Relationship between crustal finite strain and seismic anisotropy in the mantle, Pacific–Australia plate boundary zone, South Island, New Zealand

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SUMMARY
The relationship between crust and mantle deformation in plate boundary zones is an outstanding problem in geodynamics. New Zealand provides a rare opportunity to examine the way strike-slip faults relate to deep-seated zones of lower crustal and mantle flow. A conspicuous bend deflects elongated terranes such as the Dun Mountain ophiolite through >70° in the continental crust, which we interpret as being the result of distributed dextral shear between the Pacific and Australian plates in the Cenozoic. We utilized variations in the strike of two different geological markers (ophiolite terrane and fold belts) towards the Alpine fault to calculate finite strains in three crustal domains. The deflection is best matched by a transpressional, rather than a simple shear deformation. These transpressional models predict maximum horizontal finite strain azimuths (±10°) that trend anticlockwise ∼30° to ∼10° from the Alpine fault. These azimuths match published fast polarization azimuths of SKS and local (<100 km deep) shear waves. This coincidence indicates that lithospheric mantle and upper crustal deformation are broadly coupled, although the former is more widely distributed. As shear wave splitting results from mineral fabrics that are the result of finite strain, they are appropriately compared with the finite strain recorded by the deformed crustal markers. Using the geodetic data to characterize the infinitesimal displacement field, and the deformed pattern of markers to characterize the finite displacement field, one concludes that deformation in New Zealand was not steady state. The observed seismic anisotropy was probably the product of ∼45 Myr of PAC–AUS motion, not just 5 Myr as some have suggested. Our results contradict those of recent experimental studies of olivine deformation at high temperature, by suggesting that seismic anisotropy fabrics in naturally deformed mantle rocks can track finite strain at shear strains of >2.1, and strains (εs) > 1.9 without being reoriented towards the shear plane by recrystallization. In agreement with our modelling of mantle flow, the large SKS shear wave splitting delays (>2 s) on the South Island suggest that the direction of maximum finite stretch in the mantle is more likely horizontal than vertical. This inference is consistent with the 3-D strain calculated from deflected markers in the crust.

Key words: continental deformation, mantle, New Zealand, shear wave splitting, tectonics.

INTRODUCTION
The distribution and nature of upper-mantle deformation in plate boundary zones is an outstanding problem in modern geodynamics (e.g. Davis et al. 1997; Vinnick et al. 1992; Hirn et al. 1995; Silver 1996; Stern et al. 2000). Instantaneous velocities at the surface in a deforming zone are poor indicators of the width or pattern of flow at depth (e.g. Savage et al. 1999), and finite deformation of the mantle is difficult or impossible to measure directly. Controversy surrounds the relationship between crust and mantle deformation fields, the way in which strike- or oblique-slip faults at the surface may relate to deep-seated zones of lower crustal and mantle flow, and the width of those zones (e.g. Molnar 1992; Bourne et al. 1998; Teyssier & Tikoff 1998; Holt 2000). Many numerical models of collision zones assume, as a kinematic boundary condition, that the upper mantle is detached from the crustal part of the continental lithosphere and subducted, with the polarity of this subduction imparting an asymmetry to collisional mountain belt development (e.g. Willet et al. 1993; Beaumont et al. 1996; Braun & Beaumont 1995).

In recent years, geophysical data from the Southern Alps in New Zealand (Figs 1 and 2) have played a key role in the controversy concerning the relative behaviours of crust and mantle in convergent
Figure 1. 45 Ma plate reconstruction for the Pacific and Australian plates in New Zealand. In a rigid reconstruction, the curved shape of the Maitai Terrane yields a large inherited sinistral offset together with crustal-scale dextral ‘drag’. This incongruity is avoided if the marker was approximately straight in the early Cenozoic. From Sutherland (1999). The inset shows the present-day geometry of the Dun mountain ophiolite/Maitai Terrane through New Zealand.

