantle. The overall travel-time advance and the varying amplitude of this advance with back azimuth appears consistent with a high wave-speed body centred beneath the Southern Alps (Stern *et al.*, 2000) that is elongated in a southwest-northeast direction approximately parallel to the plate boundary.



z = 110 km

Figure 5.10: This map view illustrates the high-speed mantle body beneath central South Island (see Fig. 5.6 for a cross-section). *Blue, green and red raypaths* are shown for events 1 and 28 from the northwest, events 17 and 21 from the north and events 22 and 32 from the northeast, respectively.

The model of Stern *et al.* provided a good fit to arrivals along a narrow range of azimuths from the northwest. The intent of the present study is to provide a wider range of azimuths for teleseismic waves in an attempt to further define both geometry and velocity structure.

Rays travelling along the plate boundary (red raypaths in Fig. 5.10) will propagate a long way through the mantle body and also have similar path lengths through it. The travel-time advance will therefore be uniform across the WCOAST/ COOK array as all rays will have experienced similar time advances. In contrast, for earthquakes from the north with back azimuths slightly off from the mantle body orientation (green raypaths), path lengths through the mantle body will vary the most. As a result, the travel-time advance will appear as the largest. Finally, rays propagating perpendicular to the mantle anomaly (blue raypaths) will also have variable path lengths through the mantle body. These will, however, be in general shorter than those of rays arriving from the north and the time advance will therefore be moderate.

### 5.6 Modelling of the mantle wave-speed anomaly

Stern *et al.*'s (2000) vertical mantle body provides a good fit to most arrivals arriving sub-parallel to the profile line, i.e. perpendicular to the plate boundary (model VERT as a red curve in Fig. 5.11a–d, Fig. 5.12a–f). The fit however degrades for arrivals from the north that have shallow incidence angles (Fig. 5.12d).

The mantle wave-speed anomaly is modelled with a series of horizontal cylindrical shapes (Tab. 5.2) that aim at investigating the mantle body's parameters: dip, width, wave-speed anomaly and azimuth. Stern *et al.*'s (2000) mantle body is taken as initial model. The centre of the mantle body is fixed. The position of the centre of mass along the profile line is constrained to where the mantle residual gravity anomaly, i.e. the difference between the Bouguer anomaly and the gravity effect of the crustal root, is maximum (Stern *et al.*, 2000). The depth of the centre of mass is set to ~110 km, estimated by Stern *et al.* (2000) from the half-maximum of the mantle residual anomaly. In some models (e.g. models W130, W160 of Tab 5.2), however, a larger width will increase the mass of the mantle body and therefore the wavelength and amplitude of the gravity effect of such a body. This is compensated by thinning the body from the bottom, which results in shallowing the body centre. In contrast, to the centre of the mantle body, the other parameters height H, width W, dip, wave-speed anomaly  $\delta Vp$ , and azimuth from north azN, are varied independently (see Fig. 5.6 for a representation of the parameters).

Table 5.2: Models for the mantle high-speed body. Models are defined by a centre (with a lateral position indicated by a longitude for a fixed latitude of  $-43.73^{\circ}$  or a distance X from MCV along the profile and a depth z), a height H, a width W, an inclination from the vertical aV (i.e. 90° from the dip), an azimuth from north azN, a wave-speed anomaly  $\delta Vp$ , and a factor F between 0 and 1, that represents the fraction of the cylinder radius, to which Vp is constant. See Figure 5.6 for a representation of these parameters.

model	Longitude (°)	$z~(\mathrm{km})$	H (km)	W (km)	$aV~(^{\circ})$	Dip (°)	azN (°)	$\delta V p \ (\rm km/s)$	F
VERT (or $W100$ ,	170.0	110	125	100	-	-	45	0.6	0.2
Az45, $\delta$ Vp0.6)									
W130	170.0	100	110	130	-	-	45	0.6	0.2
W160	170.0	90	90	160	-	-	45	0.6	0.2
SUBE	170.0	110	110	100	-40	$50^{\circ}$ SE	45	0.6	0.2
SUBW	170.0	110	110	100	40	$50^{\circ}$ NW	45	0.6	0.2
Az30	170.0	110	125	100	-	-	30	0.6	0.2
Az60	170.0	110	125	100	-	-	60	0.6	0.2
$\delta V p 0.5$	170.0	110	125	100	-	-	45	0.5	0.2
$\delta V p 0.7$	170.0	110	125	100	-	-	45	0.7	0.2
Az30W100 $\delta$ Vp1	170.0	110	125	100	-	-	30	1.0	0.2
${\rm Az60W160} \delta {\rm Vp1.2}$	170.0	90	90	160	-	-	60	1.2	0.2
$\delta V p 0.2$	170.0	110	125	100	-	-	45	0.2	0.2
$\delta V p 0.3$	170.0	110	125	100	-	-	55	0.3	0.2

Table 5.2: continued											
model	Distance $(km)$	Depth (km)	$H (\rm km)$	W (km)	$aV~(^{\circ})$	Dip (°)	$azN~(^{\circ})$	$\delta V p \ (\rm km/s)$	F		
Stern et al. $(2000)$	12 km east	110	125	100	-	-	45	0.6	$\sim 0.2$		
$Az45\delta Vp1.1$	$12 \mathrm{km} \mathrm{east}$	110	125	100	-	-	45	1.1	$\sim 0.2$		

Dip

The vertical body (model VERT of Tab. 5.2) offers a better fit to most events from the northwest (Honshu, Irian Jaya events of Fig. 5.11b–c and events 1 and 28 of Fig. 5.12c) than the dipping bodies do (models SUBW, SUBE of Tab. 5.2). In contrast, events that show smaller travel-time advances (Banda event and events 5, 9, 18 and 19 of Figures 5.11a, 5.12b) yield a better fit with the east-dipping body (SUBE) because this has the effect of smoothing the travel-time anomaly for rays arriving from the northwest. Events 31 and 33, which have steep rays, are insensitive to the mantle body dip (Fig. 5.12a) because the time advance across the three models is similar for these nearly vertical raypaths. Finally, none of the three models offer a close fit to events 17 and 21 (Fig. 5.12d).

#### Width

The Honshu, Irian Jaya and most of the WCOAST/COOK events except for events 5, 9, 18, 19 and 17, 21 obtain a good fit with a 100–130 km wide body with vertical dip (models VERT, W130 of Tab. 5.2, Fig. 5.13–5.14). Time delays of the Banda earthquake and events 5, 9, 18 and 19, in contrast, are better matched by a wider (160 km) body (Fig. 5.14b). This is because the wider but thinner body decreases the lateral wave-speed gradient within the body, which is a similar effect to lowering the wave-speed anomaly  $\delta V p$ . Similarly, events 17 and 21 time delays (WCOAST side) are better matched by a 130–160 km wide body, i.e. a wave-speed anomaly that extends slightly further to the northwest. The discrepancy with observed travel-time delays of events 17 and 21, however, remains large with ~0.7 s for the WCOAST side.



Figure 5.11: (above) SIGHT events and predicted travel-time delays for models SUBW, VERT (*red curve*) and SUBE (Tab. 5.2) that have mantle bodies dipping  $50^{\circ}$  NW, vertical and dipping  $50^{\circ}$  SE, respectively. Note that time delays are shown relative to the mean value.

Figure 5.12: (Right) WCOAST/COOK events and predicted travel-time delays for models SUBW, VERT (*red curve*) and SUBE (Tab. 5.2) that have mantle bodies dipping 50° NW, vertical and dipping 50° SE, respectively. Time delays are shown relative to delays at station MCV. Note the major misfit for events 17 and 21 (d) with back azimuths of  $-2^{\circ}$  to  $5^{\circ}$ .





Figure 5.13: Same as Figure 5.11 for models VERT (W100; *red curve*), W130 and W160 (Tab. 5.2), which mantle bodies are 100 km, 130 km and 160 km wide, respectively.



Figure 5.14: Same as Figure 5.13 for WCOAST/COOK events.



Figure 5.15: Same as Figure 5.11 for models  $\delta Vp0.5$ , VERT ( $\delta Vp0.6$ ; red curve) and  $\delta Vp0.7$  (Tab. 5.2) with maximum wave-speed anomalies of 0.5 km/s, 0.6 km/s and 0.7 km/s, respectively.



Figure 5.16: Same as Figure 5.15 for WCOAST/COOK events.



Figure 5.17: WCOAST/COOK events and predicted travel-time delays for models az30, VERT (az45; *red curve*) and az60 (Tab. 5.2)



Figure 5.18: Rays of the COOK events have been projected onto cross-sections of model VERT (Tab. 5.2) oriented at N135°E along SIGHT T2 (top) and N50°E (middle). SIGHT events have been projected onto a cross-section oriented N135°E (bottom). Red triangles represent recording stations. Cross-sections are not to scale.

#### Amplitude of the wave-speed anomaly

A wave-speed anomaly,  $\delta Vp$ , of 0.5–0.6 km/s offers a reasonable match to most traveltime residuals, which is similar to the wave-speed anomaly deduced by Stern *et al.* (2000). However, a smaller  $\delta Vp$  is required for the Banda event, the Chile event and events 5, 9, 18 and 19 (Fig. 5.15a,d and Fig. 5.16b) and a larger one for events 17 and 21 (Fig. 5.16d). Figure 5.16d also suggests that raypaths from events 17 and 21 to the stations located northwest of station MCV are unaffected by the wave-speed anomaly. The smaller and larger wave-speed anomalies are not compatible with models by either Stern *et al.* (2000) or Kohler and Eberhart-Phillips (2002).

#### Azimuth

Rays arriving from the northwest or the southeast, i.e. nearly perpendicular to the mantle body azimuth, show little sensitivity to the mantle body's azimuth. These include all SIGHT events and some WCOAST/COOK events. In contrast, rays from the north are sensitive to the western extent of the mantle high-speed zone and, thus, its azimuth. For instance, events 17, 21 and 22, 32 show the strongest variations of travel-time residuals across the array. Travel-time delays of events 17 and 21 find a slightly better match with a body oriented N30°E than with a N60°E one (Fig. 5.17d) or, with different model parameters. Similarly, events 22 and 32 suggest an azimuth of  $30^{\circ}-45^{\circ}$  (Fig. 5.17e). As seen in the top of Figure 5.18, the raypaths of neither event groups sample the entire width of the mantle body but only the western half, which is the reason for the trade-off between azimuth and the width of the wave-speed anomaly.

#### Events 17 and 21 from the north

None of the above presented models offers a close match to travel-time residuals from events 17 and 21, with all predicting an insufficient travel-time advance across the array.



Figure 5.19: Top: Travel-time delays of events 17 and 21, and the fit provided by a set of models that illustrate the trade-off between width, azimuth and wave-speed anomaly. Bottom: events 17 and 21 rays superposed on a north-oriented slice through model Az45 $\delta$ Vp1.1. The light-purple mantle region represents the entirely unconstrained domain.

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Rays from events 17 and 21 arrive from the north at a shallow incidence of  $41^{\circ}-52^{\circ}$ . These rays graze the plate boundary and are, thus, most sensitive to the top northwestern extent of the mantle body and the lateral wave-speed gradient there in particular. Figures 5.14d and 5.17d suggest that bodies that are wider and/or have a more northward orientation (N30°E) improve the fit to the WCOAST side (west of MCV) of the residuals but degrades it on the COOK side (east of MCV). The fit to the COOK side, in contrast, is improved by a stronger wave-speed anomaly or a west-subducting body, which contribute to increasing the variation in the raypath integrated time advance (Fig. 5.12d, 5.16d).

Figure 5.19 illustrates the fit provided by (a) a 100 km wide wave-speed anomaly of 1 km/s with a more northward azimuth of N30°E than the N55°E plate boundary orientation (model Az30W130 $\delta$ Vp1.0 of Tab. 5.2), (b) an up to 160 km wide body with a N60°E azimuth and a 1.2 km/s wave-speed anomaly (model Az60W160 $\delta$ Vp1.2). An alternative model (Az45 $\delta$ Vp1.1 of Tab. 5.2; bottom of Fig. 5.19) is presented that is about 100 km wide, is oriented at N45°E, has a  $\delta$ Vp of 1.1 km/s and a northwestern corner that is wider than in the original model of Stern *et al.* (2000). Note that this is a possible model for the zone sampled by these rays only, i.e. the top northwestern corner of the mantle wave-speed anomaly. Although these models are poorly constrained, they all require a minimum wave-speed anomaly  $\delta$ Vp of 1.0 km/s and a high-speed zone extending slightly further to the top or northwest than the initial mantle model VERT.

#### Models for the Banda event and events 5, 9, 18 and 19

In contrast to the high wave-speed anomalies required for events 17 and 21 with rays arriving from the north, a lower wave-speed anomaly of  $\sim 0.3$  km/s only is required to

match time residuals of the Banda earthquake and events 5, 9, 18 and 19 with rays arriving from the northwestwest and also time residuals of the offshore Chile event with rays arriving from the southeast (Fig. 5.20).



Figure 5.20: Travel-time delays of the Banda event (top left), events 5, 9, 18 and 19 (top right), Chile event (lower left) and models  $\delta$ Vp0.3 and  $\delta$ Vp0.2 (Tab. 5.2) that have mantle bodies with azimuth and wave-speed anomalies of N55°E, 0.3 km/s and N45°E, 0.2 km/s, respectively.

#### 5.6.1 Models' misfits

Model residual distributions are presented in box-and-whisker plots (Fig. 5.21). For the SIGHT events (Fig. 5.21a), the boxplots indicate skewed distributions for models SUBW and  $\delta$ Vp0.7 of Table 5.2 and the largest interquartile range and number of outliers for models SUBE and W160, respectively.

Other models (VERT, W130,  $\delta$ Vp0.5 of Tab. 5.2), in contrast, present relatively symmetric distributions, smaller interquartile ranges and reasonable numbers of outliers. The mantle body is, therefore, best represented by a subvertical and 100–130 km wide body with an average  $\delta$ Vp of 0.5–0.6 km/s. Models Az30 and Az60 distributions are not taken into account because the SIGHT events are not sensitive to the mantle body azimuth. Similar plots for the WCOAST/COOK events (Fig. 5.21b) do not show much distinction between models because of the small number of residuals per event and the limited horizontal range. The northwest-dipping body (model SUBW) only stands out as least viable with the largest interquartile range.

Model misfits are presented as the weighted Root Mean Square (L2-norm) (Fig. 5.22). For a single event the L2-norm is defined by

$$L2norm_j = \frac{1}{N_j} \sqrt{\sum_{i=1,N_j} \frac{((x_i - \bar{x}) - (y_i - \bar{y}))^2}{\sigma_i^2}}$$
(5.1)

where  $N_j$  is the number of travel-time delays for the  $j^{th}$  event,  $x_i$  represent single travel-time delays with average  $\bar{x}$ ,  $\sigma_i$  are error bars,  $y_i$  represent calculated travel-time delays and their average  $\bar{y}$ . Values for the whole event ensemble are given by

$$L2norm_{tot} = \sqrt{\sum_{j=1,M} \sum_{i=1,N_j} \frac{((x_{ij} - \bar{x_j}) - (y_{ij} - \bar{y_j}))^2}{\sigma_i^2}} / \sum_{j=1,M} N_j$$

with M, the number of events.



Figure 5.21: Box-and-whisker plots of model residuals with (a) SIGHT and (b) WCOAST/COOK travel-time delays. *Boxes* represent the interquartile range, which includes 50 % of the data. The *red middle bar* is the median model residual. *Whiskers* are minimum and maximum values, which include 100 % of the data in the absence of outliers (*crosses*). Outliers are those residuals that are greater than 1.5 times the quartile value.

Models generally have L2-norms that are twice as large for the WCOAST/COOK data set than for the SIGHT one. This is a consequence of the smaller error bars ( $\sigma_i$  in the denominator of equations 5.1 attributed by the adaptive stacking, which was used on the WCOAST/COOK data set, than the errors attributed by picking on the SIGHT data set. The total L2-norms show little variation between models, suggesting that the data are not well suited to discriminate between the various mantle body geometries. They show a larger misfit for events from the southeast.





Figure 5.22: L2-norm of model residuals for single SIGHT (left) and WCOAST/COOK (right) events. The L2-norm for the whole set of events is given for each model on the right of the legend. *Red lines* indicate a model best fits for a L2-norm value of one.

norms indicate a greater misfit for models SUBW, SUBE,  $\delta$ Vp0.7 and W160 than for models VERT, W130 and  $\delta$ Vp0.5. For instance, models SUBW and SUBE (northwest-/southeast-dipping bodies) L2-norms for single events show the most variability with back azimuth from all the models. Their total L2-norms class them as the worst models for the SIGHT and WCOAST/COOK data sets, respectively.

In contrast, model  $\delta$ Vp0.5 displays the smallest L2-norm for the SIGHT data set and a smaller misfit than the initial VERT model with a  $\delta$ Vp of 0.6 km/s for the WCOAST/COOK data set. L2-norms of models Az30 and Az60 for the SIGHT events show little or no variation from those of model VERT, reflecting the lack of sensitivity of those events to the mantle body azimuth.

Models Az30 and Az60 misfit values for the SIGHT events have little significance as these events are inline with the profile and insensitive to variations in the mantle body's azimuth. For the WCOAST set of events, the L2-norm suggests that model Az30 azimuth is a better azimuth for the mantle body than the 45° from north of model VERT is. This, however, is the contribution of the event 21 large misfit to the L2-norm, whose rays sample the western portion of the mantle body only, and suggests that model VERT underestimates the extent of the mantle body to the northwest.

#### 5.6.2 Contribution of anisotropy to the wave-speed anomaly

A Moho shallowing at a 10° angle along-strike and to the northeast can account for a ~0.1 s error on travel-time delays for northern back azimuths (see p. 156), but not a ca. 1.2 s delay, the difference between measured travel-time residuals of events 17, 21 and corresponding synthetic residuals of tested models (VERT to  $\delta$ Vp0.7 of Tab. 5.2 with time delays shown in (d) of Fig. 5.12–5.17). Figure 5.19 suggests that the mantle

wave-speed anomaly is ca. 1.1 km/s for this particular back azimuth and an incidence angle of  $\sim 50^{\circ}$ .

