

Pn anisotropy and distributed upper mantle deformation associated with a continental transform fault

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[1] Differences in Pn speeds within continental lithosphere off the west coast of South Island, New Zealand, require anisotropy of at least $10 \pm 3\%$. These data concur with SKS splitting and are inferred to show high strain of mantle lithosphere over a zone >200 km wide. An active source seismic experiment yields Pn speeds of 8.6 and 7.7 km/s along nearly perpendicular lines. The higher speed is sub-parallel to the polarization of the faster quasi-S wave measured from SKS splitting. Most of the observed ~ 2 s of splitting in South Island can be accommodated in mantle lithosphere 150 km thick, if we assume a P to S anisotropy ratio of 1.4. The large magnitude of Pn anisotropy shows that anisotropy must occur in the relatively cold uppermost part of the mantle lithosphere and raises the possibility that dynamic recrystallization occurs in cold (~ 500 – 600°C) olivine at geologic strain rates. **INDEX TERMS:** 8120 Tectonophysics: Dynamics of lithosphere and mantle—general; 7218 Seismology: Lithosphere and upper mantle; 5199 Physical Properties of Rocks: General or miscellaneous

1. Introduction

[2] Measurements of seismic anisotropy over the past 10–15 years have made it possible to carry out quantitative structural geology of the upper mantle by determining the orientation of strain and by placing bounds on strain magnitudes within it [e.g. Silver, 1996; Savage, 1999]. Yet, despite this progress two aspects of that measured anisotropy limit its applicability: accurate knowledge of the depth range of seismic anisotropy and the cause of crystal alignment. First, constraining the depth range over which the mantle is anisotropic is much more difficult than merely demonstrating the existence of anisotropy; the splitting of SKS phases can be due to anisotropy in either the asthenosphere or the overlying mantle lithosphere, if not both. Second, two processes can cause alignment of crystals (lattice preferred orientation). Strain and rotation of individual crystals will cause anisotropy such that the faster quasi-S waves are polarized parallel to the axis of maximum elongation and the slowest quasi-S waves parallel to the axis of maximum flattening [e.g. Ribe, 1992]. In addition, during dynamic recrystallization, new crystals tend to grow parallel to the orientation of shear, at least in simple shear [Zhang and Karato, 1995]. Because of its temperature dependence, dynamic recrystallization may not occur in the shallower parts of the mantle.

[3] To examine the depth dependence of anisotropy, and therefore also its temperature dependence, an independent measure-

ment of anisotropy that samples the mantle differently from shear waves propagating vertically through it is needed. Our observations of P-waves refracted in the upper mantle (Pn), adjacent to an active continental transform fault, allow such a measurement. Moreover, Pn speeds measured using explosive or air-gun sources provide higher resolution than those made using earthquakes [Smith and Ekström, 1999].

[4] High capacity air-gun sources of lines 2 and 3W (Figure 1) of the South Island Geophysical Transect (SIGHT) program [Okaya *et al.*, 2001] provided about 4000 shots into 50 seismographs for this study. Some of the highest values of SKS splitting have been recorded within South Island [Klosko *et al.*, 1999], and by measuring Pn anisotropy in the same region, we can provide a constraint for the depth range of anisotropy. Moreover, being able to determine the magnitude of anisotropy is important, because laboratory measurements on olivine and enstatite show that there are upper limits to the magnitude of anisotropy induced by kinematic rotation and alignment of olivine crystals [Ribe, 1992]; to produce anisotropy $> \sim 10\%$ some component of dynamic recrystallization is required [Karato, 1988].

[5] Tectonically, New Zealand's South Island has undergone dextral shear since ~ 45 Ma, with displacement of 850 km (± 100 km) [e.g. Sutherland, 1999]. Its manifestation in crustal rock includes ~ 460 km of slip on the Alpine fault, and more diffuse shearing northwest and southeast of the fault. This shearing is reflected by the bending of belts of rock ("terrane") [Sutherland, 1999], and internal rotations shown by paleomagnetic declination anomalies [e.g. Mumme and Walcott, 1985]. The consistent orientations and distribution of the shear-wave splitting measured by Klosko *et al.* [1999] suggests that the mantle beneath South Island, and the western part of North Island, has been sheared over a zone 200–400 km wide [e.g., Molnar *et al.*, 1999; Moore *et al.*, 2001].

2. Pn Traveltimes and Speeds

[6] Pn is observed at the intersection of lines 2 and 3W (Figure 1) from offshore air-gun shots recorded on land. We analyzed these recordings as common receiver-gathers (i.e. traces from multiple airgun shots recorded at a single onshore seismograph). The close spacing of air-gun shots (50 m) enhances phase correlations for accurate determinations of P-wave speed [Okaya *et al.*, 2001]. On line 3W, receiver-gathers cover the NE and SW directions to determine apparent Pn speeds in both directions and therefore allow inferences of P-wave speed in the mantle and the dip of the Moho. On line 2, however, receiver gathers sample Pn from the same direction (NW). To sample the opposite direction, we re-ordered the traces to create common shot-gathers (i.e. a single airgun pop recorded on multiple seismographs) to estimate apparent speeds in the reverse direction.