Orogenic belts. In this paper, we will evaluate seismic anisotropy data from New Zealand’s South Island from the perspective of the structural geologist, placing those data in the useful context of plate motions and finite strain in the mantle. Beaumont et al. (1996) and Batt & Braun (1997) modelled intracontinental subduction of the Pacific Plate mantle lithosphere westward beneath the Australian plate—a concept first suggested by Wellman (1979). In conflict with this idea, Stern et al. (2000) have argued, on the basis of the recorded pattern of teleseismic P waves, that ~100 km of bulk shortening of the lithospheric mantle has taken place by uniform thickening across an 80–100 km wide zone beneath the Southern Alps following the onset of convergence ~6.5 Ma. Further NE in the Marlborough fault system, Bourne et al. (1998) interpreted geodetic and geological slip rate data as supporting a model in which upper crustal strike-slip faults are underlain by a >100 km-wide, dextral simple shear zone in the lower crust and mantle, which has an approximately uniform strain rate.

Seismic anisotropy studies are increasingly being used to document the orientation and intensity of deformation fabrics in the mantle (e.g. Savage 1999). Observations of shear wave splitting are attributed to the lattice-preferred orientation (LPO) of mantle minerals (chiefly olivine), resulting from crystal-plastic deformation and dynamic recrystallization of those intrinsically anisotropic crystals, with the fast polarization azimuths either being subparallel to the maximum principal stretch of the finite strain (e.g. Nicolas & Christensen 1987; Ribe & Tu 1990; Wenk et al. 1991; Ribe 1992; Ismail & Mainprice 1998; Tommasi et al. 1999) or the shear direction (e.g. Zhang & Karato 1995; Bystricky et al. 2000). Recently, shear wave splitting data have been obtained across the South Island of New Zealand, including SKS (core–mantle boundary) (Klosko et al. 1999) and S (local earthquake) (Audoine et al. 2000) phases. The fast polarization azimuths of these data typically make a small angle anticlockwise of the strike of the Alpine fault, or they are subparallel to it. On the basis of the SKS shear wave splitting data, Molnar et al. (1999) argued that the lithospheric upper mantle beneath the South Island of New Zealand comprises a zone of pervasive simple shear >300 km wide. Following inception of the plate boundary at ~45 Ma, ~400 km of the total ~850±100 km of dextral displacement between the Pacific and Australian Plates was accommodated on this zone (e.g. Sutherland 1995). In apparent accord with the distributed simple shear model, narrow terranes in the New Zealand crust are deflected across a conspicuous Z-shaped bend or orocline across a ~300 km-wide belt in the central part of the New Zealand landmass (Fig. 2). Some have argued (e.g. M. Moore, pers. comm.) that fast propagation azimuths for the SKS data in the South Island are parallel to the maximum finite stretching directions that would have accumulated if the present-day geodetic velocity field has persisted without change for the past ~6.5 Ma.

The goal of this paper is to use the deflection of crustal basement terranes in the South Island of New Zealand to calculate bulk finite strains related to the bending process. This can be viewed as a lithospheric-scale deformation experiment on real earth materials, allowing us to evaluate physical relationships between finite deformation, fabric development, and seismic anisotropy from...
Figure 2. (a) Modern plate tectonic setting of New Zealand, showing the locations of (b) a simplified tectonic map of a southern part of the South Island, and (c) a simplified tectonic map of a northern part of South Island. Numbers specify New Zealand basement terranes. Circled letters refer to South Island domains referred to in this paper.

the unscaled perspective of a ‘natural laboratory,’ rather than using the more common approach of triaxial deformation experiments on hand samples. The arrangement of narrow, crustal-scale geological markers at a high angle to the plate boundary in New Zealand provides a rare opportunity to conduct this type of analysis. The orientation of the principal stretching directions calculated from these crustal markers are compared with the fast polarization azimuths of available shear wave data from the mantle to test the idea that the finite deformation in the crust and mantle are similar, and that these layers are well-coupled. Our analysis, in contrast to some previous studies, does not rely on a strain analysis of the geodetically derived, instantaneous velocity field. This is an advantage because deformational fabrics are inherently a concept that is related to finite deformation. Strain rates in the past have not always been identical to those of the present day. Furthermore, we do not assume that the bulk deformation of the mantle is necessarily simple shear, but consider the possibility of a 3-D flow involving transpression. Lastly, our analysis is based on geological observations of finite strain using natural markers.