Both large SKS splitting (Klosko *et al.*, 1999) and Pn anisotropy (Scherwath *et al.*, 2002) have been reported beneath, west and east of the Southern Alps. SKS fast polarisation orientations (and delay times) are  $26^{\circ} \pm 7.5^{\circ}$  (1.6  $\pm$  0.24 s) and  $29^{\circ} \pm 9^{\circ}$ (1.7  $\pm$  0.27 s) at stations GLAA (West Coast) and MTCA (Mount Cook Village) (Klosko *et al.*, 1999) both located near SIGHT Transect 2. An apparent Pn anisotropy of 11.5  $\pm$  2.5 % (Scherwath *et al.*, 2002) was estimated from Pn-wave speed measurements along the mutually perpendicular NW-SE Transect 2 and NE-SW line 3W of SIGHT. Apparent and maximum Pn anisotropy of 6.5  $\pm$  3.5 % and 7–13 %, respectively, were estimated at the crossing SIGHT Transect 2 and Fiordland-Cheviot profile (Section 2.5).

Assuming simple shear and/or pure shear (Sanderson and Marchini, 1984) in the mantle with horizontal orientation of the shear axis, then shallow incident rays arriving from north are ca. 30° from the fast propagation azimuth and 40° from the horizontal plane. In contrast, rays arriving from a  $-60^{\circ}$  from north at an incidence angle of  $23^{\circ}-31^{\circ}$  sample the slow propagation orientation, while vertically incident rays may sample an intermediate wave speed (assuming minimal vertical shear). The apparent wave-speed anomaly is ca. 0.3 km/s and 1.1 km/s for back azimuths of ca.  $-60^{\circ}$  and  $0^{\circ}$  from north respectively, i.e. wave speeds of 8.4 km/s and 9.1 km/s assuming an isotropic mantle with background wave speed of 8.1 km/s. These two values suggest an apparent anisotropy of  $8 \pm 3$ %. This anisotropy estimate is a minimum because propagation paths for these two event groups are neither mutually perpendicular nor contained in the horizontal plane, where wave-speed variations should be the largest if the anisotropy has orthorombic symmetry. The true amount of anisotropy in the centre

of the mantle body may be even larger if the surrounding mantle has a background anisotropy both west and east of the mantle body as suggested by SKS splitting and Pn anisotropy studies (see Section 2.5.2). Such high anisotropy could reflect mantle shear increasing laterally towards the central vertical axis of the mantle body.

## 5.7 Conclusions

P travel-time delays were measured from a set of 35 events recorded along two consecutive deployments COOK and WCOAST that combine into one profile. Travel-time delays were also measured on four teleseisms recorded along Transect 2 during the 1996 SIGHT project. Three-dimensional forward modelling is used to calculate synthetic time delays for the large range of back azimuths recorded and to test possible geometries for a mantle high-speed anomaly beneath the Southern Alps.

Travel-time delays display large variations with back azimuth and incidence angle. These show time advances up to 1.8 s for events from the north relative to predicted travel-time delays from the SIGHT Transect 2 crustal structure (Scherwath *et al.*, 2003) but time advances of only 0.3 s for back azimuths of  $-50^{\circ}$  to  $-60^{\circ}$  from north. Travel-time advances are consistent with the presence of a mantle high-speed body beneath the Southern Alps (Stern *et al.*, 2000). Variations with most but not all back azimuths are explained by a horizontal cylindrical mantle body that stretches along the plate boundary. A sub-vertical, 100–130 km wide mantle body with a 0.5–0.6 km/s wave-speed anomaly similar to Stern *et al.* (2000) model offers a best match to northwestern back azimuths. The available data are insufficient to constrain the mantle body azimuth, which is assumed to lie in the range  $30^{\circ}$ – $60^{\circ}$  from north.

This model, however, underestimates and overestimates mantle wave-speed anomalies for back azimuths of about 0° and  $-60^{\circ}$ , respectively. Wave-speed anomalies of  $\sim 1.1$  km/s and  $\sim 0.3$  km/s are respectively required for the back azimuths of  $\sim 0^{\circ}$  and  $-60^{\circ}$  from north with epicentral distances of  $21^{\circ}-28^{\circ}$  and  $54^{\circ}-81^{\circ}$  (i.e. incidence angles of  $\sim 50^{\circ}$  and  $\sim 30^{\circ}$ ). The difference in the modelled wave-speed anomalies for both azimuths is larger than can be attributed to neglecting thinning of the crustal root to the northeast and/or a  $10^{\circ}$  error in the chosen crustal root azimuth. Both crustal root thinning and uncertainty of the azimuth may account for a 0.1 s and a 0.2 s difference, respectively, i.e. a total of 0.3 s, which is less than the 1 s difference between back azimuths.

Anisotropy is proposed as a possible explanation for the discrepancy with the fast wave-speed axis aligned with the shear direction and the low wave-speed axis being in the horizontal plane and perpendicular to the fast axis. The apparent anisotropy estimated from the modelled wave-speed anomalies from northern and northwestern back azimuths is a minimum of  $8 \pm 3 \%$  (Section 5.6.2). Note that this interpretation relies on data that present almost nonexistent ray coverage from the east to the southwest (Fig. 5.18, p. 182).

Future suggested work: More travel-time delays need to be measured on phases arriving from the north and northeast in order to confirm the observed travel-time delays dependence on the back azimuth and incidence angle, and to constrain the width and azimuth of the wave-speed anomaly. This could be done by repeating this study's experiment with a profile extending further southeast than the combined WCOAST/COOK array and east of lake Tekapo. Furthermore, a profile extending up to 100 km off-shore and for a minimum of six months is necessary to record earthquakes from the southern Chilean Coast and to constrain the dip of the wave-speed anomaly. A  $\leq 15$  km station spacing would allow sampling the gradient of the travel-time anomaly with respect to distance. Finally, including Pn phases from the northeast and southwest will help to constrain the wave-speed along the fast wave-speed axis and its orientation. Further analysis on this study data set would benefit from exploring the trade-offs between model parameters, especially those between width and azimuth or wave-speed anomaly and gradient of the mantle body

The dependence of the wave-speed anomaly on the back azimuth and the incidence angle suggests that inversions incorporating anisotropy will help to recover the full wave-speed anomaly in the mantle lid.

## Chapter 6

# Discussion and conclusions

## 6.1 Overview

Four independent analyses have been used to characterise variations in crustal thickness and mantle lithosphere properties in a region of oblique continent-continent collision. The analyses used teleseismic P-wave delays, earthquake refraction, modelling of gravity anomalies and teleseismic receiver functions.

#### Earthquake refraction

Pn speed is  $8.54 \pm 0.20$  km/s along the Southern Alps foothills on a profile ca. 5° clockwise from the plate boundary orientation (Section 2.3). The high Pn speed is shown to result from 7–13 % seismic anisotropy with a fast orientation sub-parallel to the plate boundary, as well as a relatively high isotropic wave speed of  $8.3 \pm 0.3$  km/s in the mantle lid (Section 2.5). A maximum crustal thickness of  $48 \pm 4$  km is estimated along the refraction profile near Wanaka and beneath the southern portion of the Southern Alps, which represents a ca. 18 km thick crustal root relative to a 30 km crustal thickness at the coasts (Section 2.4). This crustal root is thicker than predicted by Airy isostasy to sustain the topographic load of the 1000 m mean elevation. The

gravity effect of the over-thickened crust is apparent as a negative isostatic anomaly of -20 mGal at Wanaka (Section 2.6). For an assumed  $-300 \text{ kg/m}^3$  across-Moho density contrast (Section 2.6.2), a positive mass anomaly is deduced in the mantle lid beneath the region of over-thickened crust that has a minimum  $35 \pm 5 \text{ kg/m}^3$  density anomaly with the surrounding mantle material (Section 2.6.3). The mass anomaly has a minimum width of  $110 \pm 20 \text{ km}$  and a minimum thickness of  $70 \pm 20 \text{ km}$ . Assuming thermal contraction of mantle rocks, the density anomaly can be interpreted as a 200–400 °C negative temperature anomaly with the surrounding mantle that results from the downwarp of isotherms (Section 2.7, p. 45). Such a mass anomaly is interpreted to represent a dynamic pull on the overlying crust, and to cause enhanced crustal thickening. The Australian oceanic slab of the nearby Puysegur subduction zone located southwest of the Southern Alps region is proposed to act as a rigid backstop that contributes to the thickening of Pacific mantle lithosphere at the southwestern corner of the Southern Alps (p. 46).

#### **Receiver functions**

Forward modelling of P, S and SKS receiver functions at Geonet stations JCZ, WKZ and EAZ are used to produce a profile of crustal thicknesses across Otago and the southern portion of the Southern Alps. A model was also derived at station RPZ that provides a comparison with the crustal structure from SIGHT Transect 1 (Van Avendonk *et al.*, 2004).

Effects of dipping boundaries and anisotropy on S receiver functions have been explored with synthetics from simple one-layer models. The synthetics show that for SKS receiver functions, a steeply-dipping layer with strong impedance contrast at the top, causes a negatively polarised direct SKS-wave pulse at 0 s delay time of the receiver
functions for the updip-travelling SKS wave (Section 3.3.1, p. 65). The synthetics also show that for the shallower incident and updip-travelling S wave a steeply-dipping layer inhibits Sp conversions, when incidence angles are greater than the P-wave transmission critical angle at the dipping interface (Section 3.3.1, p. 61). Furthermore, an anisotropic layer with a strong impedance contrast at the top causes a negative direct S-wave pulse at epicentral distances less than  $\sim 45^{\circ}$  and back azimuths within  $45^{\circ}$  from the symmetry axis (Section 3.3.3). The negative pulse is interpreted to result from the elliptical particle motion that arises from the interference between the incident S wave and its free surface Sp conversion at incidence greater than the S-wave critical angle (Booth and Crampin, 1985). The back azimuth dependence of the initial pulse polarity suggests that the modified incidence angle is a function of the angle between the ray and the anisotropy symmetry axis, and results from the anomalous refraction angle that occurs at the interface between anisotropic and isotropic media (e.g. Slawinski *et al.*, 2000).

A Vp/Vs ratio of 1.60 is estimated from station RPZ P receiver functions for the upper and mid crust (Section 4.6.1, p. 98). This low value corroborates previous low Vp/Vs ratio estimates of 1.56 (Pulford, 2002) for the mid crust and 1.65 as a crustal average (Kleffmann, 1999) for the central Southern Alps. Stern *et al.* (2001) interpreted the low Vp/Vs ratio to result from high pore fluid pressure released during prograde metamorphism.

Combined P and S receiver functions image deepening of a discontinuity across western Otago from  $34 \pm 5$  km (JCZ) and  $32 \pm 3$  km (EAZ) on either side of the Southern Alps to  $39 \pm 4$  km depth beneath the highest topography (ca. 50 km NW of WKZ) assuming Vp/Vs ratios of 1.65–1.75 (Section 4.7). The interpretation of this discontinuity as the Moho is inconsistent with a -80 mGal Bouguer gravity anomaly and previous crustal thickness estimates of 43–45 km from 3D inversion of travel-time data (Eberhart-Phillips and Bannister, 2002) and 48  $\pm$  4 km from earthquake refraction. Low Vp/Vs ratios of 1.55–1.60 for the crust in western Otago could reconcile P and S receiver function delay times with the gravity and these previous crustal estimates. Alternatively, the discontinuity may be interpreted as a lower crustal boundary implying a relatively low wave-speed contrast at the Moho.

Receiver functions suggest the presence of a 50–70 km deep upper mantle discontinuity 15 km to 30 km below the Moho. S-wave receiver functions suggest that the discontinuity could represent a rotation of the anisotropy fast symmetry axis (Section 4.7, p. 140). P-wave receiver functions suggest an additional Vp increase of  $\sim 0.3$  km/s within the mantle, possibly associated with the same discontinuity. The rotation of the fast axis may be interpreted as mechanical decoupling, while partial eclogitisation or the Hales discontinuity represent viable explanations for the Vp increase.

#### Teleseismic P travel-time delays

Travel-time delays were measured across the central Southern Alps along a ~80 km long profile extending from the West Coast to east of Lake Pukaki. Depending on the ray azimuths and incidence angles, travel times show arrivals 0.3 s to 1.8 s earlier than those predicted for the presence of the Southern Alps crustal root alone, assuming a 2D model for the crust extrapolated into the third dimension at an azimuth of N50°E (Section 5.5). These early arrivals are consistent with the presence of a previously hypothesised mantle body (Stern *et al.*, 2000), which is a high wave-speed anomaly, relative to a regular mantle wave speed of 8.1 km/s. The smallest advances are observed for event back azimuths of  $-60^{\circ}$  to  $-50^{\circ}$ . The largest advances are observed for events located close to (epicentral distances of  $21^{\circ}-27^{\circ}$ ) and north from New Zealand, whose

arrivals have incidence angles of  $40^{\circ}-52^{\circ}$ .

Three-dimensional forward modelling suggests that the mantle body is sub-vertical, 100–130 km wide and is centered at ca. 110 km depth (Section 5.6). The average wave-speed anomaly required is 0.5–0.6 km/s for most events, the majority of which are located northwest of New Zealand. For the smallest and largest time advances of 0.3 s and 1.8 s, however, weaker (~0.3 km/s) and stronger (~1.1 km/s) wave-speed anomalies, respectively, are required (Section 5.6, p. 184–185). The difference is attributed to seismic anisotropy in the mantle lid (Section 5.6.2) and represents a ~8 % apparent minimum anisotropy in the mantle, consistent with the anisotropy estimate from earthquake refraction. It is proposed that the thickened mantle lid is being deformed in both horizontal and vertical planes.

### 6.2 Patterns of lithospheric deformation

Variations in lithospheric deformation within the South Island and south of the Hikurangi subduction zone are divided into three zones: (1) The central South Island, where convergence is greatest; (2) A zone between SIGHT Transect 2 and Otago, where deformation is possibly affected by lateral extrusion of the lower crust and mantle (Gerbault *et al.*, 2002; Scherwath *et al.*, 2006) as a result of the obliquity of the relative plate motion vector with the plate boundary; (3) Otago and further south, where the Southern Alps oblique collision and the Puysegur oblique subduction both contribute to thicken the mantle lithosphere.

### Central South Island

The region between SIGHT Transect 1 and Transect 2 is characterised by maximum convergence and contraction (Beavan *et al.*, 2007; Henderson, 2003). Crustal thickening is important and expressed in maximum average elevations of 1,500 m (Mt Cook region) and the presence of a crustal root with maximum thickness increasing to the southwest from  $\sim$ 7 km at SIGHT Transect 1 (Van Avendonk *et al.*, 2004) to  $\sim$ 14 km at Transect 2 (Scherwath *et al.*, 2003, relative to a 30 km crustal thickness at the coasts). Mantle shortening is apparent in a high wave-speed (Chapter 5) and a high density anomaly beneath the Southern Alps that can be explained by the downwarp of isotherms alone (Stern *et al.*, 2000). The density anomaly is a positive mass anomaly that is a pull on the overlying crust and is the reason for the enhanced crustal thickness and a -20 mGal isostatic anomaly (Reilly *et al.*, 1977).

Teleseismic P travel-time delays indicate that: (1) anisotropy is pervasive throughout the zone of mantle shortening (Section 5.6.2); and (2) mantle shortening is 100– 130 km wide (Section 5.6) and more confined than the minimum 160 km wide zone of mantle simple shear inferred from seismic anisotropy measurements (Fig. 6.1; Section 2.5; Baldock and Stern, 2009, in prep.; Duclos *et al.*, 2005; Klosko *et al.*, 1999; Scherwath *et al.*, 2002).

Modelling of GPS velocities suggests that the width of present active mantle deformation could be 100 km wide only (Beavan *et al.*, 2007; Ellis *et al.*, 2006a). An Eocene passive margin was proposed to represent a lithospheric-scale heterogeneity that controlled the development of the Alpine Fault continental transform ca. 23 Myr ago (Sutherland *et al.*, 2000), and could represent a lithospheric heterogeneity that could contribute to localise deformation. Modelling (Houseman and Billen, 2002) shows that shear weakening can also contribute to localise deformation and shortening. Strain modelling (Savage *et al.*, 2007c) predicts that shortened mantle lithosphere stiffens as it thickens and, thus, inhibits further strain to build up in the lithospheric root. Hence, pervasive anisotropy in the shortened mantle suggests that strong anisotropy was acquired before the onset of mantle shortening. Thus, the presence of a high anisotropy in the lithospheric root is consistent with plate reconstructions that deter-



after (Baldock and Stern, 2004b)

Figure 6.1: Schematic representation of the widths of mantle deformation through shortening and simple shear as deduced from teleseismic P travel-time delays (*elliptical cross-section*) and seismic anisotropy (*light coloured region*), respectively. LAB denotes the lithosphere-asthenosphere boundary. Vectors perpendicular to the cross-section denote the relative plate motion. The vertical exageration is 1:1 (after Baldock and Stern, 2004b).

mine the onset of rapid convergence ca. 35 Myr after the creation of a transcurrent plate boundary (Cande and Stock, 2004; Walcott, 1998).

The difference between the 100–130 km wide zone of mantle shortening inferred from teleseismic travel-time delays (Section 5.6) and the  $\geq$ 160 km wide zone of mantle simple shear inferred from seismic anisotropy measurements poses the following unanswered questions: Is the  $\geq$ 160 km wide zone of mantle deformation inferred from seismic anisotropy actively deforming? Is active deformation, instead, confined to the 100–130 km wide deformation zone defined by teleseismic travel-time delays? Thus, has seismic anisotropy been frozen at the edges of the mantle deformation zone? Has deformation become more localised through time by a pre-existing mantle heterogeneity or strain weakening?

### Between SIGHT Transect 2 and Otago

About 40 km south of SIGHT Transect 2, and about 70 km south of the region of maximum elevations, greater crustal root thicknesses are suggested by a 20 mGal more negative Bouguer anomaly (Reilly and Whiteford, 1979) than along SIGHT Transect 2. Inversion of travel-time data (Eberhart-Phillips and Bannister, 2002) also indicates a  $\sim 3$  km thicker crustal root there than along SIGHT Transect 2. Finite-element modelling (Gerbault *et al.*, 2002) predicts that the offset between maximum crustal thicknesses and maximum elevations can be due to lithospheric buckling perpendicular to the direction of oblique convergence and lateral extrusion of lower crustal material perpendicular to the direction of gravity data (Scherwath *et al.*, 2006) suggests that the region of greater crustal thicknesses correlates with a greater positive mass anomaly in the mantle. Scherwath *et al.* (2006) argue that mantle extrusion by creep (Meissner *et al.*, 2002) could

be the reason for the greater mantle mass anomaly south of the maximum topography. The mantle mass anomaly appears to increase south into western Otago.