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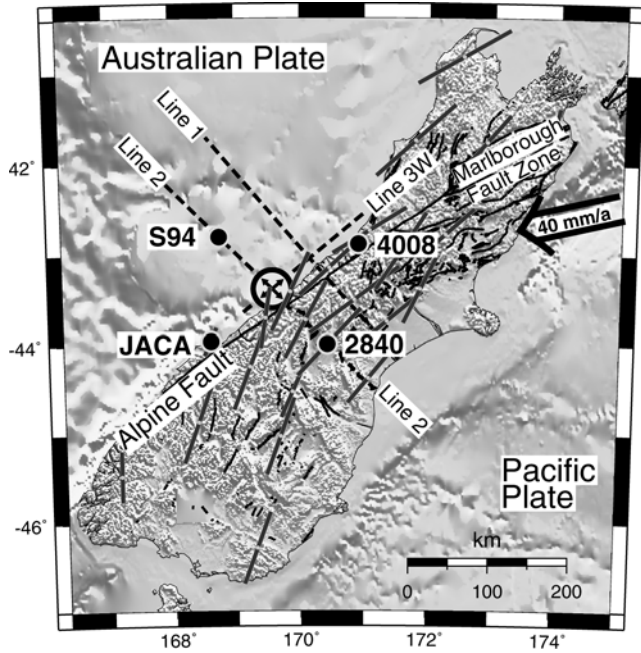


Figure 1. Map of South Island, New Zealand, with locations of three wide-angle seismic lines (dashed) and four stations (S94, 2840, 4008, and JACA) from which mantle refractions are shown in Figure 2. Pn anisotropy occurs at the intersection of the lines (cross). The size of the cross is proportional to the 50-km-wide area of the Moho from which Pn phases used here emerged (see also Figure 3). The grey bars indicate SKS fast polarization orientations [Klosko *et al.*, 1999], the bar length is proportional to the amount of shear-wave splitting.

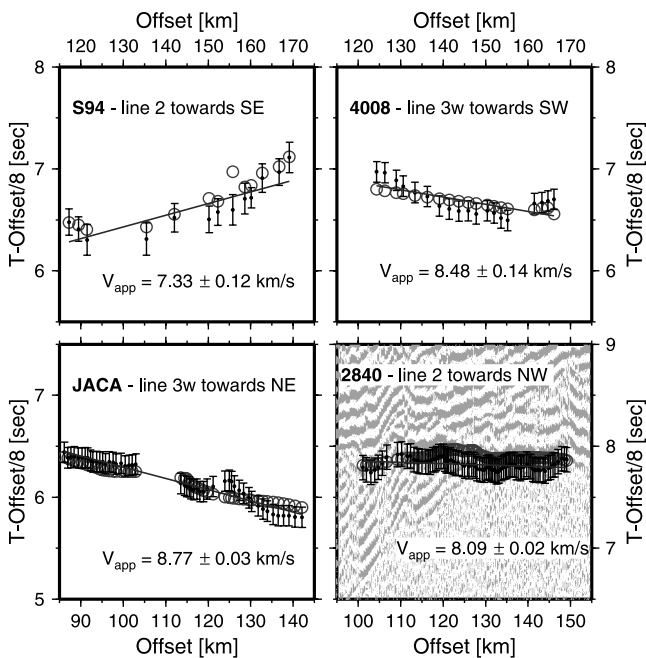


Figure 2. Pn arrival times for selected stations, plotted at reduced traveltime. Stations are ordered as they appear on the map in Figure 1. Least-squares best-fit straight lines through the arrivals yield apparent speed. Also shown are predicted arrivals (open circles) from the structures shown in Figure 3. Data quality is shown as an example for station 2840.

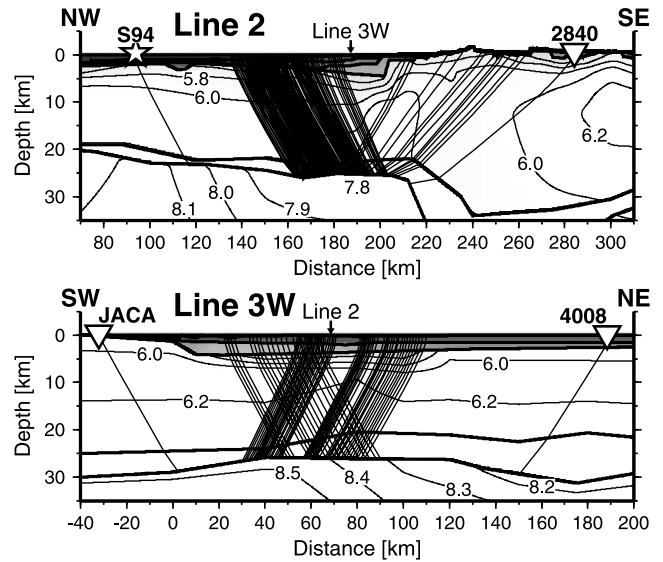


Figure 3. Ray-tracing diagrams for four stations with 50-km-wide Moho refractor surrounding the intersection of both profiles. Structures shown here are from inversions of seismic wide-angle data. Speed contours are in km/s.