DEFORMED BASEMENT TERRANES IN SOUTH ISLAND

Elongated basement terranes in the continental crust of New Zealand, especially the Maitai/Dun Mountain terrane (Coombs et al. 1976), are exposed at the surface and can be traced through areas of poor exposure using geophysical techniques (e.g. Junction magnetic anomaly, Fig. 1, inset). These terranes are deflected through >70° of strike into a Z-shaped, recurved arc. Since its inception in the early Miocene (~25 Ma Cooper et al. 1987), the Alpine Fault...
has dextrally offset the ophiolitic Maitai terrane rocks by \( \sim 460 \) km (H. W. Wellman in Benson 1952). At \( \sim 6.4 \) Ma, Pacific–Australia plate motions changed from transform to obliquely convergent in the central part of the Southern Alps, causing the Alpine Fault to become oblique-reverse and resulting in \( \sim 90 \) km of shortening across the central part of the South Island (Norris et al. 1990; Walcott 1998).

Although some have attributed much of the curvature of basement terranes across the oroclinal bend to Mesozoic deformation (e.g. Kamp 1987; Bradshaw et al. 1997), available paleomagnetic data support a Cenozoic age for that bending (e.g. Little & Roberts 1997; Mumme & Walcott 1985; Townsend 2001). In addition, seafloor-based plate reconstructions of the Pacific–Australia plate margins are most plausibly reconciled with New Zealand’s surface geology if the terranes were approximately linear (uncurved) at the time of plate boundary inception through that region in the early Cenozoic (e.g. Molnar et al. 1975; Carter & Norris 1976; Walcott 1978, 1981a,b; White 1998; Sutherland 1999). Further evidence for an originally linear margin having been deformed into a curved shape by dextral shearing of early to mid-Cenozoic age is provided by the Oligocene–Miocene age of folds and faults that are strongly deflected in a clockwise sense across the deformation zone (Little & Mortimer 2001). Regardless of the age of the bend, the most important point here is that this structure is almost certainly the product of a profound finite strain in the lithosphere of New Zealand (e.g. Norris 1979).

**BENDING-RELATED FINITE STRAIN CALCULATIONS**

Assumptions and methods

Deformation of the brittle part of the Earth’s crust takes place dominantly by slip on major faults together with rotation of fault-bounded blocks (e.g. Garfunkel & Ron 1985; Jackson & Molnar 1990; McKenzie & Jackson 1983). Viewed at a small enough scale, these movements accommodate a bulk deformation in the crust approximating that of a continuum. We calculated bending-related finite strains using two different methods (Fig. 3). Both methods assume that: (1) the Dun Mountain ophiolite belt (DMOB) was originally approximately linear and (2) length changes parallel to the NE-trending margin have been negligible, whereas those perpendicular to it may have been significant. To maintain constant-volume (incompressibility condition), horizontal length changes (stretch) in the margin-perpendicular direction \( k \) must be balanced by a corresponding strain \( e_v \) in the vertical dimension (see Table 1):

\[
(1 + e_v) = 1/k
\]

where \( e_v \) = (final vertical length - initial vertical length)/initial vertical length.

The first method applies a ‘classical’ transpression model for the deformation (Sanderson & Marchini 1984; Fossen & Tikoff 1998) (Fig. 3a). As an analytical convenience, the margin-orthogonal shortening (‘pure shear’) can be assumed to accumulate first, followed by simple shearing. In fact, these components can accrue simultaneously (e.g. Fossen & Tikoff 1993), but the chosen factorization has no bearing on finite strain calculated from a comparison of initial and final states. The crust is treated as a continuum shear zone with displacement gradients both parallel \( (\gamma) \) and perpendicular \( (k) \) to the margin or deformation zone boundary (DB). The change in azimuth of a pair of non-parallel markers, relative to the DB, allows derivation of the above two deformation tensor components, from which the orientation and magnitude of the principal finite strains can also be derived (see Sanderson & Marchini 1984, eq. 4; Ramsay & Huber 1983, pp. 283–287). Simple shear is the special case where \( k = 1.0 \). Using these methods, one can calculate the length changes recorded by passive geological markers.