### Otago and further south

Under Wanaka (Fig. 2.1), approximately 100 km southwest of the gravity minimum,  $\sim 50$  km and  $\sim 100$  km northeast of Fiordland and the subducting Australian slab (Fig. 1.6), the crustal root is up to  $\sim 4$  km thicker than along SIGHT Transect 2 (Section 2.4) but the thickness of the high wave speed and dense upper mantle body is approximately similar (Section 2.6.3). As pointed out in Section 2.7, the similar or greater crustal root thickness beneath the southern portion of the Southern Alps, relative to that beneath the central portion, is counter-intuitive from observations of: mean elevations, which are  $\sim 500$  m less in the south: compression decreasing south (Plate 1 of Beavan *et al.*, 2007); and the total shortening being 30-40 km less (Cande and Stock, 2004; Walcott, 1998). Furthermore, GPS velocity vectors in Fiordland, Otago and Southland (relative to a fixed point on the east coast) indicate more northerly directions than in the regions located to the north (Henderson, 2003), suggesting a transition in the style of motion perhaps reflecting a change in the nature of the plate boundary. This rotation could possibly indicate the transition from the Southern Alps oblique collision to the Puysegur oblique subduction. Wanaka is directly located inboard of the transition's northern limit shown by GPS vectors. Thus, it is possible that the enhanced crustal and mantle thickening of the Wanaka region (Eberhart-Phillips and Bannister, 2002; Kohler and Eberhart-Phillips, 2002; Scherwath et al., 2006, and this study's Chapter 2) could result from the contribution to thickening of the nearby converging Fiordland block and Puysegur subduction zone. The Australian slab located southwest of the Southern Alps may represent a rigid backstop (Malservisi et al., 2003) that converges with the Pacific lithosphere at ca. 3/4 of the Australian-Pacific relative plate motion (Section 2.7, p. 46). While possibly as much as half of the convergence is taken through subduction at the Puysegur margin, the rest may be accommodated through tearing or twisting of the Australian slab (Reyners *et al.*, 2002; Smith and Davey, 1984) and thickening of the Pacific lithosphere.

### 6.3 Summary

The following points summarise this thesis' main findings:

- The Pn speed is 8.54  $\pm$  0.20 km/s at N60°E along the Southern Alps eastern foothills.
- The isotropic Pn speed is  $8.3 \pm 0.3$  km/s and the maximum Pn anisotropy is 7–13 % beneath the central Southern Alps (lake Tekapo).
- A maximum crustal thickness of  $48 \pm 4$  km is estimated beneath the southern extent of the Southern Alps near Wanaka (western Otago).
- A  $48 \pm 4$  km crustal thickness represents an  $18 \pm 4$  km thick crustal root relative to a 30 km crustal thickness at the coasts that is twice or three times greater than necessary to isostatically sustain a ~1000 m topographic load.
- A mass excess is proposed in the mantle below western Otago to compensate for the crustal root mass deficit. Assuming an across-Moho density contrast of -300 kg/m<sup>3</sup>, the mass excess is an minimum density contrast of 35 ± 5 kg/m<sup>3</sup>, 110 ± 20 km width and 70 ± 20 km thickness in the mantle.
- At central South Island station RPZ, teleseismic receiver functions corroborate a low Vp/Vs ratio of 1.60 for the upper and mid crust interpreted to result from high-pore fluid pressure released during prograde metamorphism.

- P and S teleseismic receiver functions at the western Otago stations EAZ and WKZ respectively suggest Moho depths of  $32 \pm 3$  km and  $39 \pm 4$  km that are inconsistent with the -80 mGal Bouguer gravity anomaly and crustal estimates from other seismic measurements of  $48 \pm 4$  km and 43-45 km near station WKZ.
- A mantle discontinuity is interpreted at 15–30 km below the interpreted Moho that could be associated with a minimum 0.3 km/s wave speed increase and/or a rotation of the anisotropy symmetry axis.
- Teleseismic P arrivals in the central Southern Alps display relative travel-time advances of 0.3 s to 1.8 s relative to those predicted from the crustal structure alone. Largest time advances are for shallow incident rays arriving sub-parallel to the plate boundary.
- The average time advance is consistent with a 0.5–0.6 km/s wave-speed anomaly with the surrounding mantle that is a body sub-parallel to the plate boundary, sub-vertical, 100–130 km wide and centred at about 110 km depth beneath the Southern Alps.
- Wave-speed anomalies of ~0.3 km/s and ~1.1 km/s are necessary to explain the smallest and largest travel-time advances of 0.3 s and 1.8 s, respectively. The difference is attributed to a minimum ~8 % seismic anisotropy with the horizontal fast axis along the shortened mantle body.

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### Appendix A

## Earthquake refraction

### Weighted-mean wave speed

In the case of individual speed measurements,  $v_i$ , with standard errors,  $\sigma_i$ , the mean wave speed may be calculated by averaging over the individual speed measurements weighted with the inverses of their respective squared standard deviations,  $\frac{1}{\sigma_i^2}$  (Bevington, 1969). The herein so called  $\frac{1}{\sigma^2}$ -weighted mean is defined as

$$\bar{v} = \frac{\sum_{i=1}^{n} \frac{v_i}{\sigma_i^2}}{\sum_{i=1}^{n} \frac{1}{\sigma_i^2}}.$$
(A.1)

and has corresponding common variance,  $\bar{\sigma}^2$ , and mean standard deviation,  $\bar{\sigma}$  are

$$\bar{\sigma}^2 = \frac{1}{(n-1)\frac{1}{\sum_{i=1}^n \frac{1}{\sigma_i^2}}} \sum_{i=1}^n \left(\frac{v_i - \bar{v}}{\sigma_i}\right)^2 \tag{A.2}$$

$$\bar{\sigma} = \sqrt{\bar{\sigma}^2}.\tag{A.3}$$

### Timing error correction

A three- to four-second timing error was identified in station HAW data. Crosscorrelation on three teleseisms that were recorded during the two-week deployment period is used to discriminate between a 3 s or a 4 s error (top left of Fig. A.1–A.3). The three teleseisms are located north and northeast of New Zealand and approximately in line with the profile. Their arrival times are, thus expected to increase linearly along the array. For each event, waveforms of stations MCV, OHA and WWS are cross-correlated with that of station HAW in order to estimate the corresponding arrival delay times relative to station HAW (top right of Fig. A.1–A.3). A regression line is fitted to stations MCV, OHA and WWS delay times to approximate a relative travel-time curve through the array. The timing error is estimated by interpolating the regression line at station HAW epicentral distance (bottom left Fig. A.1–A.3). The single timing error estimates and mean do not allow to discriminate between a 3 s or a 4 s (Tab. A.1). Station HAW picks were, therefore, not included in the wave-speed analysis. The cause to the large errors may be attributed to variations of the crustal thickness along the array of stations.

Event	ID	$\delta t$ predicted	σ
Loyalty Is.		3.55	4.31
Tonga		3.97	79.17
Vanuatu Is.		3.42	22.24
weighted mean		3.55	0.24

Table A.1: Timing errors,  $\delta t$ , represent the time difference predicted for station HAW. The time difference is estimated from the regression line on stations MCV, OHA and WWS time shifts that are estimated by cross-correlation (bottom left of Fig. A.1–A.3). Standard deviations,  $\sigma$ , are used to weight single timing errors when calculating the mean. Note that the Tonga and Vanuatu events contribute little to the weighted mean due to the large error bars on the estimated time shifts.





Figure A.1: *Top left*: single seismograms after instrument response correction and band-pass filtering with corner frequencies of 1 Hz and 3 Hz. *Top right*: station HAW cross-correlated seismograms for an 8 s correlation window. *Bottom left*: linear regression (*green line*) on cross-correlation delay times of stations MCV, OHA and WWS (*blue crosses*). The *red star* denotes the predicted time delay at station HAW.





Table A.2: Events 1–4 and 6 are the Fiordland aftershocks re-located by Martin Reyners (GNS Science) using a temporary seismograph deployment and Eberhart-Phillips & Reyners' 1D model for Fiordland (2001). Event 5 is the off-shore Cheviot event located by GeoNet using the standard 1D model for New Zealand Maunder (2001).

	Event	Origin	ı time		on	Magnitude	
nb	ID	Date	Time	Lat.	Long.	Depth (km)	$M_L$
1	2105255	2003/08/25	03:36:30.26	-45.111	166.964	20.5	5.0
2	2106280	2003/08/26	23:56:26.57	-45.486	166.596	19.7	5.5
3	2106314	2003/08/27	01:29:40.41	-45.442	166.716	22.7	5.1
4	2106319	2003/08/27	01:42:54.12	-45.314	166.945	24.4	5.6
5	2106361	2003/08/27	03:39:37.23	-42.802	173.758	35	4.1
6	2110611	2003/09/04	08:40:44.25	-45.224	166.921	22.7	6.1

Table A.3: Crustal (Pg) and mantle phase (Pn) wave speeds are estimated through chisquare fitting of a regression line (Press *et al.*, 1992) to the arrival picks. Regression coefficients are  $T_i$ , the time intercept, and  $v_P^{-1}$ , the slope, that have standard deviations,  $\Delta T_i$  and  $\Delta v_P^{-1}$ , respectively.  $v_P$  and  $\Delta v_P$  are the P-wave speed and its uncertainty. The last column indicates the 95 % confidence interval on the wave-speed estimate. A  $\frac{1}{\sigma^2}$ -weighted mean wave speed is calculated from the single wave-speed estimates and error bars. A pooled regression is computed for comparison with the weighted mean. In some cases a static shift has been applied to groups of picks for individual events in order to reduce the scatter between events.

event id	$T_i$	$v_P^{-1}$	$\Delta T_i$	$\Delta v_P^{-1}$	$v_P$	$\Delta v_P$	95~%
	(s)	(s/km)	(s)	(s/km)	$(\rm km/s)$	$(\rm km/s)$	$(\rm km/s)$
1	1.55	0.146	3.40	0.021	6.85	0.08	0.97
2	-0.90	0.156	1.59	0.008	6.41	0.03	0.31
3	-0.54	0.155	2.24	0.011	6.45	0.04	0.47
4	0.10	0.148	6.82	0.039	6.76	0.14	1.78
6	1.16	0.143	6.52	0.038	7.00	0.15	1.85
weighted mean ve	elocity:				6.44	0.07	0.18
pooled regression:	0.32	0.150	1.25	0.007	6.64	0.14	0.30

Fiordland events / linear regression results:

crustal phase

mantle phase							
event id	$T_i$	$v_P^{-1}$	$\Delta T_i$	$\Delta v_P^{-1}$	$v_P$	$\Delta v_P$	95~%
	(s)	(s/km)	(s)	(s/km)	$(\rm km/s)$	$(\rm km/s)$	$(\rm km/s)$
1	9.86	0.110	1.70	0.005	9.06	0.42	5.34
2	10.01	0.114	1.04	0.003	8.81	0.23	0.99
3	9.34	0.113	1.01	0.003	8.89	0.23	1.00
4	9.00	0.110	0.95	0.003	9.09	0.24	1.05
7	8.25	0.113	1.88	0.006	8.87	0.46	2.00
weighted mean ve	elocity:				8.92	0.07	0.21
pooled regression:	9.51	0.112	1.71	0.005	8.89	0.20	0.42
pooled regression	after co	orrecting f	or time	e offsets:			
(applied correction	ns are -	-0.1 s, -0	.6 s , -	-0.2 s, 0.8	s, -0.5 s	)	

Table A.3: continued

pplied	corrections	are -0.1	s, -0.6	s ,	-0.2  s,	$0.8 \ s$	s, -0.5  s)	
	1							

000		0.1 0, 0		0.2 0, 0.0	0, 0.0 0)		
	9.50	0.111	0.48	0.001	9.00	0.06	0.12

### Off-shore Cheviot event:

### mantle phase

event id	$T_i$	$v_P^{-1}$	$\Delta T_i$	$\Delta v_P^{-1}$	$v_P$	$\Delta v_P$	95~%
	(s)	(s/km)	(s)	(s/km)	$(\rm km/s)$	$(\rm km/s)$	$(\rm km/s)$
5	4.34	0.122	0.55	0.001	8.21	0.10	0.27

Table A.4: Previous studies are used to constrain the crustal structure of a 2D velocity model along the Southern Alps (SA) region (Fig. 2.3). LVZ: low-velocity zone within approximately 85 km east of the inferred Alpine Fault location. This LVZ is induced by high-pore fluid pressure in the mid-crust of the Southern Alps due to the release of fluids during prograde and strain-induced metamorphism (Stern *et al.*, 2001).

reference study	location	upper crust		middle crust		lower crust	
		z (km)	$v_P \ (\rm km/s)$	z (km)	$v_P \ (\rm km/s)$	z (km)	$v_P \ (\rm km/s)$
(Davey & Broadbent, 1980)	Fiordland	0–3	5.3	3-8	6.5 - 6.8	> 8	7.3
(Eberhart-Phillips & Reyners,	Fiordland	0-4	5.5	4-62.5			6.25 - 7.5
2001)							
(Scherwath et al., 2003)	SIGHT Transect 2	0 - 5	5.5	5-32	6.0 (LVZ) - 6.2	32-42	7.0
(Van Avendonk et al., 2003)	SIGHT Transect 1	0 - 5	5.5	5-30	6.0 (LVZ) - 6.2	30-35	6.8
(Reyners et al., 1993)	north Canterbury	2–11	5.7	11-20	6.24	20 - 27	7.1
This study 2D model	Fiordland	0 - 5	5.5	5-10	6	10	6.8
	central SA $(T2)$	0 - 5	5.5	5-32	6	32-42	6.8
	southern SA (Wanaka)	0 - 5	5.5	5–38	6	38-48	6.8
	East Coast	0 - 5	5.5	5-20	6	20-27	6.8

### Appendix B

# **P-wave anisotropy**

This appendix is a summary of the general P anisotropy (Cerveny, 2001; Farra and Psencik, 2003; Zheng, 2004; Zheng and Psencik, 2002) and Pn anisotropy (Backus, 1965) derivation from the Christoffel equation. Also included is a derivation of the equations for the specific cases of orthorombic and hexagonal symmetries with horizontal fast axes and their application to the teleseismic travel-time delays of Chapter 5.

# B.1 Christoffel equation and small perturbations of the medium

The Christoffel equation is the basis for the calculation of wave speeds and polarisation vectors of plane waves in homogeneous isotropic, anisotropic and also dissipative media (Cerveny, 2001). The Christoffel equation derives from the wave equation and is given by

$$(\Gamma_{il} - G_m \delta_{il}) \mathbf{P}_i^{(m)} = 0$$
 (m = 1, 2, 3) (B.1)

where  $\Gamma_{il} = \mathbf{a}_{ijkl}\mathbf{n}_j\mathbf{n}_k$  is the Christoffel matrix,  $\mathbf{a}_{ijkl}$  is the density-normalised elastic tensor,  $\mathbf{n}_j$  is a unit vector in the direction of the wavefront normal.  $G_m$  and  $\mathbf{P}_i^{(m)}$ (m=1, 2, 3) are the three eigenvalues and eigenvectors of the Christoffel matrix, which respectively equal the squares of the wave speeds,  $v_m^2$ , and the mutually orthogonal polarisation vectors of the three plane wave solutions.

In isotropic media, the P and S wave polarisation vectors are coincident with the dynamic axes formed by the propagation vector and wavefront, respectively. The two S wave solutions are degenerate, which means that the S wave polarisation vectors can be in any orthogonal transverse directions (e.g. Keith and Crampin, 1977). In anisotropic media, in contrast, the three eigenvalues are non-degenerate. The Christoffel equation has three solutions corresponding to the qP, qS1 and qS2 phases with distinct wave speeds,  $v_m$ , and mutually orthogonal and fixed polarisation vectors,  $\mathbf{P}^{(m)}$ .

In weakly anisotropic media, perturbations of the medium are small, and can thus be approximated by the first-order terms of their power series. Second- and higher-order terms are neglected. The perturbed elastic parameters  $\mathbf{a}_{ijkl}$  are, therefore, expressed as

$$\mathbf{a}_{ijkl} = \mathbf{a}_{ijkl}^0 + \Delta \mathbf{a}_{ijkl} \tag{B.2}$$

where  $\mathbf{a}_{ijkl}^{0}$  is the reference value in the isotropic case and  $\Delta \mathbf{a}_{ijkl}$  is its perturbation in the anisotropic medium. Similarly, other parameters of the Christoffel equation  $\Gamma_{il}$ ,  $G_m$  and  $\mathbf{P}_i^{(m)}$  can be expressed as sums of their isotropic values with their respective perturbations:

$$\Gamma_{il} = \Gamma_{il}^0 + \Delta \Gamma_{il} \tag{B.3}$$

$$G_m = G_m^0 + \Delta G_m \tag{B.4}$$

$$\mathbf{P}_i^{(m)} = \mathbf{P}_i^{(m)0} + \Delta \mathbf{P}_i^{(m)} \tag{B.5}$$

The eigenvalues  $G_m^0$  are solutions of the Christoffel equation in the isotropic medium. The eigenvalues are  $c_P^2$ , the square of the P wave speed, and  $c_S^2$ , the square of the S wave speed, i.e. the two degenerate eigenvalues of the isotropic case.  $\Delta\Gamma_{il}$ , the perturbation of the Christoffel matrix, is due to perturbations in the elastic parameters,  $\Delta \mathbf{a}_{ijkl}$ , and in the normal to the wavefront,  $\Delta \mathbf{n}_j$ .  $\Delta \Gamma_{il}$  is thus:

$$\Delta\Gamma_{il} = \Delta \mathbf{a}_{ijkl} \mathbf{n}_j \mathbf{n}_l + \mathbf{a}_{ijkl}^0 (\mathbf{n}_j \Delta \mathbf{n}_l + \mathbf{n}_l \Delta \mathbf{n}_j). \tag{B.6}$$

Substituting the expressions of  $G_m$ ,  $\Gamma_{il}$  and  $\mathbf{P}_i^{(m)}$  in the Christoffel equation (B.1) and keeping first-order terms yields

$$\underbrace{(\Gamma_{il}^{0} - G_{m}^{0}\delta_{il})\mathbf{P}_{l}^{(m)0}}_{0} + (\Gamma_{il}^{0} - G_{m}^{0}\delta_{il})\Delta\mathbf{P}_{l}^{(m)} + (\Delta\Gamma_{il} - \Delta G_{m}\delta_{il})\mathbf{P}_{l}^{(m)0} = 0.$$
(B.7)

### **B.2** General solution for the P phase

The P phase index is arbitrarily taken as m = 1. Substituting m in Equation (B.7) results in

$$(\Gamma_{il}^0 - G_1^0 \delta_{il}) \Delta \mathbf{P}_l^{(1)} + (\Delta \Gamma_{il} - \Delta G_1 \delta_{il}) \mathbf{P}_l^{(1)0} = 0.$$
(B.8)

