Evolution of the middle and lower crust during the transition from contraction to extension in Fiordland, New Zealand

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ABSTRACT

A deeply eroded orogen in southwest New Zealand preserves a record of changing flow patterns in the middle and lower crust during a transition from contraction and crustal thickening to extension and crustal thinning. The New Zealand exposures show that deformation patterns at mid-lower crustal depths were strongly influenced by local variations in crustal structure, temperature, composition, magmatic activity and rheology. Kinematic parameters, including the orientation of shear zone boundaries, the degree of non-coaxiality and kinematic partitioning, strain symmetry, and whether shear zones were thickening or thinning in different planes of observation, were extremely variable spatially and changed repeatedly over an 8–10 Ma period. However, despite this variability, several aspects of superposed deformations remained constant and can be assigned to distinctive tectonic settings. All shear zones that formed during the 119–111 Ma period in Northern Fiordland record flow involving bulk horizontal (layer-parallel) shortening, vertical (layer-perpendicular) thickening and >50% pure shear regardless of shear zone orientation, degree of non-coaxiality, strain symmetry, and temperature conditions. In contrast, all shear zones that formed during the 114–90 Ma period in Central Fiordland record flow involving vertical thinning, subhorizontal stretching and 40%–50% pure shear. These patterns are correlative with regional contraction and regional extension, respectively. The data suggest that at length scales of ~100 km and time scales of ca. 10 Ma, the effects of changing plate boundary dynamics on deformation patterns in the middle and lower crust can be distinguished from the effects of changing local boundary conditions, including steep temperature gradients and variable rheology.

INTRODUCTION

One of the most important reasons for the failure of plate tectonics as a description of how continental lithosphere deforms is lower crustal flow. Unlike in oceanic regimes, many zones of continental deformation are hundreds to thousands of kilometers wide and cannot be described by the rigid motion of plates that move laterally past and into one another along narrow boundaries (Buck, 1991; Clark and Royden, 2000; McKenzie et al., 2000; Jackson, J., 2002; Giorgis et al., 2004). Differences in the thickness, thermal profile, and strength of continental crust make lower crustal flow much more likely and important in continental...
regions than it is in oceanic regions (Royden, 1996; Beaumont et al., 2001). However, despite recognition of its importance, the characteristics and consequences of lower crustal flow are among the least understood aspects of geodynamics. For example, what types of flow patterns evolve in the lower crust as tectonic settings and the driving forces of continental deformation change? How do flow patterns change as local rheologies and physical conditions in the deep crust change? What are the length and time scales of middle and lower crustal heterogeneity?

The study of deep-crustal exposures provides an important and useful approach to answering these questions. Geophysical images provide instantaneous views of lower-crustal structure but the age and kinematic significance of deep-crustal fabrics commonly is difficult to resolve (Nemes et al., 1997; McBride and Knapp, 2002; Jackson, H.R., 2002). Natural exposures of ancient middle and lower crust potentially allow us to determine directly how deformation in the deep crust relates to other observable features. However, studies of the middle and lower crust in different settings also have shown that the rheology and thermal structure of the lower crust are extremely heterogeneous and can change rapidly during orogenic cycles (Miller and Paterson, 2001, Whitney et al., 2004; Rusmore et al., 2005; Karlstrom and Williams, 2006). This time-dependent, heterogeneous nature of deformation in most natural systems commonly obscures the nature of the forces that control lower-crustal deformation and complicates the application of numerical models to natural phenomena. Many numerical and analytical techniques employ algorithms that require steady-state conditions or limit the number of variables that operate simultaneously over geologic length and time scales (Royden, 1996; Lin et al., 1998, Jiang and Williams, 1998; Jiang et al., 2001; Beaumont et al., 2001). Our ability to predict the response of deforming continental lithosphere to changes in driving forces, including plate motions, relies on an adequate determination of lower-crustal behavior during orogenesis.