The second strain calculation method uses a ‘trellis’ or ‘rotating blocks’ model for bulk crustal strain (e.g. Garfunkel 1989; Jackson & Molnar 1990; Mandl 1987) (Fig. 3b). The narrow, fault-bounded nature of crustal terranes in New Zealand, and evidence for reverse (and in some cases, sinistral) movement along their boundaries, suggest that this might be an appropriate deformation mechanism (e.g. Norris et al. 1978; Little & Mortimer 2001). In contrast to the transpression model, faults do not lengthen or shorten, but always remain as lines of no finite elongation during deformation. The deformation is specified by only one parameter (either \( \gamma \) or \( k \)) and the azimuth change of a single marker is sufficient to calculate both of these (for the equations used, see Fig. 1 of Little & Roberts 1997). Depending on the orientation of the faults, the width of the deforming zone will expand or contract. Simple shear is not possible unless the faults are parallel to the deformation zone boundary.

**Application**

To apply these to the bending-related crustal deformation of the South Island we use distinct changes in the strike of the DMOB to
Table 1. Results of 3-D finite strain modelling using classical transpression and ‘trellis’ models for deflection of crustal markers in South Island, New Zealand.

| Domain | DB azimuth* | DMOB azimuth† | Fold azimuth‡ | Finite deformation components§ | Finite principal stretches** | Finite horizontal strain ratio†† | Finite horizon principal extension azimuth | Octahedral strain§§ (equivalent) | Converg. angle** (degrees) | Kinematic vorticity number*** |
|--------|--------------|---------------|---------------|-----------------|-----------------------------|-------------------------|---------------------------------|----------------------------------|-----------------------------|-----------------------------|
|        | (degrees)    | (degrees)     | (degrees)     | **γ**          | **k**                       |                         |                                 |                                  |                             |                             |
| A      | NA           | 310           | NA            | NA              | NA                         | NA                      | NA                              | NA                               | NA                          | NA                          |
| B      | 220          | 000           | 198           | 1.19 (1.19)     | 1.00                       | 1.76                    | 0.57                            | 1.00                             | 3.1                          | 190                         | 0.8                          |
| C      | 248 044 (031) | 230 (217)     | 1.32 (1.13)   | 0.75            | 3.06                       | 0.22                    | 1.49                            | 13.9                             | 219                         | 1.9                          | 12                           |
| D      | 235 037      | 221           | 0.56 (0.53)   | 0.89            | 1.06                       | 0.23                    | 4.11                            | 4.6                              | 217                         | 2.0                          | 69                           |

Note. – See text and Fig. 3 for explanation ideal transpression and trellis models. See Fig. 2 for definition of domains A–D.

*DB = deformation boundary’s azimuth: taken as parallel to Moonlight fault (domain B), Wairau fault (domain C), and Alpine fault (domain D).
†Strike of Dun Mountain ophiolite marker (uncertainty of ±5°). Parenthetical values account for the difference in strike of the Alpine fault between the southern (domains B, D) and northern (Wairau fault, domain C) regions. The parenthetical azimuth represents a marker trend, observed in domain C, which has been converted to an ‘equivalent’ azimuth in the southern South Island (fault-marker angle held constant).
‡Axial trace of kilometre-scale folds of Haast Schist (uncertainty of ±10°). See above for explanation of parenthetical data.
§Shear strain, γ, and margin-perpendicular stretch, k. Left-hand shear strain value corresponds to pure-then-simple shear factorization of finite deformation (after Sanderson & Marchini 1984), whereas the bracketed value corresponds to simultaneous simple and pure shear, using the equations in Fossen & Tikoff (1993). These latter values are not quoted for the trellis model because that deformation type is inherently non-steady state.
**(1 + e₁h) and (1 + e₂h) are horizontal maximum and minimum principal finite stretches; (1 + e₅) is vertical stretch.
††Rₑᵥ = (1 + e₁h)/(1 + e₂h).
§§After Nadai (1963).
***See Fig. 3 for definition of deformation boundary convergence angle, β.