From the requirement of  $\mathbf{P}_{i}^{(m)}$  to be unit vectors, the perturbation  $\Delta \mathbf{P}_{i}^{(1)}$  is perpendicular to the polarisation vector  $\mathbf{P}_{i}^{(1)0}$  of the isotropic case.  $\Delta \mathbf{P}_{i}^{(1)}$  can therefore be expressed as a linear combination of vectors  $\mathbf{P}_{i}^{(2)0}$  and  $\mathbf{P}_{i}^{(3)0}$ 

$$\Delta \mathbf{P}_i^{(1)} = c_2 \mathbf{P}_i^{(2)0} + c_3 \mathbf{P}_i^{(3)0} = 0 \tag{B.9}$$

and

$$\mathbf{P}_{i}^{(1)0} \Delta \mathbf{P}_{i}^{(1)} = 0. \tag{B.10}$$

The above condition (B.10) can be used to simplify equation (B.8) by multiplying it with  $\mathbf{P}_{i}^{(1)0}$ :

$$(\Delta\Gamma_{il} - \Delta G_1 \delta_{il}) \mathbf{P}_l^{(1)0} \mathbf{P}_i^{(1)0} = 0.$$
(B.11)

 $\Delta G_1$  is the perturbation of the square of the P wave speed,  $v_1^2 - c_P^2$ , also note that

$$\Delta G_1 = B_{11},\tag{B.12}$$

where B11 is an element of the so called weak anisotropy matrix with elements  $B_{mn}$ . Because the P polarisation vector  $\mathbf{P}_i^{(1)0}$  of the isotropic case is taken as a unit vector and parallel to the vector  $\mathbf{n}_i$ ,  $B_{11}$  can be rewritten as

$$B_{11} = \Delta a_{ijkl} n_i n_j n_k n_l. \tag{B.13}$$

### B.3 P anisotropy in the horizontal plane (Backus equation)

For the P headwave propagating in the horizontal plane, the propagation vector has two non-zero components

$$(n_1, n_2, n_3) = (\cos(\varphi), \sin(\varphi), 0)$$

with  $\varphi$  the P wave propagation azimuth. The weak-anisotropy element  $B_{11}$  is a trigonometric polynomial of degree 4 in  $\varphi$ , whose expression in the case of general anisotropy is (Backus, 1965):

$$B_{11} = \Delta a_{1111} cos^4(\varphi)$$

$$+ 4\Delta a_{1112} cos^3(\varphi) sin(\varphi)$$

$$+ (2\Delta a_{1122} + 4\Delta a_{1212}) cos^2(\varphi) sin^2(\varphi)$$

$$+ 4\Delta a_{1222} cos(\varphi) sin^3(\varphi) + \Delta a_{2222} sin^4(\varphi). \qquad (B.14)$$

This polynomial's Fourier expansion results in a more well-known formula (Backus, 1965)

$$\alpha^{2}(\varphi) - \alpha_{0}^{2} = A + C\cos(2\varphi) + D\sin(2\varphi) + E\cos(4\varphi) + F\sin(4\varphi), \tag{B.15}$$

#### B.3. P ANISOTROPY IN THE HORIZONTAL PLANE (BACKUS EQUATION)223

with  $\alpha(\varphi)$ , the P-wave speed for azimuth  $\varphi$ , and

$$8A = 3\Delta a_{1111} + 2\Delta a_{1122} + 4\Delta a_{1212} + 3\Delta a_{2222}$$
  

$$2C = \Delta a_{1111} - \Delta a_{2222}$$
  

$$D = \Delta a_{1112} + \Delta a_{1222}$$
  

$$8E = \Delta a_{1111} - 2\Delta a_{1122} - 4\Delta a_{1212} + \Delta a_{2222}$$
  

$$2F = \Delta a_{1112} - \Delta a_{1222}.$$
  
(B.16)

A minimum of five wave speeds measured along non-colinear profiles is necessary to solve for the five parameters A to F. A minimum of seven wave speeds is necessary in the case of a dipping Moho (Backus, 1965).

Based on the  $2\varphi$  dependence displayed by real wave-speed measurements, Smith and Ekström (1999) proposed following approximation for Equation B.15:

$$\alpha(\varphi) = \alpha_0 + C\cos(2\varphi) + D\sin(2\varphi), \tag{B.17}$$

Solving for the three unknown parameters,  $\alpha_0$ , the average Pn speed, and both constants, C and D of Equation (B.17), requires only three known wave speeds measured along intersecting profiles. If only two Pn-speed estimates,  $\alpha_1$  and  $\alpha_2$ , are known along two intersecting profiles, a third equation is needed to solve for the three unknowns  $\alpha_0$ , C and D. A third equation,  $\frac{d\alpha(\varphi)}{d\varphi}|_{\varphi=\Phi} = 0$ , is found by assuming  $\alpha(\varphi)$  is maximum for the fast propagation azimuth,  $\Phi$ , from a nearby SKS-splitting fast polarisation orientation. The resulting system of three equations is

$$\alpha_{1} = \alpha_{0} + C\cos(2\varphi_{1}) + D\sin(2\varphi_{1})$$

$$\alpha_{2} = \alpha_{0} + C\cos(2\varphi_{2}) + D\sin(2\varphi_{2})$$

$$0 = C\sin(2\Phi) - D\cos(2\Phi),$$
(B.18)

where  $\alpha_1$  and  $\alpha_2$  are the Pn-speed estimates along two intersecting profiles with azimuths  $\varphi_1$  and  $\varphi_2$ , respectively, and  $\Phi$  is the fast propagation orientation assumed from a nearby SKS fast polarization azimuth. The three unknowns  $\alpha_0$ , C and D represent the isotropic wave speed, i.e. the average wave speed, and the two anisotropy constants of Equation (B.17). The solutions are:

$$C = \frac{\alpha_1 - \alpha_2}{G} cos(2\Phi)$$
  

$$D = \frac{\alpha_1 - \alpha_2}{G} sin(2\Phi)$$
  

$$\alpha_0 = \alpha_1 - Ccos(2\varphi_1) - Dsin(\varphi_1)$$
  
(B.19)

with  $G = \cos(2\Phi) \left[\cos(2\varphi_1) - \cos(2\varphi_2)\right] + \sin(2\Phi) \left[\sin(2\varphi_1) - \sin(2\varphi_2)\right]$ 

Standard deviations,  $\sigma_C$  and  $\sigma_D$ , depend on the uncertainties in wave speeds and the fast azimuth,  $\sigma_{\alpha_1}$ ,  $\sigma_{\alpha_2}$  and  $\sigma_{\Phi}$ , which propagate into the solutions of C and D. Uncertainties  $\sigma_{\alpha_1}$ ,  $\sigma_{\alpha_2}$  and  $\sigma_{\Phi}$  being uncorrelated, variances  $\sigma_C^2$  and  $\sigma_D^2$  are as following:

$$\sigma_{C}^{2} = \left[\frac{\cos(2\Phi)}{G}\right]^{2} (\sigma_{\alpha_{1}}^{2} + \sigma_{\alpha_{2}}^{2})$$

$$+ \left[\frac{\alpha_{1} - \alpha_{2}}{G} \left(-2\sin(2\Phi) - \frac{\cos(2\Phi)}{G}\frac{\partial G}{\partial \Phi}\right)\right]^{2} \sigma_{\Phi}^{2}$$

$$\sigma_{D}^{2} = \left[\frac{\sin(2\Phi)}{G}\right]^{2} (\sigma\alpha_{1}^{2} + \sigma_{\alpha_{2}}^{2})$$

$$+ \left[\frac{\alpha_{1} - \alpha_{2}}{G} \left(2\cos(2\Phi) - \frac{\sin(2\Phi)}{G}\frac{\partial G}{\partial \Phi}\right)\right]^{2} \sigma_{\Phi}^{2}$$
(B.20)
(B.21)

with

$$\frac{\partial G}{\partial \Phi} = -2\cos(2\Phi) \left[ \sin(2\varphi_1) + 2\sin(2\varphi_2) \right] - 2\sin(2\Phi) \left[ \cos(2\varphi_1) - \cos(2\varphi_2) \right].$$

and  $\sigma_{\alpha_0}^2$ , the variance of the isotropic wave speed, is:

$$\sigma_{\alpha_0}^2 = \sigma_{\alpha_1}^2 + \cos^2(2\varphi_1)\sigma_C^2 + \sin^2(2\varphi_1)\sigma_D^2 + 2\cos(2\varphi_1)\sin(2\varphi_1)\sigma_{CD}^2$$
(B.22)

Note that in the case where the fast orientation,  $\Phi$ , is averaged over a whole region, the error,  $\sigma_{\Phi}$ , is small as is the contribution of the second term of Equations (B.20) and (B.21) to uncertainties in *C* and *D*.

### B.4 P anisotropy for orthorombic and hexagonal symmetries with horizontal fast axis

Expressions of the  $B_{mn}$  elements of the weak anisotropy matrix have been given in the case of general anisotropy and for P and S waves propagating in a three-dimensional space (Farra and Psencik, 2003; Zheng, 2004; Zheng and Psencik, 2002).  $B_{11}$  depends on 15 independent parameters, which requires a minimum of 15 observations. In the following, two expressions are derived to calculate the three-dimensional anisotropy by taking Equation (B.17) as derived by Smith and Ekström (1999) and assuming anisotropy with orthorombic and hexagonal symmetries with horizontal fast propagation axes of azimuth  $\Phi$ .

As described in the former paragraph, from the requirement that the P-wave speed is maximum along the fast propagation azimuth,  $\Phi$ , follows

$$Csin(2\Phi) - Dcos(2\Phi) = 0$$

and

$$2\Phi = \arctan\left(\frac{D}{C}\right).$$

The above expression for  $\Phi$  is substituted in the Equation (B.17) that, using trigonometric identities, is reshaped as:

$$\alpha(\varphi) = \alpha_0 + \sqrt{C^2 + D^2} \cos(2\varphi). \tag{B.23}$$

with

$$\sqrt{C^2 + D^2} = (\alpha_{max} - \alpha_{min})/2,$$

where  $\alpha_{max}$  and  $\alpha_{min}$  are the maximum and minimum wave speeds. Thus, by fixing a fast propagation azimuth, the number of unknowns in Equation (B.17) is reduced to the parameters  $\alpha_0$  and  $M = \sqrt{C^2 + D^2}$ .

In the first stage, orthorombic symmetry is assumed for the anisotropy from the alignment of olivine minerals under simple shear with a horizontal axis and small pure shear in the vertical plane. The minimum and maximum wave speeds are therefore contained in the horizontal plane, and the isotropic wave speed, i.e. the average wave speed, is assumed along the vertical direction. Taking Equation (B.17) for the azimuth-dependent wave speed,  $\alpha(\varphi)$ , in the horizontal plane, the variation of the wave speed with the incidence angle,  $\theta$ , is:

$$\alpha(\varphi,\theta) = \alpha_0 + \frac{M}{2} \cos(2\varphi - 2\Phi) \left[1 - \cos(2\theta)\right].$$
(B.24)

Second, hexagonal symmetry with a horizontal axis is assumed for the anisotropy from the alignment of olivine minerals under simple shear along a horizontal axis. The wave speed is, therefore, maximum for propagation parallel to the symmetry axis of azimuth,  $\Phi$ , while the wave speed is minimum for propagation in the perpendicular plane. The wave speed may be expressed as:

$$\alpha(\varphi, \theta) = \alpha_0 + \frac{M}{2} \cos(2\varphi - 2\Phi) \left[1 - \cos(2\theta)\right] - \frac{M}{2} \left[1 + \cos(2\theta)\right].$$
(B.25)

### B.4.1 Estimate of the maximum anisotropy using Chapter 5 wavespeed anomaly estimates

As described in Section B.3, the fast propagation azimuth is constrained to the fast polarisation azimuth,  $\Phi$ , of a nearby SKS-splitting measurement. The fast azimuth is taken to be  $\Phi \sim 29^{\circ}$  (Klosko *et al.*, 1999, station MTCA), and eight wave-speed anomaly estimates are used (Tab. B.1), which are assumed to be relative to a wavespeed of 8.1 km/s (IASP91; Kennett and Engdahl, 1991). These estimates define a system of linear equations that are solved through least-squares techniques (Bevington, Assuming anisotropy with orthorombic symmetry, the isotropic wave speed and the maximum anisotropy are ~8.6 km/s and ~11 %, respectively; the model misfit is 0.25 km/s (RMS in Tab. B.2). Assuming hexagonal symmetry, results in a lower anisotropy of ~9 % than in the orthorombic case and a larger isotropic wave speed of ~8.9 km/s (Tab. B.2). The misfit is 0.22 km/s and, thus, is similar in value to that of the orthorombic case.

Both orthorombic and hexagonal symmetries suggest P anisotropy that is consistent with Pn anisotropy estimates for the central Southern Alps:  $11 \pm 2$  % at the intersection of SIGHT Transect 2 with Transect 3 (Scherwath *et al.*, 2002); and 7–13 % at the intersection of SIGHT Transect 2 with the Fiordland-Cheviot profile (Chapter 2). The isotropic wave speed of ~8.9 km/s calculated for the hexagonal symmetry, in contrast, appears unreasonably large if it were to be explained by thermal contraction of mantle rocks produced by the downwarp of isotherms and mantle shortening alone. A ~8.6 km/s calculated for the orthorombic symmetry may represent a better estimate for the isotropic wave speed.

Event	Honshu	Banda	Irian Jaya	$5,\!9,\!18,\!19$	$17,\!21$	31,33	22,32	1,28
$\delta v_P \ ({\rm km/s})$	0.5	0.3	0.6	0.3	1.1	0.5	0.5	0.5
heta (°)	23	30	23.6	32.5	52	8	40	17.5
$\varphi$ (°)	-26	-61	-52	-45	-2	-66	27	-30

Table B.1: Wave-speed anomaly estimates  $(\delta v_P)$  that are included in the calculation of the maximum anisotropy.  $\theta$  is the incidence angle, and  $\varphi$  is the azimuth from north.

symmetry	$\alpha_0$	$\sigma_{lpha_0}$	M	$\sigma_M$	$\delta P$	$\sigma_{\delta P}$	RMS
	$(\rm km/s)$	$(\rm km/s)$	$(\rm km/s)$	$(\rm km/s)$	(%)	(%)	$(\rm km/s)$
orthorombic	8.6	0.09	0.4719	0.3874	10.9	9.0	0.25
hexagonal	8.93	0.18	0.3961	0.2133	8.9	4.6	0.22

Table B.2: Calculation of the maximum anisotropy from wave-speed estimates of Table B.1 for both orthorombic and hexagonal symmetries with a horizontal fast propagation orientation fixed to the assumed azimuth of  $\Phi = 29^{\circ}$  (station MTCA; Klosko *et al.*, 1999). Parameters are the isotropic wave speed,  $\alpha_0$ , the anisotropy parameter M, the maximum anisotropy,  $\delta P$ , their respective standard deviations and the model misfit given as the RMS (Root Mean Square).

# Appendix C

# P-wave travel-time delays



Figure C.1: Honshu Island earthquake and picks. Picked is not the first arrival but the more prominent and more continuous phase across the array. A band-pass filter is used that is acausal and has cut-off and corner frequencies of 0.1-0.5-1.5-2.5 Hz. The reduced wave-speed is based on ak135-predicted move-out across SIGHT Transect 2.



Figure C.2: Banda Sea earthquake and picked first arrivals. Traces are band-pass filtered at cut-off and corner frequencies of 0.1-0.5-1.5-2.5 Hz.



Figure C.3: Irian Jaya earthquake and picks. Traces are band-pass filtered at cut-off and corner frequencies 0.01-0.03-1.0-2.0 Hz.



Figure C.4: Off coast Chile earthquake and picks. Traces are band-pass filtered at cut-off and corner frequencies 0.05-0.1-1.0-2.0 Hz.

			Hypocentre	Epicentral	Back			
Event	Date	Julian day	Time	Latitude (°)	Longitude (°)	z (km)	distance (°)	azimuth (°)
Honshu Island	1996/02/14	45	21:26:57.33	29.2395	140.6366	151.5	77.2 - 78.5	-27.025.9
Banda Sea	1996/02/17	48	10:18:04.00	-6.9767	125.2151	549.9	53.9 - 54.7	-61.060.2
Irian Jaya	1996/02/17	48	20:17:46.90	-0.8884	136.106	9.4	52.0 - 53.4	-45.744.7
Off coast Chile	1996/02/19	50	07:10:07.00	-42.038	-75.2682	7.6	75.3 - 76.7	135.1 - 136.1

Table C.1: Teleseisms recorded during SIGHT (Fig. C.1–C.4) and used by Stern et al's study 2000.

Table C.2: Phase parameters of the SIGHT events based on ak135 Earth model Kennett *et al.* (1995).  $i_c$  is the incidence angle at the Moho assuming a mantle wave speed of 8.1 km/s.

Event	Epicentral	Back	Phase	$\mathrm{d}T/\mathrm{d}\Delta$	$i_c$	Move-out	ic projected onto
	distance (°)	azimuth (°)		$(s/^{\circ})$	$(^{\circ})$	$(\rm km/s)$	SIGHT T2 (°)
Honshu Island	77.2 - 78.5	333.0 - 334.1	Р	5.5	23.52 - 23.95	21.63	21.99
Banda Sea	53.3 - 54.7	299.0 - 299.8	Р	7.0	30.62 - 31.08	16.17	30.06
Irian Jaya	52 - 53.4	314.3 - 315.33	Р	7.3	32.37 - 32.87	14.98	32.73
Irian Jaya	52 - 53.4	314.3 - 315.33	sP	7.4	32.48 - 32.98	15.00	32.69
Off coast Chile	75.3 - 76.7	135.1 - 136.1	Р	5.6	24.29 - 24.77	-19.45	24.61

Event	Origin time			Hypocentre			Epicentral	Back
ID	Date	Julian day	Time	Latitude (°)	Longitude (°)	Depth (km)	distance (°)	azimuth (°)
1	2003	231	08:46:59.00	-5.8703	146.9950	77.0	42.9 - 43.1	325.0 - 325.1
2	2003	232	16:42:40.00	-11.4389	166.2160	55.9	32.4 - 32.6	352.7 - 352.9
3	2003	234	21:35:34.14	-13.3700	167.1400	159.2	30.4 - 30.6	354.0 - 354.3
4	2003	235	02:45:59.00	-0.8340	133.7510	32.1	53.6 - 53.8	312.5 - 312.6
5	2003	257	18:45:53.00	18.7207	121.0290	52.4	76.6 - 76.9	312.5 - 312.6
6	2003	264	18:16:12.00	19.8741	95.6346	4.5	92.9 - 93.1	294.7 - 294.9
7	2003	265	22:59:04.00	-5.0283	153.8590	112.4	41.2 - 41.4	334.8 - 334.9
8	2003	279	18:29:37.00	-10.7772	164.4550	30.7	33.2 - 33.5	349.6 - 349.8
9	2003	282	22:19:15.00	13.7255	119.9830	51.6	73.2 - 73.5	308.7 - 308.8
10	2003	290	10:19:09.00	-5.5037	154.1790	150.2	40.6 - 40.9	335.1 - 335.2
11	2003	291	22:27:15.00	0.4920	126.1060	51.0	58.9 - 59.1	305.7 - 305.9
12	2003	295	11:45:30.00	-6.0661	147.7630	45.6	42.4 - 42.6	325.8 - 325.9
13	2003	297	01:26:58.00	-3.4996	145.5050	33.0	45.7 - 45.9	324.3 - 324.4
14	2003	310	08:15:00.00	2.0057	127.0190	28.2	59.6 - 59.9	307.7 - 307.8

Table C.3: Teleseisms recorded during COOK (events 1–17) and WCOAST (events 18–35) deployments.