[7] To isolate Pn arrivals that cover only the region where the two lines intersect, we analyzed Pn arrivals emerging from the Moho within an area 50 km wide. We disregarded gathers whose Moho penetrations span less than ~ 15 km of the region or that consist of fewer than 5 arrival times. This selection resulted in two gathers, 4008 and JACA, for line 3W, and 24 shot and 24 receiver-gathers for line 2.

[8] On line 3W, apparent Pn speeds from both 4008 and JACA (Figure 2) give $V_p = 8.62 \pm 0.14$ km/s. For line 2, the two gathers that cover the zone of interest best, S94 and 2840, yield $V_p = 7.69 \pm 0.13$ km/s. With $V_{max} = 8.62 \pm 0.14$ km/s, and $V_{min} = 7.69 \pm 0.13$ km/s, the anisotropy is $11.5 \pm 2.4\%$. Taking all 48 gathers of line 2 that fulfill above requirements reduces the possible bias from choosing only two representative gathers for line 2. The results are similar, if with slightly larger error: $V_{up} = 8.00 \pm 0.08$ km/s, $V_{down} = 7.39 \pm 0.12$ km/s, and $V_p = 7.68 \pm 0.15$ km/s.

[9] Pn speeds determined previously from a combination of forward modeling and inversion of all data for line 2 [Schervath *et al.*, 1998] and 3W (unpublished work by A. Melhuish and others, 1998) were 7.80 and 8.45 km/s (± 0.25 km/s), respectively (Figure 3). This corresponds to P-wave anisotropy of $8 \pm 4\%$. However, we regard this result as inferior to the direct and localized analysis of Pn-slopes as both inferred structures contain a degree of smoothing, which is common in tomographic inversion [e.g. Zelt and Barton, 1998]. Extreme values of speeds that may be real will be “smeared,” along ray paths into adjacent areas with speeds closer to the average values of the respective layers. Therefore, analyzing apparent speeds by linear fits to raw measurements of Pn arrival times should yield less biased determinations of Pn speeds. On the other hand, inferring Pn speeds by fitting straight-line segments to arrival times assumes a homogeneous overlying crust and planar rather than curved interfaces.

[10] In the inversions shown in Figure 3, neither is the speed distribution above the Moho, particularly on line 2, homogeneous, nor is the Moho itself planar. To estimate the error introduced by assuming homogeneous layers and a planar Moho, we ran tests using the structures in Figure 3 as a guide. We traced rays through the structures for lines 3W and 2 and then analyzed these synthetic Pn arrivals as above, by fitting straight-line segments by least squares. The Pn speed recovered for line 3W matched that used to generate the synthetic Pn arrival times, but for line 2 we obtained a

Pn speed of 7.6 km/s, 0.2 km/s less than the value sampled by the synthetic Pn phases. Thus, plausible lateral variations in crustal structure can be mapped into Pn traveltimes to yield erroneous speeds. Accordingly, the apparent error in assuming a planar Moho and uniform crustal structures is 0.2 km/s, and this would increase the total uncertainty in estimating anisotropy to at least 4%. The difference of 0.2 km/s could be due to systematic non-uniformity in the overlying crust, or to random errors linked to imperfect knowledge of the crustal structures and artifacts introduced in the inversion process. We choose to ascribe some uncertainty in this correction to random effects as discussed above. In effect, we adopt an increased estimate for Pn speeds of line 2 by 0.1 km/s and assigning the other 0.1 km/s as error due to uncertainty in crustal structure. Hence, although we directly measure a magnitude of anisotropy of $11.5 \pm 2.4\%$ from the plot of traveltimes vs. distance, we use $10 \pm 3\%$ for the difference in Pn speeds along the two lines.