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shearing (wrench) deformation in the New Zealand crust that was parallel to the Moonlight fault zone. Domain C lies on the NW side of the Alpine fault where the DMOB has a NE trend. The Wairau section of the Alpine fault in domain C strikes 10°–15° clockwise of its ~055° azimuth ~460 km further south. Domain D, adjacent to the Alpine fault in the Southern Alps, lies in the zone of active (post-6.4 Ma) oblique collision, and in the region where present-day geodetic shear strain rates are at a maximum (e.g. Pearson et al. 1995; Beavan et al. 1999). This domain has the most strongly de-flected markers relative to the local strike of the Alpine fault, and was affected by a late Cenozoic increment of transpressive deformation that is likely to have reinforced and tightened the crustal bend in this region (Little et al. 2002a,b).

We interpret domains from A to D as reflecting a strain gradient, with an increase in marker deflection and accumulated finite strain towards the Alpine fault. This view is supported by the marked structural attenuation and dismemberment of the NE-striking parts of terranes 3–5 (Fig. 2) as they approach the Alpine fault, relative to their apparently less rotated, north- or NW-striking parts (e.g. Rattenbury et al. 1998). In addition to the DMOB, a second set of angular strain markers, not parallel to the first, is provided by a regionally extensive set of kilometer-scale, upright folds deforming metamorphic foliation in the Haast Schist (Figs 2 and 4). These folds include the Earnslaw and Goulter synforms, which can be traced towards the Alpine fault for tens to hundreds of kilometers. Oligocene–early Miocene in age, these structures were referred to as ‘Moonlight generation’ by Norris et al. (1978) and Craw (1985), and called ‘Alpine folds’ by Cooper et al. (1987) (see also Rattenbury et al. 1998; White 1998; Turnbull 2001).

Figure 4. (a) Schematic plan-view of the four crustal domains in the South Island considered in this paper. Azimuths of markers and deformation boundaries in each domain used in the strain analysis are specified with three digit number (000–360). The letters A, B, C and D refer to the four crustal strike domains labelled in Fig. 2. Calculated horizontal finite strains are plotted as ellipses (relative to the undeformed circle in domain A). Relative convergence vectors across each domain boundary (transpression model only) are plotted as black arrows. The angle, $\beta$, is the angle between this vector and the domain boundary (see Fig. 3a). Contemporary (post ~3 Ma) Pacific–Australia convergence vector (Nuvel-1) is from DeMets et al. (1990). (b) Calculated vertical strain ellipses (in plane perpendicular to the margin). Domain D is interpreted as a remnant of domain C on the south side of the present-day Alpine Fault, which has been subjected to an additional late Cenozoic transpression increment in the Southern Alps.