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Event	Origin time			Hypocentre			Epicentral	Back		
ID	Date	Julian day	Time	Latitude (°)	Longitude (°)	Depth (km)	distance (°)	azimuth (°)		
15	2003	310	10:38:05.00	-19.2786	168.8900	119.7	24.4 - 24.7	357.0 - 357.3		
16	2003	316	08:26:43.00	33.2416	137.0520	381.8	82.3 - 82.5	332.4 - 332.5		
17	2003	318	04:28:32.00	-16.6373	172.3080	2.6	27 - 27.2	4.4 - 4.7		
18	2003	344	04:38:14.00	23.0988	121.3840	27.6	79.7 - 80.1	315.4 - 315.6		
19	2003	345	00:01:45.91	22.72	121.52	10.0	79.4 - 79.7	315.3–315.5		
20	2003	360	08:18:32.00	-17.5018	-178.8230	548.5	27.6 - 27.8	23.2 - 24.0		
21	2004	003	16:23:17.00	-22.3037	169.6600	9.4	21.1 - 21.4	358.9 - 359.7		
22	2004	011	08:07:3.00	-16.3484	-176.1010	366.0	29.5 - 29.7	27.5 - 28.3		
23	2004	011	08:07:3.00	-16.3484	-176.1010	366.0	29.5 - 29.7	27.5 - 28.3		
24	2004	011	09:29:7.00	-20.2041	-179.1340	661.8	24.9 - 25.1	24.4 - 25.3		
25	2004	051	05:58:43.00	-11.6435	166.3890	76.6	31.8 - 32.1	353.1 - 353.7		
26	2004	053	06:46:26.00	-1.5802	100.4130	51.3	74.0 - 74.3	283.1 - 283.4		
27	2004	053	23:17:16.00	18.4743	145.6270	215.3	65.5 - 65.8	334.5 - 334.7		
28	2004	053	23:17:16.00	18.4743	145.6270	215.3	65.5 - 65.8	334.4 - 334.8		
Event	Origin time				Hypocentre	Epicentral	Back			
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ID	Date	Julian day	Time	Latitude (°)	Longitude (°)	Depth (km)	distance (°)	azimuth (°)		
29	2004	056	16:52:30.00	-20.7756	-174.9790	55.4	25.9 - 26.1	33.3 - 34.2		
30	2004	058	06:14:31.00	-34.0944	-112.4920	10.0	59-59.3	109.1 - 109.4		
31	2004	072	22:45:16.00	36.4092	70.7435	207.4	119.8 - 120.1	293.0 - 293.3		
32	2004	074	16:30:40.00	-17.2749	-172.2640	12.0	30.2 - 30.3	35.0 - 35.8		
33	2004	096	21:24:0.00	36.5576	71.0033	168.5	119.7 - 120.0	293.3 - 293.6		
34	2004	114	01:50:30.00	-9.5471	122.7330	88.1	53.0 - 53.3	295.1 - 295.3		
35	2004	124	04:36:50.00	-37.6950	-73.4060	21.0	80.6 - 81.0	133.9 - 134.2		
Events with Pn phases										
36	2003	233	12:12:47.0	-45.0874	167.0890	6.8	2.54	236.1 - 241.9		
37	2003	315	15:39:33.0	-30.6140	-179.1360	30.4	15.5 - 15.7	35.2 - 36.1		

159.7420

10.0

11.5 - 11.7

211.2 - 212.6

057

38

2004

22:58:39.0

-53.0830

Table C.3 continued

Table C.4: Picked phases for COOK teleseismic events and related parameters based on ak135 Earth model Kennett *et al.* (1995).  $i_c$  is the incidence angle at the Moho assuming a mantle wave speed of 8.1 km/s.

			DI			0 11
Event	Epicentral	Back	Phase	$dT/d\Delta$	$i_c$	Quality
ID	distance (°)	azimuth (°)		$(s/^{\circ})$	(°)	
1	42.9 - 43.1	-35.0	ScP	4.04	17.10 - 17.15	В
2	32.4 - 32.6	-7.2	Р	8.76	39.62 - 39.68	А
3	30.4 - 30.6	-5.8	Р	8.80	39.85 - 39.89	В
4	53.6 - 53.8	-47.4	Р	7.32	32.19 - 32.27	В
5	76.6 - 76.9	-47.4	Р	5.63	24.18 - 24.26	А
6	92.9 - 93.1	-65.2	pP	4.61	19.63 - 19.64	В
7	41.2 - 41.4	-25.1	Р	8.18	36.54 - 36.63	А
8	33.2 - 33.5	-10.3	Р	8.72	39.38 - 39.46	А
9	73.2 - 73.5	-51.2	Р	5.89	25.38 - 25.46	А
10	40.6 - 40.9	-24.9	Р	8.20	36.64 - 36.73	А
11	58.9 - 59.1	-54.2	Р	6.93	30.29 - 30.36	А
12	42.4 - 42.6	-34.2	Р	8.12	36.21 - 36.30	А
13	45.7 - 45.9	-35.7	pP	7.89	35.05 - 35.14	В
14	59.6-59.9	-52.3	pР	6.89	30.08 - 30.16	В
15	24.4-24.7	-2.9	Р	9.08	41.40 - 41.46	А
16	82.3 - 82.5	-27.5	PcP	4.39	18.65 - 18.66	А
17	27.0 - 27.2	4.6	sP	9.00	40.93 - 41.01	А
18	79.7 - 80.1	-44.5	pP	5.42	23.20 - 23.33	В
19	79.4 - 79.7	-44.6	pP	5.45	23.31 - 23.44	В
20	27.6 - 27.8	23.6	Р	8.74	39.52 - 39.57	С
21	21.1 - 21.4	-0.7	pP	10.79	51.70 - 51.93	В
	•					

Event	Epicentral	Back	Phase	$\mathrm{dT}/\mathrm{d}\Delta$	$i_c$	Quality
ID	distance (°)	azimuth (°)		$(s/^{\circ})$	(°)	
22	29.5 - 29.7	27.9	Р	8.76	39.64 - 39.68	В
23	29.5 - 29.7	27.9	PcP	2.60	10.91 - 10.96	С
24	24.9 - 25.1	24.8	Р	8.77	39.67 - 39.71	В
25	31.8 - 32.1	-6.6	Р	8.77	39.66 - 39.74	А
26	74.0 - 74.3	-76.7	Р	5.82	25.06 - 25.15	А
27	65.5 - 65.8	-25.4	Р	6.38	27.66 - 27.79	В
28	65.5 - 65.8	-25.4	PcP	4.15	17.59 - 17.62	В
29	25.9 - 26.1	33.7	Р	9.04	41.17 - 41.21	В
30	59.0 - 59.3	109.2	Р	6.93	30.27 - 30.38	D
31	119.8 - 120.1	-66.8	PKPdf	1.92	8.05 - 8.05	А
32	30.2 - 30.3	35.4	sP	8.84	40.10 - 40.12	С
33	119.7 - 120.0	-66.6	PKiKP	1.96	8.19 - 8.20	В
34	53.0 - 53.3	-64.8	Р	7.34	32.27 - 32.38	С
35	80.6 - 81.0	134.1	Р	5.34	22.85 - 22.97	D
36	2.54	-123.9118.1	pPn	13.75	90.0	С
37	15.5 - 15.7	35.2 - 36.1	pPn	13.14 - 13.18	73.24 - 73.77	А
38	11.5–11.7	-148.8147.4	sPn	13.68	85.16-85.28	$\mathbf{C}$

Table C.4 continued

Table C.5: Static corrections for sites of the COOK and WCOAST deployments are calculated for a reference elevation of 0 m and are based on Kleffman's (1999) velocity model for SIGHT Transect 2 shallow structure. Indices 1–3 denote surface gravels, Tertiary sediments and basement, respectively. The differences of the station static corrections *ttot* with those of the MCV station are the corrections applied to the travel-time delays. The error *terr* is dependent on the velocity uncertainty.

site	$h_1$ (m)	$Vp_1 \ (km/s)$	$h_2$ (m)	$Vp_2 \ (km/s)$	$h_3$ (m)	$Vp_3 \ (km/s)$	ttot (s)	ttot-tmcv (s)	terr (s)
GIL	24	1.6	0	0	0	0	-0.014	0.141	0.050
KAQ	0	0	61	3.7	0	0	-0.016	0.139	0.004
MAI	33	1.6	34	3.7	0	0	-0.030	0.125	0.100
KCJ	73	1.6	0	0	0	0	-0.046	0.109	0.010
WEL	0	0	0	0	467	5.4	-0.086	0.069	0.004
UCV	0	0	0	0	1175	5.4	-0.218	0.063	0.004
MCV	0	0	0	0	840	5.4	-0.155	0.	0.004

site	$h_1$ (m)	$Vp_1 \ (km/s)$	$h_2$ (m)	$Vp_2 \ (km/s)$	$h_3$ (m)	$Vp_3 \ (km/s)$	ttot (s)	ttot-tmcv $(s)$	terr(s)
UWN	0	0	0	0	635	5.4	-0.118	0.037	0.004
BIR	0	0	0	0	718	5.4	-0.133	0.022	0.004
BAP	57	1.37	229	3	311	5.5	-0.174	-0.019	0.01
MIP	171	1.37	286	3	158	5.5	-0.174	-0.019	0.01
BRM	171	2.00	286	3	246	5.5	-0.226	-0.071	0.05

Table C.5 continued



Figure C.5: Map view of delay times measured at the COOK and WCOAST arrays relative to ak135-predicted travel times (Kennett *et al.*, 1995) and to the average delay time at Maimai Creek (third triangle from the left). Triangles denote COOK and WCOAST stations. Positive (*red circles, delay*) and negative (*green crosses, advance*) travel-time delays are corrected for statics. Delays are projected near the station towards the direction of the incoming rays, with the distance to the station proportionnal to the incidence angle of the arrival and the symbol size proportional to the delay time.



Figure C.6: COOK events 1–6 and and predicted travel-time delays for models SUBW, VERT (*red curve*) and SUBE of Table 5.2 that have mantle bodies dipping 50° NW, vertical and dipping 50° SE, respectively. Time delays are presented relative to delays at station MCV and to the ak135 Earth model (Kennett *et al.*, 1995).



Figure C.7: Same as Figure C.6 for COOK events 7–12.



Figure C.8: Same as Figure C.6 for COOK and WCOAST events 13–18.



Figure C.9: Same as Figure C.6 for WCOAST events 19–26.



Figure C.10: Same as Figure C.6 for WCOAST events 27–32.



Figure C.11: Same as Figure C.6 for WCOAST events 33–35.

## Appendix D

# **Published** papers

- Crust and mantle thickening beneath the southern portion of the Southern Alps, New Zealand (2007), Bourguignon, S., Stern, T. A. and Savage, M. K., *Geophys. Jour. Int.*, Vol. 168, pp. 681-690;
- Crustal thickness and Pn anisotropy beneath the Southern Alps oblique collision, New Zealand (2007), Bourguignon, S., Savage, M. K. and Stern, T. A., in A Continental Plate Boundary: Tectonics at South Island, New Zealand, D. Okaya, T. A. Stern and F. J. Davey (eds), Geophysical Monograph Series, Vol. 175, American Geophysical Union, pp. 115–122.

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### Crust and mantle thickening beneath the southern portion of the Southern Alps, New Zealand

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#### SUMMARY

An 8.54  $\pm$  0.20 km s<sup>-1</sup> Pn speed is estimated on a line oriented ca. N5°E from the Alpine Fault beneath the Southern Alps of South Island based on a refraction experiment that uses the Fiordland earthquake of August 2003 as a source. This high Pn speed results from both strong anisotropy in the mantle lid of 7–13 per cent and a high Pn speed average of 8.3  $\pm$ 0.3 km s<sup>-1</sup>. A maximum crustal thickness of  $48 \pm 4$  km is calculated for the southern South Island near the town of Wanaka. This represents a crustal root of about 18 km, compared to measured crustal thicknesses at the east and west coasts of the South Island. The average topography in the southern Southern Alps is of the order of  $\sim 1000$  m, which is less than half that predicted by Airy isostasy for an 18 km crustal root. As recently proposed for the central South Island  $\sim$ 120 km to the north, we propose that thickened cold, and therefore more dense, mantle lithosphere exists beneath southern South Island, and that this excess of mass is an effective load that pulls down the overlying crust. The load is similar to that beneath the central Southern Alps, despite the predicted convergence across the Alpine Fault there being nearly twice that at Wanaka. Gravity modelling of crustal structure along a profile through Wanaka suggests that this mass excess has a minimum density contrast of  $35 \pm 5$  kg m<sup>-3</sup> between thickened mantle and asthenosphere, assuming an across-Moho density contrast of -300 kg $m^{-3}$ . We speculate that the reason for the enhanced thickening beneath Wanaka is that the subducted Australian Plate at the southwestern corner of the South Island acts like a backstop onto which Pacific mantle collides at  $\sim 26 \text{ mm yr}^{-1}$ , ca. 3/4 the full plate speed.

Key words: earthquake refraction, gravity modelling, lithospheric deformation, New Zealand, *Pn* anisotropy, oblique collision.

#### **1 INTRODUCTION**

In continental collision zones, we can readily observe how the crust thickens by reverse faulting and mountain growth. What is not so obvious is how shortening occurs in the mantle: as intracontinental subduction (Mattauer 1986) or continuous thickening of the entire lithosphere (England & Houseman 1986; Molnar 1992). In intracontinental subduction the elastic strength of the delaminating uppermost mantle provides resistance to bending and deformation is localized at the interface with the subducting plate. In contrast, in continuous thickening the mantle lithosphere is a continuum whose strength provides the resistance to deformation. Deformation is continuously distributed and accommodated in a ductile manner. Direct imaging of the mantle lithosphere beneath orogens and measurements of seismic anisotropy should provide an insight into mantle deformation. Both modes of shortening cause downwarp of isotherms in the mantle and, hence, produce a zone of wave speeds faster than the surrounding region. However, the geometry of the shortened mantle lid should differ between both modes, being either symmetric (continuous thickening) or asymmetric (intracontinental subduction). Similarly, if shortening is taking place along a narrow fault in the mantle (intracontinental subduction), seismic anisotropy should be more localized than if shortening is occurring as continuous thickening.

At the Australian–Pacific Plate boundary of the central South Island of New Zealand, the component of convergence has increased for the past 20 Myr (Cande & Stock 2004) and produced a mountain range, called the Southern Alps, along a pre-existing continental transform, the Alpine Fault. Here too it is debated how the continental mantle lithosphere accommodates shortening. Some consider that the transition between two subduction zones of opposite polarities, north and south of South Island, occurs as intracontinental subduction (Wellman 1979; Beaumont *et al.* 1996; Waschbusch *et al.* 1998; Beavan *et al.* 1999), while others prefer continuous thickening (Molnar *et al.* 1999; Stern *et al.* 2000). A further question is what is the reason for the absence of intermediate depth seismicity beneath the South Island collision zone (Anderson & Webb 1994).

The  $M_W$  7.2 Fiordland earthquake (2003 August 21) and aftershocks (Table 1, Fig. 1) enable us to analyse refraction traveltimes along the Southern Alps crustal root to measure properties of the mantle lid along the Southern Alps and perpendicular to the former

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**Table 1.** Events 1–4 and 6 are the Fiordland aftershocks re-located by Martin Reyners (GNS Science) using a temporary seismograph deployment and Eberhart-Phillips & Reyners' 1-D model for Fiordland (2001). Event 5 is the off-shore Cheviot event located by GeoNet using the standard 1-D model for New Zealand (Maunder 2001).

Event		Orig	in time		Magnitude		
Number	ID	Date	Time	Lat.	Long.	Depth (km)	ML
1	2105255	2003/08/25	03:36:30.26	-45.111	166.964	20.5	5.0
2	2106280	2003/08/26	23:56:26.57	-45.486	166.596	19.7	5.5
3	2106314	2003/08/27	01:29:40.41	-45.442	166.716	22.7	5.1
4	2106319	2003/08/27	01:42:54.12	-45.314	166.945	24.4	5.6
5	2106361	2003/08/27	03:39:37.23	-42.802	173.758	35	4.1
6	2110611	2003/09/04	08:40:44.25	-45.224	166.921	22.7	6.1



**Figure 1.** *Triangles:* deployed seismographs. *Stars:* earthquakes used in this study. *Solid lines:* previous SIGHT seismic profiles T1, T2, 3W and 4E (Okaya *et al.* 2002), and this study as the Fiordland–Cheviot refraction profile. *Bars:* SKS-splitting measurements (Klosko *et al.* 1999; Duclos *et al.* 2005). Lengths are proportional to delay times, while azimuths indicate fast-polarisation orientations. *Black circles with 4 arrows:* apparent values of *Pn* anisotropy at profile intersections T2/T3W (Scherwath *et al.* 2002), T1/T4E and T2/T4E (Baldock & Stern 2005), and this study's estimates of maximum *Pn* anisotropy.

South Island GeopHysical Transect (SIGHT) seismic experiment (Okaya *et al.* 2002). We determine the wave speed and infer the corresponding Pn anisotropy beneath the Southern Alps. We also estimate the crustal root thickness and model its gravity effect in order to define how much of the root is due to topographic loading and how much can be ascribed to the positive load of the subjacent shortened mantle lithosphere.