3. Discussion

[11] An estimate of Pn anisotropy from only two lines has to be regarded as a minimum unless the orientations of the lines are fortuitously parallel to those with the maximum and minimum speeds. Let us assume that the source of anisotropy is the same for Pn and SKS and that anisotropy does not change with depth. The closest SKS result (Figure 1) was obtained within 25 km of our measurement, and its fast SKS orientation is $\sim 30 \pm 10^\circ$ counter-clockwise from the Alpine fault. The orientation of our faster Pn is approximately 10° counter-clockwise from the Alpine fault, a difference of 20° . Assuming that azimuthal variations in speed match the orientation to the strain ellipse [Ribe, 1992], $V(\vartheta) = \bar{V} + 0.5\Delta V \cos(2\vartheta)$, where ϑ is measured with respect to the orientation of the fastest P or quasi-S wave, \bar{V} is the average speed, and ΔV the difference between minimum and maximum speeds, we could correct our "apparent" Pn-anisotropy of $10 \pm 3\%$ to become $13.4 \pm 5\%$. Instead, for an orientation of the fast direction parallel to the Alpine fault, as suggested by SKS stations in the north of South Island and east of the Alpine fault (Figure 1), the same logic yields a Pn-anisotropy of $11 \pm 5\%$.

[12] If we relax the assumption that the average orientation of lattice preferred orientation within the layer traversed by SKS is the same as that sampled by Pn, the speeds measured along the two lines might represent the maximum and minimum P-wave speeds at the top of the mantle by considering finite shear. If the lattice preferred orientation results from finite strain due to displacement of amount d over a width W , the orientation of the maximum elongation, ϕ , relative to the orientation of the zone of simple shear is given simply by $\tan(2\phi) = 2W/d$. For $d = 850$ km and $W = 400$ km, ϕ should be 22° counter-clockwise from the strike of the Alpine fault, and therefore 12° from our faster Pn. For a maximum Pn speed oriented parallel to the maximum elongation, the Pn-anisotropy implied by our measurements should be $11 \pm 3\%$. Therefore, despite the estimated $\sim 10\%$ difference in Pn speeds, consideration of likely fast orientations, either from fast SKS orientations or from an argument based on finite strain, suggests anisotropy to be in the 11–13% range.

[13] From a compilation of measured P- and S-wave speeds of mantle xenoliths, Ismail and Mainprice [1998] reported ratios of P- and S-wave anisotropy around 1.4, though Ribe [1992] calculated ratios of 1.5–1.7 for different amounts and types of strain. Assuming a ratio of 1.4, a Pn anisotropy of $10 \pm 3\%$ corresponds to S-wave anisotropy of $7 \pm 3\%$, and that of $13.4 \pm 5\%$ for P corresponds to $9.5 \pm 5\%$ for S. Using an interval between fast and slow quasi-S waves of 1.8 s [Klosko et al., 1999] and an average mantle S-wave speed in the upper mantle of 4.7 km/s, S-wave anisotropy of $7 \pm 3\%$ implies a thickness of the anisotropic layer of 120 km (thicknesses between 85 and 212 km, given the uncertainty) and $9.5 \pm 5\%$ implies 89 km (between 58 and 188 km).

These thicknesses show that most, if not all, SKS splitting could occur in mantle lithosphere.

[14] Pn speeds less than 8 km/s are distributed over a region at least 100 km west of the Alpine fault (Figure 3). The eastward extent of low Pn speeds is more difficult to establish as tight control on Pn is not acquired until at least 100 km east of the Alpine fault [Schervath et al., 1998]. On line 1, parallel but ~ 60 km northeast of line 2 (Figure 1), low Pn speeds are also found beneath and west of the surface trace of Alpine fault [Van Avendonk et al., 1999]. The Pn anisotropy reported here implies that the distributed strain extends offshore at least 100 km west of the Alpine Fault.

[15] Ribe's [1992] numerical calculations for a medium consisting of 70% olivine and 30% enstatite, show that extensional strains of about 1.6 in a zone of simple shear (compared with values of 2.5 to 4.6 for New Zealand given above) will produce P-wave anisotropy of 8%. At this level of strain his plots show that anisotropy continues to increase, albeit at a decreasing rate as strain increases; a limiting value of about 10% is indicated. If so, deformation and rotation of olivine crystals with some dynamic recrystallization may be required to account for the magnitude of the P-wave anisotropy reported here, whether it be 10% or 13%. Dynamic recrystallization not only permits higher levels of P-wave anisotropy, perhaps as high as 20% [Ismail and Mainprice, 1998], but also facilitates rotation of the fast orientation to become parallel to the plane of simple shear [Zhang and Karato, 1995]. Zhang and Karato [1995] showed that at a strain rate of 10^{-5} s^{-1} , dynamic recrystallization affected anisotropy of samples as hot as 1300°C , but significantly less those at 1200°C . As the average strain rates for New Zealand ($\sim 2-3 \times 10^{-15} \text{ s}^{-1}$) have been nearly 10 orders of magnitude slower, the large magnitude of P-wave anisotropy reported here raises the possibility that the temperature dependence of dynamic recrystallization permits it to occur throughout the entire mantle lithosphere at geologic strain rates, at temperatures as low as $500-600^\circ\text{C}$.

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