Results

For each crustal domain, the strike angles of the markers, and the shear strains, convergence factors, and bending-related finite strain estimates are calculated from the deflection of these markers (Table 1 and Fig. 4). Both strain models calculate large finite strains during bending of the South Island crust. The bending-related finite strains calculated from the observed marker deflections differ in their orientation, magnitude, and degree of vertical stretch (thickening) for the several domains. Across the domain A–B boundary, the change in the azimuth of the DMOB is known to $\sim \pm 10^\circ$. For this angular uncertainty, the two methods combine to yield an estimate for azimuth of the maximum horizontal principal stretch (domain B) of $195^\circ \pm 5^\circ$. Significantly, this direction coincides with the mean axial trace ($190^\circ$–$200^\circ$) of the spatially coincident ‘Moonlight generation’ folds. This result supports a ‘wrench’ origin for those folds, and a simple shear ($\beta = 0$) type of deformation in domain B (Fig. 4a). In domains C and D, finite strain calculations using both the trellis and transpression models deviate from simple shear. This is especially true of the trellis model, which maximizes bending-related crustal thickening estimates as it minimizes the calculated horizontal strain ratios ($R_{hs} < 5.0$ in domain D). For the trellis model, the calculated largest principal stretch axes in domains C and D are vertical (Fig. 4b). The transpression model yields smaller, but still significant values of crustal thickening (up to a factor of 1.5), greater horizontal strain ratios, and greatest principal stretch axes that are horizontal. Despite these differences, the azimuths of the largest horizontal finite principal strain direction in domains C and D are relatively insensitive to the kinematic model, and the two methods yield maximum strain directions trending anticlockwise of the Alpine fault that differ from one another by $<5^\circ$. The uncertainty in the
mean trend of marker azimuths in domains C and D (±5°) yields a similar magnitude of uncertainty in the calculated azimuth of the finite principal strain. These relationships indicate that we can estimate, to within ±10°, azimuths of the finite principal strain direction. This relationship is important because that direction can be easily compared with the fast polarization azimuth of shear wave splitting.

The relative convergence azimuth (angle β in Fig. 3) for each deformation increment depends on the relative magnitudes of the shear strain (γ) to the margin-perpendicular stretch, (k):

\[ \tan(\beta) = (1 - k)/(k\gamma). \]

Results using the transpression model indicate a lower convergence angle (β) in the past than exists today between the Pacific and Australian plates (Fig. 4a), a result that is in accord with plate reconstructions (e.g. Sutherland 1995). This is true even in domain D, which overlaps the geologically observed locus of present-day oblique convergence (Beavan et al. 1999), and which may include a late Cenozoic component of bending. There, the present-day Pacific–Australian convergence angle, β, is ~18° (Nuvel-1A model of DeMets et al. 1990); this value exceeds the convergence angle calculated by us for the domain C–D deformation increment. The trellis model, on the other hand, yields estimates of the relative convergence angle that are much higher than the present-day value for all three deformation increments. We therefore view the ‘trellis’ concept as an unviable model for lithospheric deformation in New Zealand, and consider it no further in the discussion below.

**SEISMIC ANISOTROPY DATA**

The calculated directions of maximum horizontal stretch related to bending of the crustal terranes in the South Island of New Zealand, especially for domains B and D, correlate well with available data on the fast polarization azimuth of shear waves propagating through the subjacent lithospheric mantle (Fig. 5). Both follow a similar pattern of the distinct anticlockwise obliquity to the Alpine–Wairau fault, with this angle decreasing from ~30° to ~10° in spatial proximity to that structure (Klosko et al. 1999; Audoine et al. 2000). The available SKS shear wave splitting data indicate that the above-described zone of NE-trending fast polarization azimuths in the New Zealand lithosphere spans the >200 km onshore width of the southern South Island, but does not extend as far east as the Chatham Islands on the Pacific Plate (Fig. 1, inset). This supports the view that the strong seismic anisotropy fabric is the product of mantle flow in the Pacific–Australia plate boundary zone during the Cenozoic (Molnar et al. 1999). The relative contribution of the asthenosphere and the lithosphere in SKS studies is often a concern (e.g. Silver 1996; Savage 1999). Lack of variation of splitting parameters with distance across the slab in the lower North Island suggests that asthenospheric flow parallel to the slab may contribute to some of the observed anisotropy in that region (Marson-Pidgeon et al. 1999). However, the correspondence in fast polarization azimuths for SKS phases with those of local earthquakes down to depths of 100 km suggests a lithospheric source for some of the splitting, and increasing splitting with depth for earthquakes in the southern North Island also suggest a lithospheric source for the splitting (Audoine et al. 2000). Finally, recent data indicates that Pn anisotropy affecting refracted waves travelling horizontally through the upper mantle of the West Coast of the South Island reaches a minimum value of 10 ± 3 per cent; these data require that the Pn anisotropy resides in the lithospheric mantle (Scherwath et al. 2002), and it suggests that the anisotropy can reside entirely within the lithosphere, unless the anisotropy decreases with depth.