#### 2 EARTHQUAKE REFRACTION ANALYSIS

We deployed an eight-seismograph linear array in the eastern section of the Southern Alps, 5° clockwise from the strike of the range (Fig. 1). The array recorded five Fiordland aftershocks of  $M_L \ge 5$ 



**Figure 2.** Top:  $M_L$  6.1 Fiordland aftershock (event 6 of Table 1). Traces are bandpass filtered between 1 and 10 Hz. Pair of arrows indicate Pn first arrivals and ~1.5 s peg-leg. The uncertainty in Pn speed is given by its 95 per cent confidence interval. (Note that the third station from the left was not included in the regressions because of a timing error of the recording instrument). *Dashed curves*: predicted Pg and Pn traveltime curves for 2-D model (Fig. 3). Blow-up of the Pn first arrivals is in the top right corner. Bottom: same as in top figure for the  $M_L$  4.1 off-shore Cheviot event. Traces are bandpass filtered between 1.5 and 5 Hz. The Pn speed estimate is the result of a single regression. Note that the 2nd trace (from the left) was despiked, and one pick (3rd from the left) was not included in the regression.

at the SW-end and one  $M_L$  4.1 earthquake off-coast Cheviot, at the NE-end of the profile line (Table 1). Maximum epicentral distances of 490 km enabled us to pick Pg and Pn first arrivals and to analyse refraction traveltimes along the root of the Southern Alps. Pg and Pn apparent speeds were determined for single events, from the inverse of regression slopes on the first break picks (Fig. 2). We obtained a single speed value by weighting the individual speeds with their corresponding inverse standard deviations. The Pn speed of 8.54  $\pm$  0.20 km s<sup>-1</sup> and the apparent dip of 2.5  $\pm$  1.3° SW were determined by assuming a uniform dipping Moho along the refraction profile (e.g. Stein & Wysession 2003). Error bars presented are 95 per cent confidence intervals.

Crustal phases are seen in the Fiordland aftershock records (top of Fig. 2) but not in the record from Cheviot (bottom of Fig. 2).

© 2006 The Authors, *GJI*, **168**, 681–690 Journal compilation © 2006 RAS The determined Pg speeds of the Fiordland records are relatively high and show spatial variations with approximately 6.8 km s<sup>-1</sup> for events 1, 4 and 6, and approximately 6.4 km s<sup>-1</sup> for events 2 and 3 (Fig. 1). These wave speeds are slightly smaller, but consistent with values of 6.7–6.9 km s<sup>-1</sup> at 4–8 km depth and 7.1–7.4 km s<sup>-1</sup> from 8 km depth as determined from seismic refraction profiles (Davey & Broadbent 1980) in the exhumed Fiordland crustal block. They are also consistent with 6.25–7.5 km s<sup>-1</sup> from 4 to 62.5 km depth from 3-D inversion of local earthquakes (Eberhart-Phillips & Reyners 2001). The 6.4–6.8 km s<sup>-1</sup> wave speeds, however, are not representative of the lower 6.0–6.2 km s<sup>-1</sup> average *P*-wave speed in the Southern Alps mid-crust (Eberhart-Phillips & Bannister 2002; Scherwath *et al.* 2003; Van Avendonk *et al.* 2004).

The apparent *Pn* speeds determined from off-shore Cheviot and the reverse events in Fiordland are  $8.21 \pm 0.22$  and  $8.92 \pm$  $0.18 \text{ km s}^{-1}$ , respectively (Fig. 2). Taking a  $6.0-6.2 \text{ km s}^{-1}$  midcrustal wave speed (more representative of the Southern Alps crustal wave speed than the Fiordland  $6.8 \text{ km s}^{-1}$ ), that is, a 6.1- $6.23 \text{ km s}^{-1}$  average for the entire crust, results in an average of  $8.54 \pm 0.20 \text{ km s}^{-1}$  *Pn* speed and an apparent  $2.5 \pm 1.3^{\circ}$  SW dipping Moho. This is a conservative value compared to an apparent dip value of  $2.6-3.1^{\circ}$  SW for a  $6.4-6.8 \text{ km s}^{-1}$  crustal wave speed, but is also a much lower value than the apparent  $\sim 8^{\circ}$ SW dip calculated between where the Moho depths at transect T1 (Van Avendonk *et al.* 2004) and T2 (Scherwath *et al.* 2003) intersect with our refraction profile. The inconsistency may result from our assumption of a uniform dipping Moho.

### 3 VELOCITY MODEL AND CRUSTAL THICKNESS

Previous crustal studies in South Island (Table 2) are included to constrain the crustal structure (Fig. 3). Our refraction profile intersects SIGHT T2 15–30 km east of the maximum crustal thickness of  $44 \pm 1.4$  km (Scherwath *et al.* 2003). Here, the Moho is 42 km deep (Scherwath *et al.* 2003) and accordingly fixed to 42 km depth. Similarly, the Moho depth is fixed to 33 km at the intersection with SIGHT T1 (Van Avendonk *et al.* 2004). South of the crossing with SIGHT T2, the dip of the Moho is set to  $2.5^{\circ}$  and the upper-mantle wave speed is set to 8.54 km s<sup>-1</sup>, as determined above. We apply a forward modelling technique (Luetgert 1992) on this 2-D model. Rays propagating from both source locations, Fiordland and off-shore Cheviot, indicate that an approximately 150 km Moho portion, extending from Wanaka (southern SI) to Tekapo (central SI), is constrained. A maximum Moho depth of  $48 \pm 4$  km is estimated

near Wanaka at the southwestern tip of this zone (Fig. 3). Hence, the crustal root is  $18 \pm 4$  km thick at Wanaka (relative to average coastal values of 30 km in South Island, e.g. Godfrey *et al.* 2001; Melhuish *et al.* 2005) and ~4 km thicker than imaged along SIGHT T2 near Mount Cook (Scherwath *et al.* 2003) suggesting thickening of the crustal root from north to south along the Southern Alps.

#### 4 Pn ANISOTROPY

Knowing the Pn speed at two intersecting profiles, this study and SIGHT T2 (Scherwath *et al.* 2003), we infer the Pn anisotropy at the intersection.

Upper-mantle anisotropy is widely interpreted as being primarily due to lattice preferred orientation (LPO) of olivine minerals induced by finite strain (McKenzie 1979). Hence, shearing in the upper mantle can be deduced from anisotropy observations. Moreover, if we know both the total SKS splitting through the whole mantle and the value of anisotropy in the very top of the mantle lid, we can ask the question: is it reasonable that all the SKS splitting is due to shear in the ~100 km thick mantle lid rather than being distributed in the asthenosphere?

In the South Island, SKS-splitting observations (Klosko *et al.* 1999; Duclos *et al.* 2005) separate into a central and a southern domain. The boundary between these two domains is approximately coincident with SIGHT T2. In central South Island, fast propagation orientations,  $\Phi$ , of SKS are subparallel to the Alpine Fault, but in the southern South Island these are consistently oriented ~28° anticlockwise from the strike of the Alpine Fault, that is, the expected orientation of shear (Fig. 1) (Klosko *et al.* 1999; Molnar *et al.* 1999; Baldock & Stern 2005). Coincidence of fast polarisation azimuths and the deflection of crustal markers suggests lithospheric mantle and upper crustal deformation are broadly coupled in the South Island (Little *et al.* 2002). In central South Island, fast polarisations subparallel to the orientation of shear suggest some dynamic recrystallization component and/or higher strain rates than further south (Little *et al.* 2002; Scherwath *et al.* 2002; Savage *et al.* 2004).

Taking a *Pn* speed of  $8.0 \pm 0.2$  km s<sup>-1</sup> on the nearly perpendicular profile SIGHT T2 (Scherwath *et al.* 2003), and our result of  $8.54 \pm$ 0.20 km s<sup>-1</sup>, implies a  $6.1 \pm 5.0$  per cent apparent anisotropy. If we take the fast orientation of *Pn* propagation to be that of the SKS fast polarization orientation, we can rotate the intersecting profiles into fast and slow orientations to infer a maximum anisotropy. However, the  $\Phi$  of  $44^{\circ}$ , measured at the profile intersection (Klosko *et al.* 1999), lies at the turning point between the two distinct domains of SKS anisotropy. We infer that  $\Phi$  is likely to be an intermediate value

**Table 2.** Previous studies are used to constrain the crustal structure of a 2-D velocity model along the Southern Alps (*SA*) region (Fig. 3). *LVZ*: low-velocity zone within approximately 85 km east of the inferred Alpine Fault location. This LVZ is induced by high-pore fluid pressure in the mid-crust of the Southern Alps due to the release of fluids during prograde and strain-induced metamorphism (Stern *et al.* 2001).

Reference study	Location	Upper crust		Middle crust		Lower crust	
Reference study	Location	<i>z</i> (km)	$v_P ({\rm km}~{\rm s}^{-1})$	<i>z</i> (km)	$v_P ({\rm km}~{\rm s}^{-1})$	<i>z</i> (km)	$v_P ({\rm km}{\rm s}^{-1})$
Davey & Broadbent (1980)	Fiordland	0–3	5.3	3-8	6.5-6.8	>8	7.3
Eberhart-Phillips & Reyners (2001)	Fiordland	0-4	5.5	4-62.5			6.25-7.5
Scherwath et al. (2003)	SIGHT T2	0-5	5.5	5-32	6.0 (LVZ)-6.2	32-42	7.0
Van Avendonk et al. (2004)	SIGHT T1	0-5	5.5	5-30	6.0 (LVZ)-6.2	30-35	6.8
Reyners & Cowan (1993)	North Canterbury	2-11	5.7	11-20	6.24	20-27	7.1
This study 2-D model	Fiordland	0–5	5.5	5-10	6	10	6.8
	Central SA (T2)	0-5	5.5	5-32	6	32-42	6.8
	Southern SA (Wanaka)	0-5	5.5	5-38	6	38-48	6.8
	East Coast	0–5	5.5	5-20	6	20–27	6.8

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**Figure 3.** Top left: location map of the main divide (*dashed line*) and Fiordland–Cheviot (*thick line*) profiles along which mean elevations are calculated. *Middle*: mean topography (*thin curve*) and Bouguer gravity anomaly (*thick curve*) (Reilly & Whiteford 1979) in a 10 km wide band along the Fiordland–Cheviot profile (Fig. 1); mean topography along the main divide (*dashed curve*). Bottom: 2-D velocity model; *triangles*: deployment; *T1/T2*: intersections with SIGHT previous crustal studies; *dashed line*: constrained portion of the Moho between Wanaka and Tekapo; *interrogation marks*: unconstrained interfaces. Predicted traveltime curves are dashed curves of Fig. 2.

resulting from the overlap of Fresnel zones over the two domains of anisotropy: central South Island with mean  $\Phi$  of 56  $\pm$  2° and southern South Island with  $21 \pm 1^{\circ}$ . Distinct maximum Pn anisotropies are calculated for each domain. Given the first terms of the Taylor expansion of the azimuth-dependent wave speed,  $\alpha(\phi) =$  $\alpha_0 + \text{Ccos}(2\phi) + \text{Dsin}(2\phi)$  (Smith & Ekström 1999), an assumed fast propagation orientation and Pn speeds in the two crossing profile orientations, three equations are defined which enable solving for the three parameters:  $\alpha_0$ , C and D. Taking southern South Island mean  $\Phi$  of 21  $\pm$  1° implies a *Pn* anisotropy  $\delta P$  of 13.3  $\pm$ 3.5 per cent, an average Pn speed  $\alpha_0$  of 8.42  $\pm$  0.28 km s^{-1} and constants C and D of  $0.42 \pm 0.08$  and  $0.78 \pm 0.21$  km s<sup>-1</sup>, respectively. Taking a central South Island  $\Phi$  of 56  $\pm$  2° implies  $\delta P = 7.0 \pm$ 3.5 per cent,  $\alpha_0 = 8.25 \pm 0.24$  km s<sup>-1</sup>,  $B = -0.11 \pm 0.03$  km s<sup>-1</sup> and  $C = 0.27 \pm 0.15$  km s<sup>-1</sup>. 8.3  $\pm 0.3$  km s<sup>-1</sup> fits in the error bars of both average Pn speeds and the anisotropy is in the range 7-13 per cent.

The average Pn speed of  $8.3 \pm 0.3$  km s<sup>-1</sup> is consistent with a zone of fast Pn speed inferred in two former studies for the southern South Island: an average  $8.3 \pm 0.1$  km s<sup>-1</sup> from regional events (Haines 1979) and 8.4 km s<sup>-1</sup> from a joint hypocentre determination of intermediate depth earthquakes in Fiordland (Smith & Davey 1984).

While 7 per cent Pn anisotropy can be explained by finite strain alone, 13 per cent anisotropy requires infinite strain or additional dynamic recrystallization. If the calculated anisotropy at the top of the mantle lid is constant throughout the mantle lid, then a thickness of the anisotropic layer can be estimated from the observed SKSsplitting delay times. For the central South Island fast polarisation end-member, with 7 per cent Pn anisotropy, a layer of about 100 km thickness would account for the observed SKS-delay time of 1.76 s (Klosko et al. 1999) (assuming a P- to S-anisotropy ratio of 1.4 and a 4.7 km s<sup>-1</sup> average S-wave speed in the uppermost mantle). This thickness is only half as large, that is, 50 km, for the southern South Island end-member of 13 per cent Pn anisotropy. Comparison with parallel and crossing lines off-shore suggests similar anisotropy within error bars with  $6.5 \pm 3.0$  per cent, 230 km east of the surface trace of the Alpine Fault (Baldock & Stern 2005) and 11.5  $\pm$ 2.0 per cent on the Australian side, 30 km west of the Alpine Fault (Scherwath et al. 2002) (Fig. 1). Note that the 11.5  $\pm$  2.0 per cent Pn anisotropy on the Australian side and the here inferred 7-13 per cent Pn anisotropy on the Pacific side are at a similar distance to the Alpine Fault at Moho depth if we take the dip of the Alpine Fault to be 45° SE. The uncertainty of our anisotropy estimate doesn't allow us to predict how anisotropy correlates with distance from the Alpine Fault, while SKS delay times show no significant

correlation with distance at all. Overall, mantle anisotropy appears more widely distributed than upper-crustal deformation (Scherwath *et al.* 2003).

### **5 GRAVITY MODELLING**

Intracontinental subduction and continuous thickening both involve displacement of asthenosphere with colder mantle lithosphere and, therefore, downwarp of isotherms. Thus, in both models, the negative temperature contrasts and resultant thermal contraction within the upper mantle produce positive density contrasts, which appear as positive gravity anomalies.

The Southern Alps region exhibits a negative Bouguer gravity anomaly (Fig. 4b), as is usually observed above crustal roots that sustain the load of mountain ranges (Airy isostasy). However, a closer look at the Southern Alps shows that the topography (Fig. 4a) and the Bouguer anomaly (Fig. 4b) do not correlate well (Woodward 1979) and trend at different angles. Moreover, the mean elevations of ca. 1000 m, as seen in the Wanaka region (Fig. 3), should only require the support of a ca. 6-9 km thick crustal root underneath, if Airy load compensation and a -300 to -400 kg m<sup>-3</sup> density contrast between crustal root and mantle are assumed (e.g. Watts 2001). However, the crustal root is  $18 \pm 4$  km at Wanaka (relative to a 30 km coastal average in South Island). Therefore, at Wanaka the crust is at least two times thicker than needed to support the topography. The anomalous gravity effect of the overthickened crust, that is, the deviation from Airy load compensation, is visible in the negative isostatic anomaly of the Southern Alps region (Fig. 4c). We hypothesize that a mass excess exists in the mantle that pulls the crustal root down and maintains equilibrium by balancing part of the mass deficit of the crustal root. The positive gravity effect of such a mantle body, here called the mantle residual anomaly, is obscured by the large negative anomaly of the crustal root. As a result the observed Bouguer anomaly low is -85 mGal (Reilly & Whiteford 1979, Fig. 3) and less than that expected for an 18 km thick crustal root alone.

The present modelling aims at defining the minimum density contrast and the lateral and vertical extent of the mantle body that fits the gravity.

### 5.1 Model domain

We estimate the mantle residual anomaly along a profile crossing the South Island at Wanaka, herein called Jackson Bay-Dunedin profile (JB-D in Figs 4b–5). A  $2^{3/4}$ -D gravity modelling software (GM-SYS<sup>TM</sup>) is used that allows bodies of finite extent in the dimension perpendicular (*Y*-axis, Figs 4b–c) to the calculated gravity profile (*X*-axis, Figs 4b–c). The chosen gravity model is *ca*. 400 km long in the orientation parallel to our JB-D interpretation profile (*X*-axis, Figs 4b–c, 5a) and extends 200 km NE and 100 km SW from profile JB-D (*Y*-axis, Figs 4b–c and 5b).

### 5.2 Crustal model

Modelling is done relative to a reference crust of 30 km thickness above a mantle of 3300 kg m<sup>-3</sup> density. An average density contrast of -300 kg m<sup>-3</sup> is adopted for the crustal root. This value is less than the -450 kg m<sup>-3</sup> (Stern *et al.* 2000) density contrast estimated at SIGHT T2. As discussed below, using a density contrast of -450 kg m<sup>-3</sup> in this study results in an unreasonably large density contrast within the upper mantle. The SIGHT T2 mostly traverses greywacke, apart from a ~60 km wide strip of Alpine Schist directly south east of the Alpine Fault. In the mid and lower crust,

rocks are inferred to be greywacke/schist (Scherwath et al. 2003) and oceanic crust (Kleffman 1999), respectively. In contrast, the JB-D line is substantially within the Otago Haast Schist (greenschist facies). These schists represent the deeply exhumed part of a Mesozoic accretionary prism on the margins of Gondwana (e.g. Mortimer 2004). The JB-D line strikes parallel to the axis of an antiform that corresponds to the largest amount of exhumation within the Otago Schist. Here, the schists retain a first metamorphic event that reached temperatures and pressures up to 200-400°C and 4-8 kbar, respectively (Grapes & Watanabe 1992; Grapes 1995; Mortimer 2000) at estimated depths of 10-25 km. Away from the antiform both degree of metamorphism and exhumation decrease. Thus, we conclude that along the line JB-D rocks at a present depth of 30 km were once possibly  $\sim$ 50 km deep. At these depths and temperatures, continental crust starts to transform to eclogite (e.g. Wyllie 1992) of  $\sim$ 3550 kg m<sup>-3</sup> density (Hacker & Abers 2004). If the lower crust were to be partially transformed to eclogite, the across-Moho density contrast would be low or even absent.