Several variations to a simple pattern are apparent in Fig. 5. The correlation between crustal strain and seismic anisotropy fabrics in domain C, although satisfactory, is complicated by an increase in the local variability of SKS shear wave polarization azimuths in the central part of the Southern Alps. In particular, two of the stations yielded fast polarization azimuths that are approximately parallel to the Alpine fault, rather than being anticlockwise of it (stations EWZA and MAYA in Klosko et al. 1999). Unlike the calculated finite strain, which increases in proximity to the Alpine fault, the magnitude of the time delay between the polarized phases does not show a corresponding increase in that direction. This relationship is expected, because experimental and simulated fabric data indicate that mantle peridotites will attain a peak magnitude of seismic anisotropy (δFs) of 8–9 per cent at finite strains (ε) as low as 50–60 per cent (e.g. Mainprice & Silver 1993).

At three stations west of the Alpine fault the fast polarization azimuth of SKS phases are parallel to the Alpine fault, rather than anticlockwise of it. This result, however, is contradicted by local S-wave polarizations for the same area, which are strongly anticlockwise, together leading to a somewhat inconclusive result for that region. Finally, our calculations predict no bending-related strain in domain A—the region used as an ‘undeformed’ reference, yet the SKS waves have measurable anisotropy there (again anticlockwise of the Alpine fault). In reality, domain A (SE Otago) lies in a zone of active reverse-slip faulting and folding (e.g. Norris et al. 1990). As it is a deformed reference, rather than an undeformed one, and as deformation has continued to the present day well after early mid-Cenozoic development of the oroclineal bend, our calculations probably underestimate the total crustal strain at this observation scale (especially in domain A).

**DISCUSSION AND CONCLUSIONS**

Coupling mantle and crustal deformation

This paper has presented evidence for a first-order similarity in azimuthal pattern between the fast polarization directions of shear waves and the principal axes of finite strain calculated for the crust using the deflection of markers across part of the Pacific–Australia plate boundary zone. This correlation supports the view that plate boundary deformation of the lithospheric mantle is widely distributed and at least in part coherent with the deformation field of the overlying crust, and contradicts the mantle-subduction model of lithospheric deformation in convergent orogens (e.g. Beaumont et al. 1996). The wide variation of shear wave splitting patterns observed at subduction zones (e.g. Savage 1999) does not allow us to argue, on the basis of the polarization directions alone, however, that a ‘mantle drip’ or subduction model could not be applicable in New Zealand.

In detail, the available shear wave splitting data suggest that the deep anisotropy sampled by SKS waves is wider in plan view than the modelled zone of bending-related strain. Thus SKS polarization azimuths in domains A and B are similar despite the discordance in strike of the crustal terranes between those two regions. The SKS fast azimuths are concordant with the calculated strain...
Figure 5. Simplified map of crustal terranes in South Island, New Zealand, including the Dun Mountain ophiolite/Maitai Terrane (black) showing calculated finite strains (transpression model, this study) together with measured fast polarization azimuths (and delays) of shear wave seismic anisotropy measurements in South Island, New Zealand (SKS data from Klosko et al. 1999; Molnar et al. 1999, local S-wave data from Audoine et al. 2000). Local S-wave data attributed by Audoine et al. (2000) to fracture-controlled crustal anisotropy are excluded.

Directions in the domains across which there is a clockwise deflection of terranes (domains B, C and D); however, measurable shear wave splitting persists eastward beyond this zone into domain A, where bending-related finite strain has been assumed to be negligible. At present, the total width of the zone of measurable seismic anisotropy into the offshore region to the east of domain A is unknown.