The crustal root is 14 km thick at SIGHT T2 (Y = -140 km) and thickens to 18 km midway (Y = -100 km) between SIGHT T2 and profile JB-D (Fig. 5b). The crustal structure along profile JB-D is not known in detail and is constrained by only three points of known crustal thickness. The crustal thickness is ca. 30 km off-shore Jackson Bay (Melhuish et al. 2005),  $48 \pm 4$  km thick near Wanaka (this study) and ranges between 27 and 33 km off-shore from Dunedin (Godfrey et al. 2001). In between these three points the shape of the crustal root and the location of its deepest point are not well constrained. The simplest hypothesis is that the crustal root is asymmetric as imaged along SIGHT T1 (Van Avendonk et al. 2004) and T2 (Scherwath et al. 2003). However, geodetic strain-rates (Henderson 2003) and Holocene reverse faulting show that contraction occurs as far as eastern Otago (Fig. 4a) in the southern South Island (Norris & Cooper 2000) and is here more distributed than in central South Island (SIGHT T2). Indeed, low dip-slip angles on the southern section of the Alpine Fault (Jackson Bay) (Sutherland 1994) are associated with the widest zone of active deformation across the southern South Island (Norris & Cooper 2000). 3-D crustal structure obtained from simultaneous inversion of earthquake and shot arrival times and gravity data included below 20 km (Eberhart-Phillips & Bannister 2002, their Fig. 12) also suggests a wider crustal root in south than in north. Hence, the crustal root may be distributed further southeast from the Alpine Fault along profile JB-D than it is in the north (SIGHT T2). We, therefore, assume a symmetric crustal root with respect to the gravity minimum.

The maximum crustal thickness is fixed to 48 km at the intersection with the Fiordland–Cheviot profile (X = 100 km of profile JB-D) and the crustal root is symmetric with respect to the gravity minimum (X = 123 km, Figs 4b and 5a).

Due to the strong trade-off between shape of the crustal root and symmetry of the mantle body, the mantle residual anomaly, that is, the difference between the Bouguer anomaly and the modelled crustal root gravity effect, is symmetric (Fig. 6) and requires the presence of a symmetric positive mantle mass excess. In contrast, an asymmetric crustal root would require an asymmetric, that is, dipping, mantle body. In other words, we can place reasonable constraints on the mass excess of the mantle body, but not its shape.

### 5.3 Models for the mantle body

Assuming the mantle mass excess is a cylinder-type source (Nettleton 1976), a first-order maximum depth of 90–100 km is estimated for the centre of mass from the half-maximum of the





**Figure 4.** (a) Topography of South Island, SIGHT T2 and Jackson Bay-Dunedin profile (JB-D). Highest elevations (Mt Cook region) appear in yellow. (b) Bouguer gravity anomaly (adapted from Scherwath *et al.* 2002). *Rectangles*: domain of the gravity model used in  $2^{3/4}$ -D modelling and extent of model cross-sections taken along SIGHT T2 and profile JB-D (Fig. 5a); *hatched rectangle*: mantle body (model 1 of Table 3). (c) Isostatic gravity anomaly on-shore (DSIR) superposed on free-air anomaly off-shore (Seasat) and axis of the  $2^{3/4}$ -D model.

mantle gravity anomaly. The mass per unit length of strike of the mantle body is *ca*.  $2.7 \times 10^{11}$  kg m<sup>-1</sup> as determined by the mass balance between topography and crustal root. Thus, although the density contrast, width and depth extent of the mantle body are free

parameters, the mass excess is required to attain the above mass per unit length of  $2.7 \times 10^{11}$  kg m<sup>-3</sup> within 10 per cent. In addition, the combination of the Moho depth, that is, the minimum depth that the top of the mass excess can reach and the first-order depth



**Figure 5.** Cross-sections through gravity model 1 (Table 3). The thin oceanic crust, the crustal root and the mantle body are represented with their density contrasts determined relative to a reference crust of 30 km thickness and an average mid-lower crustal density of 3000 kg m<sup>-3</sup> above a mantle of 3300 kg m<sup>-3</sup> density. (a) Mean topography in a 10 km wide band, Bouguer gravity anomaly (Reilly & Whiteford 1979) and *X*-cross-section (Y = 0 km) taken along the Jackson Bay-Dunedin profile (JB-D). (b) *Y*-cross-section (X = 100 km in Figs 4b–c) taken perpendicular to profile JB-D. JB-D and T2 denote the intersections with crossing profiles.

estimate of the centre of mass, provides bounds to the vertical extent of the mantle body. A minimum density contrast is found for which the body's vertical dimension is maximum but contained within the vertical bounds mentioned above, and the resulting gravity effect satisfies the amplitude and wavelength of the Bouguer anomaly. For density contrasts smaller than this minimum, vertical stretch of the mantle body is necessary in order to fit the maximum amplitude of the gravity anomaly. However, because the gravity effect is proportional to  $r^{-2}$ , deeper mass is less effective in producing a gravity effect and more mass needs to be added than required to attain mass balance. Hence, there is no body found with density contrast below this minimum that can fulfil all requirements.

A minimum density contrast,  $\Delta\rho$ , of  $35 \pm 5$  kg m<sup>-3</sup> is required for a mantle body centred at 90–100 km depth in order to satisfy the mass balance and the wavelength of the gravity anomaly (model 1 of Table 3, Figs 5 and 6) with a misfit of the order of 10 mGal. The lateral and vertical dimensions of this body are  $110 \pm 20$  and  $70 \pm$ 20 km, respectively. The minimum density contrast ( $\Delta\rho$ ) is  $20 \pm$ 5 kg m<sup>-3</sup> for an across-Moho density contrast of -250 kg m<sup>-3</sup> (model 2 of Table 3) or  $55 \pm 5$  kg m<sup>-3</sup> for the across-Moho density contrast of -350 kg m<sup>-3</sup> (model 3 of Table 3). The minimum density contrast is even larger,  $\Delta\rho \sim 120$  kg m<sup>-3</sup>, if we adopt a crustal root contrast of  $-450 \text{ kg m}^{-3}$ , as assumed under SIGHT T2 (Stern *et al.* 2000). However, such a density contrast is far beyond the average of 60 kg m<sup>-3</sup> (e.g. Houseman *et al.* 2000) that can be considered as a reasonable maximum to be explained by isotherm deflections alone. Further chemical heterogeneities would be required within the mantle if we used such a large density contrast.

In summary, the mantle body is wider and less thick than previously inferred along the SIGHT T2 line (Stern *et al.* 2000) (model 0, Table 3), but provides a similar mass excess in the case of a crustal root with  $-300 \text{ kg m}^{-3}$  density contrast.

### 6 DISCUSSION

We interpret the 8.54  $\pm$  0.20 km s<sup>-1</sup> *Pn* speed estimated at *ca*. N5°E from the Alpine Fault to be the result of seismic anisotropy and high average wave speed in the mantle lid. Seismic anisotropy is recognised to be mostly the product of 850 km shear between the Pacific and the Australian Plates in the past 45 Myr (Molnar *et al.* 1999; Little *et al.* 2002; Savage *et al.* 2004; Baldock & Stern 2005). *Pn* speeds higher than the worldwide average upper-mantle wave speed of 8.1 km s<sup>-1</sup> (Kennett & Engdahl 1991, IASP91) are indicative of shortened and cold mantle beneath the Southern Alps



**Figure 6.** Bouguer gravity anomaly (Reilly & Whiteford 1979). Gravity anomalies are calculated for model 1 (Table 3, Fig. 5) using GM-SYS<sup>TM</sup>, a  $2^{3/4}$ -D gravity modelling software. These are the anomaly for a symmetric crustal root alone, that for the entire model (crustal root + mantle body), the mantle residual anomaly (Bouguer – crustal root) and the total misfit (Bouguer anomaly – entire model).

after 20 Myr (Cande & Stock 2004) continuously increasing convergence, as discussed below.

The average *Pn* speed of ~8.3 km s<sup>-1</sup> is a 2–3 per cent perturbation relative the upper-mantle wave speed of 8.1 km s<sup>-1</sup> (Kennett & Engdahl 1991). Taking  $\delta V_P / \delta T = 5 \times 10^{-4}$  km s<sup>-1</sup> °C<sup>-1</sup> (Anderson & Isaak 1995), the relationship between lateral wave speed variation and temperature, this perturbation of speed could be explained by a ~400 °C negative temperature contrast with the surrounding mantle. Similarly, the temperature contrast,  $\Delta T$ , caused by the deflection of isotherms can be estimated from the density contrast,  $\Delta \rho$ , with  $\Delta \rho = -\rho \alpha \Delta T$ . Taking  $\alpha = 3.5 \times 10^{-5}$  (Anderson *et al.* 1992) as the coefficient of thermal expansion and  $\rho = 3300$  kg m<sup>-3</sup> for the uppermost-mantle density, the equivalent average temperature contrast ranges from -170 to -480°C in the case of a 20– 55 kg m<sup>-3</sup> density contrast in the mantle lid, as suggested by the above gravity modelling.

The bulk of the inferred mantle body (Fig. 5) compares well with a zone of fast wave speed below the Southern Alps imaged by 3-D inversion of teleseismic traveltime residuals (Kohler & Eberhart-Phillips 2002, their Fig. 7). Along profile JB-D, their inversion displays a zone of anomalous mantle with *P*-wave speed perturbations of 1.5–3 per cent relative to 8.1 km s<sup>-1</sup> existing in a *ca.* 100 km wide zone located *ca.* 50 km offset east of the Alpine Fault. Similar to their 3-D inversion (Kohler & Eberhart-Phillips 2002) our modelling suggests that the anomalous mantle extends deeper north (Mt Cook region) than south (Wanaka region) (compare models 1 and 0 of Table 3).

Crustal roots of  $14 \pm 2$  km thickness in central South Island (Scherwath *et al.* 2003) and  $18 \pm 4$  km thickness in southern South Island (relative to a coastal average of 30 km) as well as mantle mass of similar excess beneath both regions seem, at first, counterintuitive with the total convergence across the Alpine Fault being *ca.* 40 km less (Cande & Stock 2004) and elevations *ca.* 500 m less across southern than central South Island. Lower crustal extrusion (Bird 1991) in an oblique convergent setting was suggested as a possible mechanism for maximum crustal thickening south east of the Southern Alps topographic maximum and at *ca.* 15° counterclockwise from the Alpine Fault (Gerbault *et al.* 2002). Although lower crustal extrusion is a possible explanation for the large crustal thickness beneath Wanaka, a further process is required that thickens the mantle lithosphere and provides the mass excess to fit the gravity anomaly beneath the Wanaka region.

Two observations let us speculate that the nearby Puysegur margin may contribute to thickening of the Pacific lithosphere of the southwestern South Island. First, hypocentres image north-eastward steepening of the Benioff zone in the Australian slab beneath Fiordland (Smith & Davey 1984; Reyners et al. 2002). Second, 3-D inversion of local-earthquake data indicates a zone of high mantle  $V_P$  ( $V_P > 8.5$  km s<sup>-1</sup>, that is, to 2–3 per cent faster  $V_P$  than the surrounding) beneath Fiordland that is east of and adjacent to the Australian slab (Eberhart-Phillips & Reyners 2001, their Fig. 6d). This high  $V_P$  zone shallows from 90 km depth beneath Fiordland to 60 km depth beneath the Southern Alps (Eberhart-Phillips & Reyners 2001), while the eastern extent is unresolved. We interpret this zone of high wave speed as the southernmost expression of the thickened Pacific mantle lithosphere. Here, the Australian slab may act as a rigid backstop, a buttress (Malservisi et al. 2003), which contributes to thickening of the Pacific mantle lithosphere of the southern South Island from the southwest. Projecting the 34 mm yr<sup>-1</sup> relative plate motion at the latitude of Fiordland (DeMets et al. 1994) onto the Australian slab (Reyners et al. 2002) results in a

**Table 3.** Gravity models are derived for an 18 km thick crustal root below a 30 km deep Moho, symmetric to X = 123 km and maximum between X = 100 and X = 146 km of profile JB-D (Figs 4b-c and 5a). The crustal root is thickened midway (Y = -100 km) from 14 km at SIGHT T2 to 18 km at JB-D (Fig. 5b). Density contrasts,  $\Delta \rho$ , in the range -250 to -350 kg m<sup>-3</sup> are adopted for the crustal root. The minimum density contrast and dimensions of a mantle anomaly are varied with the requirement to fit the Bouguer anomaly (Reilly & Whiteford 1979) (Fig. 6) and to attain the mass balance between topography, crustal root and mantle body within 10 per cent:  $\delta m$ (mantle) +  $\delta m$ (topography)  $\approx -\delta m$ (crustal root).  $X_1 - X_2$  is the lateral extent of the mantle body,  $Z_1 - Z_2$ , the depth range (Fig. 5a),  $A = (X_2 - X_1)(Z_2 - Z_1)$ , the cross-section and  $\delta m = A \cdot \Delta \rho$ , the linear mass excess of the mantle body. The temperature contrast,  $\Delta T$ , is calculated by assuming that the density contrast  $\Delta \rho$  is solely due to thermal contrast and taking  $\Delta \rho = -\rho \alpha \Delta T$  with  $\alpha = 3.5 \times 10^{-5}$  (Anderson *et al.* 1992) the coefficient of thermal expansion and  $\rho = 3300$  kg m<sup>-3</sup> the uppermost-mantle density.

Model	Topography $\delta m$ (kg m <sup>-1</sup> )	Root		Mantle body					
		$\Delta \rho$ (kg m <sup>-3</sup> )	$\delta m$ (kg m <sup>-1</sup> )	$\Delta \rho$ (kg m <sup>-3</sup> )	$\begin{array}{c} X_{1m} - X_{2m} \\ \text{(km)} \end{array}$	$\begin{array}{c} Z_1 - Z_2 \\ (\text{km}) \end{array}$	$A (m^2)$	$\delta m$ (kg m <sup>-1</sup> )	Δ <i>T</i> (°C)
1	$2.4 \times 10^{11}$	-300	$-5.1 \times 10^{11}$	+35	60-175	60-130	$8.0 \times 10^{9}$	$2.8 \times 10^{11}$	305
2	$2.4 \times 10^{11}$	-250	$-4.1 \times 10^{11}$	+20	50-200	75-125	$7.5 \times 10^{9}$	$1.5 \times 10^{11}$	170
3	$2.4 \times 10^{11}$	-350	$-5.9  imes 10^{11}$	+55	60–170	60-125	$7.1 \times 10^9$	$3.9 \times 10^{11}$	480
			Mantle body d	lerived at SIGI	HT T2 (Stern et a	<i>l.</i> 2000)			
0	$3.4 \times 10^{11}$	-450	$-5.5 \times 10^{11}$	+30	70–160	70–170	$9.0 \times 10^9$	$2.7 \times 10^{11}$	260

© 2006 The Authors, *GJI*, **168**, 681–690 Journal compilation © 2006 RAS convergence rate as large as 26 mm yr<sup>-1</sup>, that is, six times greater than the convergence rate perpendicular to the Alpine Fault at Jackson Bay. Alternatively, Malservisi et al. (2003) interpret a backstop wider than the slab inferred from seismicity (Reyners et al. 2002) onto which the Pacific mantle collides at the almost full plate speed of  $\sim$ 34 mm yr<sup>-1</sup>. Assuming that the Pacific mantle has been converging for 10-20 Ma at a rate of 26 mm yr<sup>-1</sup> onto the Australian slab, then the total shortening across the margin would be as large as 250-480 km. The length of the Australian slab Benioff zone implies that at least 150 km of the total shortening must have been accommodated as subduction, while the 100-330 km remainder may have been accommodated by thickening of the Pacific mantle lithosphere. Hence, in the southern South Island, shortening of the mantle may occur both at a slow convergence rate oriented perpendicular to the Alpine Fault, for example, 30 km of total convergence (Cande & Stock 2004), but also at a faster rate oriented perpendicular to the Australian slab. As a result, the thickened mantle lithosphere is an effective load that pulls down and thickens the overlying crust.

### 7 CONCLUSIONS

This earthquake refraction study offers new constraints on the uppermost-mantle properties beneath the Southern Alps, in a direction almost parallel to the Australian–Pacific Plate boundary and perpendicular to former crustal studies across the Southern Alps.

(i) The average Pn speed along the N60°E profile is 8.54  $\pm$  0.20 km s<sup>-1</sup> and the Moho is dipping at an apparent angle of 2.5  $\pm$  1.3° SW. We infer a 48  $\pm$  4 km crustal thickness near Wanaka. At 80 km east of the Alpine Fault but *ca*. 50 km east at Moho depth, the *Pn* anisotropy is in the range 7–13 per cent and the average *Pn* speed is 8.3  $\pm$  0.3 km s<sup>-1</sup>.

(ii) The Southern Alps crustal root near Wanaka is  $18 \pm 4$  km thick (relative to a coastal average of 30 km in South Island) and is twice that required by Airy isostasy for a crustal root of -300 kg m<sup>-3</sup> density contrast.

(iii) Mass balance predicts the presence of a mantle mass excess per unit strike length of  $2.7 \times 10^{11}$  kg m<sup>-1</sup> beneath the southern Southern Alps (Jackson Bay-Dunedin profile), for an 18 km thick crustal root of assumed -300 kg m<sup>-3</sup> density contrast with the lithospheric mantle. This mantle mass excess is approximately the same across the central (Stern *et al.* 2000, SIGHT T2) South Island, but would be greater for larger across-Moho density contrasts, for example, 40 per cent greater for a crustal root of -350 kg m<sup>-3</sup>.

(iv) The mantle body has a positive density contrast of 35  $\pm$  5 kg m<sup>-3</sup> minimum with 110  $\pm$  20 km width and 70  $\pm$  20 km thickness for an across-Moho density contrast of -300 kg m<sup>-3</sup>.

(v) For crustal roots with density contrasts of  $-400 \text{ kg m}^{-3}$  and more, the minimum density contrast required for the mantle body is larger than can be explained by the downwarp of isotherms alone and would require chemical heterogeneities within the mantle.

(vi) We speculate that the Puysegur margin, located southwest of the Southern Alps collision zone, contributes to thickening of the Pacific mantle lithosphere beneath the southern South Island by its subducted slab acting as a rigid backstop. This thickened mantle lithosphere is an effective load that pulls down the overlying crust.

(vii) The present gravity modelling is limited by the lack of constraints on the crustal structure beneath the southern South Island. Here crustal and mantle investigations are needed in order to model the gravity effect of the crustal root more precisely, and constrain the bulk and geometry of the mantle body.