This disparity between the predictions of the bending strain model and the observations of shear wave splitting could be cited as evidence for crustal and mantle deformation being poorly coupled in the SE part of the South Island. An alternate, perhaps simpler explanation for this discrepancy, however, acknowledges that the present-day zone of active plate boundary deformation spans the entire width of the SE part of the South Island. The seismic anisotropy fabric in domain A may be the deep-lithospheric expression of the active reverse faulting, folding and WNW shortening taking place in the crust of coastal Otago, a late Cenozoic deformation post-dating orocline development.

Role of finite strain in development of olivine LPOs and seismic anisotropy

Fast shear wave polarization azimuths SE of the Alpine fault of the South Island New Zealand are chiefly oblique to the structural trend (Alpine fault) rather than parallel to it. The overall consistency of fast polarization azimuths suggests that their anticlockwise obliquity to the Alpine fault is real, despite the 10° error associated with the azimuth of fast polarization at any one site. This obliquity supports the view that the LPO of olivine, with or without recrystallization, is controlled chiefly by finite strain (e.g. Tommasi et al. 1999).
calculations suggest that seismic anisotropy fabrics can track finite strain trajectories in the lithospheric mantle at horizontal strain ratios (\(R_{xy}\)) of \(>6.0\), finite shear strains (\(\gamma\)) of \(>2.0\), and octahedral strains (\(\varepsilon_{o}\)) of \(>1.9\) without being substantially reoriented towards the shear plane by dynamic recrystallization. However, the fault-parallel SKS fast polarizations observed at several stations in (and to the west of) the central southern Alps may be caused by a reorientation of [100] crystallographic axes of olivine toward the shear plane by dynamic recrystallization (Zhang & Karato 1995; Bystricky et al. 2000). There is no evidence for such a process taking place at values of finite shear strains less than \(\sim2.0–2.5\). This is an important result of our natural ‘laboratory’ experiment, as it contrasts with the cited experimental studies, in which recrystallization-induced LPO changes took place at finite shear strains of only \(\sim1–2\). The apparently restricted distribution of this process beneath the central part of the Southern Alps on the eastern side of the Alpine fault may reflect locally higher strain rates (narrower deformation zone?) or higher heat flow in that region.

An alternate, non-rheological explanation for the locally fault-parallel trend of fast polarization azimuths at the two stations SE of the Alpine fault is possible. The patterns of flow in the crust and mantle may differ locally from one another beneath the central part of the Southern Alps, which is the region underlain by a zone of thickened lithospheric mantle (Stern et al. 2000). This part of the mantle may accommodate a larger component of margin-orthogonal convergence than the superjacent crust, a situation that would lead to fast azimuths being parallel to the margin (e.g. Tommasi et al. 1999). Further work is clearly required to test these competing hypotheses.

Implications of strain modelling for strain history and mantle fabrics

One conclusion derived from our finite strain modelling is that the deformation which caused bending of South Island terranes was a result of dextral transpression—with components of both horizontal shortening and vertical thickening—and not solely a dextral wrenching (simple shear) deformation. This result is consistent with the modelling of Norris (1979), whose work was based on a different, more arbitrary set of boundary conditions. It is, of course, possible to model the development of mantle fabrics using the present-day deformation. This result is consistent with the cited experimental studies, in which recrystallization-induced LPO changes took place at finite shear strains of only \(\sim1–2\). The apparently restricted distribution of this process beneath the central part of the Southern Alps on the eastern side of the Alpine fault may reflect locally higher strain rates (narrower deformation zone?) or higher heat flow in that region.

An alternate, non-rheological explanation for the locally fault-parallel trend of fast polarization azimuths at the two stations SE of the Alpine fault is possible. The patterns of flow in the crust and mantle may differ locally from one another beneath the central part of the Southern Alps, which is the region underlain by a zone of thickened lithospheric mantle (Stern et al. 2000). This part of the mantle may accommodate a larger component of margin-orthogonal convergence than the superjacent crust, a situation that would lead to fast azimuths being parallel to the margin (e.g. Tommasi et al. 1999). Further work is clearly required to test these competing hypotheses.

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