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# Crustal Thickness and Pn Anisotropy Beneath the Southern Alps Oblique Collision, New Zealand

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Over-thickened crust and fast, anisotropic mantle material are interpreted beneath South Island, New Zealand, from an earthquake refraction study along the Southern Alps foothills. An  $8.54 \pm 0.20$  km/s Pn speed is estimated along the N60°E striking refraction profile and a maximum crustal thickness of  $48 \pm 4$  km is inferred near Wanaka township, at the southern end of the profile. The crustal thickness represents an 18 km thick crustal root relative to a 30 km coastal average. Thus, the root is 2-3times thicker than expected for Airy isostatic compensation of the mean ~1000 m Southern Alps topographic load. This suggests that the underlying mantle plays an active role in depressing topography. Comparison of the  $8.54 \pm 0.20$  km/s Pn-speed estimate with cross profiles suggests anisotropy arising from finite strain of the mantle lid rocks. The Pn anisotropy is estimated near Lake Tekapo, at the northern end of the profile, to be a minimum of  $6.5 \pm 3.5\%$ . We predict a maximum Pn anisotropy of 7-13% and an average isotropic Pn speed of ~8.3 km/s by adopting the fast polarization orientation from previous SKS splitting measurements done at the profile intersection. The Pn speed of 8.3 km/s is consistent with previous studies showing high average Pn speeds below the southern half of South Island and the presence of cold, dense mantle lithosphere.

# 1. INTRODUCTION

A long standing question on collision zones is how the upper mantle accommodates shortening. Two end-member models, intra-continental subduction [*Wellman*, 1979; *Beaumont*, 1996] and continuous thickening [*Molnar et al.*, 1999; *Stern et al.*, 2000] have been proposed for the Southern Alps of South Island, New Zealand. These two modes of deformation are analogous to the simple shear [*Wernicke*, 1985] and pure shear [e.g., *McKenzie*, 1978] models of extension, respectively. In the subduction-type end-member, deformation by simple shear is localized in a narrow and obliquely-

dipping shear zone at the slab top interface. In the continuous thickening end-member, shortening is accommodated by distributed pure shear. While both end-member models involve cold temperature contrasts in the mantle that produce faster wave speeds, the distribution of deformation and that of seismic anisotropy, i.e., localized on a narrow discontinuity vs. widespread, may help discriminate between them.

Continuous and distributed thickening has been suggested based on teleseismic traveltime residuals [Kohler and Eberhart-Phillips, 2002; Stern et al., 2000] showing a symmetric pattern. However, the simple shear model is preferred by a number of numerical models that intend to fit the GPS velocity field across the Southern Alps [Beavan et al., 1999; Ellis et al., 2006; Liu and Bird, 2006]. Numerical investigations of the development of continental collision [Pysklywec et al., 2002] have shown that both modes of shortening may be combined depending on the thermal structure and the convergence rate.

Determining Pn speeds and their azimuthal variation with respect to the orientation of the plate boundary can provide insight on physical conditions and deformation experienced in the uppermost mantle. Pn anisotropy combined with SKS-splitting measurements [*Savage*, 1999;





**Figure 2.** Top: Arrivals from the Fiordland aftershock are bandpass filtered at cut-off and corner frequencies of 0.5-1-5-10 Hz. First-break Pn are indicated by the bottom pair of arrows and single arrows in blow-up on the right, and predicted Pg and Pn travel-time curves by dashed curves (see model Figure 3c). Pn arrivals are followed ca. 1.5 s later by arrivals (~1.5-s peg-leg indicated with top pair of arrows) with much larger amplitude. These second arrivals have the same apparent wave speed as the Pn and are interpreted as an internal reflection ~5 km near the source. The Pn-speed estimate and corresponding 95% confidence interval (right-hand side of graph) is the mean of single regression slopes weighted with their respective standard deviations. Bottom:Arrivals from the ML 4.1 offshore Cheviot event are bandpass filtered at cut-off and corner frequencies of 0.5-1-3-5 Hz. Note the offset axis is in the opposite direction to that of the top figure. First-break Pn are indicated by the pair of arrows and the predicted Pn travel-time curve by a dashed curve (see model Figure 3c). The Pn-speed estimate is the result of a single linear regression and is given with corresponding 95% confidence interval (right-hand side of graph). In both graphs the trace of the third station from the left was shifted by 3.5 s to correct a timing error. However, the pick wasn't included in Pn-speed calculations because of uncertainty in the timing error [after *Bourguignon et al.*, 2007].

**Table 1.** Events 1–4 and 6 are the Fiordland aftershocks located using data collected by a temporary seismograph deployment [*Reyners* et al., 2003] and a 1-D velocity model for Fiordland [*Eberhart-Phillips and Reyners*, 2001]. Event 5 is the offshore Cheviot event located by GeoNet using the standard 1-D model for New Zealand [*Maunder*, 1999].

Event		Origi	n time		Magnitude		
nb	ID	Date	Time	Lat.	Long.	Depth (km)	Magintude
1	2105255	2003/08/25	03:36:30.26	-45.111	166.964	20.5	5.0
2	2106280	2003/08/26	23:56:26.57	-45.486	166.596	19.7	5.5
3	2106314	2003/08/27	01:29:40.41	-45.442	166.716	22.7	5.1
4	2106319	2003/08/27	01:42:54.12	-45.314	166.945	24.4	5.6
5	2106361	2003/08/27	03:39:37.23	-42.802	173.758	35	4.1
6	2110611	2003/09/04	08:40:44.25	-45.224	166.921	22.7	6.1

of this experiment with others of this type in the Sierra Nevada [*Jones et al.*, 1994; *Savage et al.*, 1994; *Ruppert*, 1998; *Louie et al.*, 2004] suggests that two factors may have made the observation of Pn along the crustal root possible. (1) The Moho is smoothly dipping along the profile line, as a result of both the obliquity of the profile relative to the Southern Alps crustal root and gentle thickening of the root in the east. (2) Fast wave speeds within the cold mantle lithosphere may contribute to efficient refraction of seismic waves along the Moho boundary.

### 3. CRUSTAL THICKNESS

Previous crustal studies in South Island [Davey and Broadbent, 1980; Reyners et al., 1993; Eberhart-Phillips and Reyners, 2001; Eberhart-Phillips and Bannister, 2002; Scherwath et al., 2003; van Avendonk et al., 2003] were integrated into a crustal model taken along our refraction profile (Figure 3c). In addition, the above measurements were used to constrain the dip of the Moho to 2.5°SW, to the southwest of the intersection of the Fiordland-Cheviot profile with SIGHT T2. The uppermost mantle wave speed was set to 8.54 km/s. Ray tracing [Luetgert, 1992] on this 2-D crustal model (Figure 3c) shows that rays propagating from Fiordland and offshore Cheviot constrain a ~150 km Moho portion extending from Wanaka (southern SI) to the intersection of SIGHT T2 with the Fiordland-Cheviot profile, i.e., west of Tekapo (central SI). Along this profile, the Moho depth is maximum near Wanaka, located at the southwestern tip of the constrained zone, and estimated to  $48 \pm 4$  km (Figure 3c). This crustal thickness represents an  $18 \pm 4$  km thick crustal root beneath the Southern Alps relative to a coastal average of 30 km crustal thickness in South Island [Godfrey et al., 2001; Melhuish et al., 2005]. Hence, near Wanaka the crustal root is estimated to be 4 km thicker than near Mount Cook [Scherwath et al., 2003] and suggests thickening of the Southern Alps crustal root from the NE to the SW. In contrast, mean elevations decrease from ~1500 m near Mt Cook to ~1000 m near Wanaka (Figure 3b). Thus, topography and crustal root thickness have an inverse relationship in the Southern Alps region. For 1000 m elevation and an Airy root with a density contrast of -300 kg/m<sup>3</sup> [Bourguignon et al., 2007] or -400 kg/m<sup>3</sup> [Scherwath, 2002] with the surrounding mantle, the crustal root should be 9 or 6 km thick, respectively, i.e., less than half the ca. 18 km inferred near Wanaka. A similar conclusion was drawn further north where Stern et al. [2000] deduced the presence of a dense mantle body beneath the central Southern Alps. Modeling of the crustal structure along SIGHT T2 with an assumed  $-450 \text{ kg/m}^3$  density contrast between crustal root and mantle predicts a Bouguer gravity anomaly far more negative than observed [Reilly and Whiteford, 1979], indicating a ca. 10 km excess of crustal root thickness (relative to that expected from Airy isostasy). Stern et al. [2000] attributed this excess of crustal thickness to the downward pull of a cold, and therefore dense, lithospheric root (Figure 3d).

Intermediate depth seismicity beneath Fiordland indicates steepening of the Australian slab in proximity to the Southern Alps collision zone [*Reyners et al.*, 2002, also see Figure 3d]. It has been suggested that the Australian slab may act as a backstop that converges at ~26 mm/yr, i.e., 3/4 of the full plate speed, with the Pacific lithosphere and contributes to thicken the Pacific mantle lithosphere from the southwest (Figure 3d; *Malservisi et al.*, 2003; *Bourguignon et al.*, 2007].

### 4. PN ANISOTROPY

We combine the Pn speed from this study's earthquake refraction and that from SIGHT T2's seismic line [*Scherwath et al.*, 2003] to infer the Pn anisotropy at these two profile intersections. The first terms of the Taylor expansion of the azimuth-dependent wave speed,  $\alpha(\varphi) = om$ ,  $\alpha_0 + Ccos(2\varphi) +$  $Dsin(2\varphi)$  [*Smith and Ekström*, 1999] are employed to rotate

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**Figure 3.** (*a*): Bouguer gravity anomaly [*Reilly and Whiteford*, 1979] in 50 mGal contours is supplemented with locations of Mt Cook (triangle) and of the main divide (dashed line) and Fiordland-Cheviot (solid line) profiles of the graph below. (*b*): Mean topography in a 10 km wide swath along the Fiordland-Cheviot profile (thin curve) and the Main Divide (thick dashed curve) and Bouguer anomaly (thick curve; [*Reilly and Whiteford*, 1979]) along the Fiordland-Cheviot profile (solid line in Figure 3a). (*c*): 2-D velocity model based on this study's Pn speed and Moho dip estimates and on results from: *Davey and Broadbent* [1980], *Reyners et al.* [1993], *Eberhart-Phillips and Reyners* [2001], *Eberhart-Phillips and Bannister* [2002], *Scherwath et al.* [2003] and *van Avendonk et al.* [2003]. Triangles denote deployed instruments, T1/T2 indicate intersections with SIGHT previous crustal studies, ray-tracing is for events 5 and 6 (predicted travel-time curves in Figure 2), thick dashed line indicates the constrained portion of the Moho that extends from Wanaka to Tekapo, question marks denote unconstrained interfaces [after *Bourguignon et al.*, 2007]. (*d*): 2-D model of the Fiordland-Cheviot profile (no vertical exaggeration) with seismicity within a 10 km wide swath and interpretation.

the intersecting profiles into fast and slow orientations and infer a maximum anisotropy. Three equations are found, which enable us to solve for the three unknown parameters,  $\alpha_{o}$ , the average Pn-speed and both constants C and D, simultaneously. Two equations are determined by substituting  $\alpha(\phi)$  and  $\phi$  for the known Pn-speed values and azimuths of the two respective intersecting profiles. A third equation,

 $\frac{d\alpha(\phi)}{d\phi} |_{\phi=\Phi} = 0], \text{ is found by assuming } \alpha(\phi) \text{ is maximum for the fast propagation azimuth, } \Phi, \text{ from a nearby SKS-splitting}$ 

measurement.

Taking our result of  $8.54 \pm 0.20$  km/s and a Pn speed of  $8.0 \pm 0.2$  km/s on the nearly perpendicular profile SIGHT T2

[Scherwath et al., 2003], implies  $6.5 \pm 3.5\%$  apparent anisotropy (T in Figure 1). Assuming the fast orientation to be that of the nearby SKS fast polarization orientation, we can calculate the maximum anisotropy. However, the nearby fastpolarization measurement [*Klosko et al.*, 1999] is located at the transition between two domains of anisotropy and this must be an average resulting from the overlap of Fresnel zones over the two domains. In central South Island, SKS fast polarization orientations,  $\Phi$ , are sub-parallel to the Alpine Fault, while in southern South Island these are consistently oblique to the Alpine fault, i.e., the orientation of shear (Figure 1) [*Klosko et al.*, 1999; *Molnar et al.*, 1999]. Therefore, two possible fast orientations need to be considered.

Taking central South Island  $\Phi$  of  $56 \pm 2^{\circ}$  implies a maximum Pn anisotropy  $\delta P$  of  $7 \pm 3.5$  % and an average Pn speed  $\alpha_{\circ}$  of  $8.25 \pm 0.24$  km/s while taking southern South Island mean  $\Phi$  of  $21 \pm 1^{\circ}$  implies a  $\delta P$  of  $13.3 \pm 3.5$ % and an  $\alpha_{\circ}$  of  $8.42 \pm$ 0.28 km/s. A mean  $\alpha_{\circ}$  of  $8.3 \pm 0.3$  km/s fits both results and is consistent with Haines' average  $8.3 \pm 0.1$  km/s for southern South Island (1979) and wave-speed perturbations of  $\sim 2$  % [relative to the IASP91 Earth's model; *Kennett and Engdahl*, 1991] imaged by inversion of teleseismic travel times [*Kohler and Eberhart-Phillips*, 2002]. The 8.3 km/s Pn speed is  $\sim 3\%$  more than the 8.1 km/s world-wide average [*Kennett and Engdahl*, 1991] and suggests cold and dense upper mantle material.

We replicate the analysis further north at the intersection with SIGHT T1 and make the important assumption that the Pn speed northeast of the constrained portion of the Cheviot-Fiordland refraction profile is  $8.54 \pm 0.20$  km/s as well. The Pn speed on the intersecting SIGHT T1 line is 7.9–8.0 km/s [*van Avendonk et al.*, 2003], i.e., slightly less than along SIGHT T2. Here the azimuths of SIGHT T1 and the Cheviot-Fiordland profile almost align with the slow and fast orientations of wave propagation. Taking a 7.9 ± 0.2 km/s Pn speed along SIGHT T1, an 8.54 ± 0.20 km/s along our refraction profile and a  $\Phi$  of 56 ± 2°, results in a 8.1 ± 3.5 % Pn anisotropy and an average Pn speed  $\alpha_0$  of 8.21 ± 0.24 km/s.

If the 7% or 13% anisotropy at the intersection with SIGHT T2 is constant throughout the mantle lid, then an anisotropic layer of about 100 km or 50 km thickness, respectively, would account for the observed SKS-delay time of 1.76 s [*Klosko et al.*, 1999] (assuming a P-to S-anisotropy ratio of 1.4 and a 4.7 km/s average S-wave speed in the uppermost mantle).

Dynamic slip alone can't explain in situ anisotropy greater than a theoretical maximum of 10% [Ribe, 1992] as calculated for southern South Island, but requires additional dynamic re-crystallization by subgrain rotation and grainboundary migration [Nicolas et al., 1973; Karato, 1988], additional pure shear or infinite strain. All these processes have the effect of rotating fast propagation orientations parallel to the shear orientation, i.e., reducing the obliquity of fast orientations to that of shear. However, SKS fast polarizations of southern South Island are ~28° oblique to the Alpine Fault and the shear orientation. Hence, a  $13.3 \pm 3.5\%$  Pn anisotropy as calculated for southern South Island  $\Phi$  seems incompatible with the obliquity of fast polarization orientations from SKS splitting to that of shear. A  $7 \pm 3.5\%$  anisotropy, calculated for central South Island, is a more reasonable result. However, there fast polarizations are oriented parallel to the shear orientation. An amount of anisotropy intermediate to 7% and 13% or rotation of material independent of strain would resolve this paradox.

# 4.1. Comparison with previous Pn-anisotropy measurements in South Island

Three other Pn-anisotropy measurements were made on crossing refraction lines. The Pn anisotropy is  $11.5 \pm 2.0\%$ on the Australian side, 30 km west of the surface trace of the Alpine Fault (S in Figure 1) [Scherwath et al., 2002]. If we take the dip of the Alpine Fault as 40°SE [Kleffmann et al., 1998], then at the Moho, the measurement on the Australian side is at a similar distance to the fault as our measurement of 7-13% anisotropy on the Pacific side. Offshore 230 km east of the Alpine Fault, two null Pn anisotropy measurements on crossing lines SIGHT T1 and T3 and SIGHT T2 and T3 (B1 and B2 in Figure 1) show that upper mantle anisotropy does not extend 50 km east of South Island [Baldock and Stern, 2005; correction in prep.]. The Pn speed is  $8.1 \pm$ 0.1 km/s in both transect azimuths and can be assumed as the isotropic Pn speed. Beneath the Canterbury plains (east of the Southern Alps) into the offshore, however, northwestsoutheast raypaths define a broad region of  $7.8 \pm 0.1$  km/s Pn speed [Baldock, 2004]. Assuming 7.8 km/s and 8.1 km/s are the minimum and isotropic Pn speeds, respectively, the Pn anisotropy beneath the Canterbury plains is  $7.5 \pm 3.0\%$ [Baldock and Stern, 2005; correction in prep.].

These measurements suggest that the Pn anisotropy is strong up to ~70-80 km distance from the Alpine Fault at depth. Scherwath et al. [2002] noted that the  $11.5 \pm 2.0\%$  Pn anisotropy (S in Figure 1) is slightly greater than the theoretical maximum of ~10% for strain-induced anisotropy [Ribe, 1992]. They suggested dynamic recrystallization, some pure shear component and/or infinite strain as possible mechanisms to explain the high observed anisotropy. In the east of South Island, the Pn anisotropy is less strong, and possibly extends as far as the east coast, ~150 km east from the Alpine Fault at depth. Modeling of Ellis et al. [2006], in contrast, predicts a width of ~100 km maximum. Duclos et al. [2005] interpret constant SKS-splitting delay times but decreasing Pn anisotropy with distance to the Alpine Fault as the result of widening of a continuously distributed zone of mantle deformation with depth.

### 5. CONCLUSIONS

A Pn speed of  $8.54 \pm 0.20$  km/s and a Moho apparent dip of  $2.5 \pm 1.3^{\circ}$ SW are determined from an earthquake refraction travel-time analysis along the Southern Alps crustal root. The profile line is oriented N60°E, ~N5°E from the Alpine Fault.

We estimate a  $48 \pm 4$  km crustal thickness near Wanaka, which is a ca. 18 km thick crustal root (relative to a coastal average of 30 km). Here the root is at least twice as thick

as expected for Airy isostatic compensation of the Southern Alps topographic load.

The relatively high wave speed of  $8.54 \pm 0.20$  km/s is interpreted to be the result of both anisotropy in the mantle lid and a relatively high average Pn speed of ca. 8.3 km/s below southern South Island. The Pn anisotropy is 7–13% 80 km east of the Alpine Fault. Pn anisotropy values across South Island confine the deformation to a maximum ~100 km thick layer in the mantle lid assuming anisotropy is constant with depth. An average 8.3 km/s P-wave speed is interpreted as mantle lithosphere colder and, hence, denser and with higher P-wave speeds than surrounding mantle rocks. This colder zone results from the downward deflection of isotherms and acts as an effective load at the base of the crust.

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