In this chapter, we report the results of a semiquantitative, field-based investigation of ductile flow in a deeply eroded section of exposed middle and lower crust in Fiordland, New Zealand (Fig. 1). The young age and well-constrained tectonic setting of these exposures allowed us to examine the physical response of the middle and lower crust to a major tectonic change from

Figure 1. General geologic map of western New Zealand after Wood (1972), Oliver and Coggon (1979), J.Y. Bradshaw (1989a), Daczko et al. (2002a), Tulloch and Kimbrough (2003), and Klepeis et al. (2004). Area near Resolution Island from Turnbull et al. (2005). Abbreviations are as follows: CS—Caswell Sound; DS—Doubtful Sound; LM—Lake Manapouri; LTA—Lake Te Anau; MS—Milford Sound; RI—Resolution Island. Boxes show areas of study.
contraction to extension as plate boundary forces evolved from 114 to 111 Ma. During this interval, both the plate boundary kinematics and the regional style of deformation in western New Zealand changed (Tulloch and Kimbrough, 1989; Gibson and Ireland, 1995; Spell et al., 2000; Scott and Cooper, 2006; Klepeis et al., 2007). We obtained the first direct measure of changing lower-crustal flow fields over length scales of several hundred square kilometers during this transitional period. The exposures allowed us to relate the kinematic evolution of different fabrics in areas that record distinctive histories of magmatism and high-grade metamorphism. The data suggest a lower-crustal structure that is much more heterogeneous and transient than previously believed.

CONFLICTING OBSERVATIONS AND TESTABLE HYPOTHESES IN FIODRLAND

Most convergent margins record cycles of contraction and crustal thickening followed by extension and crustal thinning over periods of several tens of millions of years. In western New Zealand, Early Mesozoic contraction and arc magmatism thickened the ancient margin of Gondwana to a crustal thickness of at least 45 km by ca. 120 Ma (Oliver, 1980; Bradshaw, J.D., 1985; Bradshaw, J.Y., 1989a; Clarke et al., 2000; Tulloch and Kimbrough, 2003). This activity accompanied subduction and convergence along the ancient continental margin of Gondwana (Tulloch and Kimbrough, 2003) and produced a linear belt of Triassic–Early Cretaceous intrusive rocks known as the Median batholith (Fig. 1; Mortimer et al., 1999).

By ca. 110 Ma, structures that are typical of extending lithosphere had formed within the crustal column, including metamorphic core complexes and rift basins (Gibson et al., 1988; Tulloch and Kimbrough, 1989; Gibson and Ireland, 1995; Spell et al., 2000; Kula et al., 2005; Scott and Cooper, 2006; Klepeis et al., 2007). Geochemical and geochronologic data indicate that subduction-related magmatism lasted until ca. 105 Ma (Tulloch and Kimbrough, 2003) and appears to have overlapped with the transition to extension (Waight et al., 1998). This transition coincided with a reorganization of plate boundaries that ended subduction along the Gondwana margin (Bradshaw, J.D., 1989; Tulloch and Kimbrough, 2003). By ca. 84 Ma, the New Zealand continent had rifted away from Australia and Antarctica as the Tasman Sea opened (Gaina et al., 1998). This history and the unusual degree of exposure allow us to test two leading hypotheses that describe how local variations in lower-crustal strength (effective viscosity) and temperature affect lithospheric behavior during orogenic contraction and extension.

One leading hypothesis that explains conditions in the lower crust during the period 120–110 Ma, suggests that the development of extensional features following a period of crustal thickening reflects horizontal flow in lower crust that has been weakened by heat, magmatism and partial melting. Published models suggest that gravitational instabilities caused by a thick, weak lower crust and variations in crustal thickness may promote the flow of material away from thick zones resulting in extension within the crustal column (McKenzie et al., 2000; Jackson, J., 2002). Relationships observed in Central Fiordland may support this hypothesis. Gibson and Ireland (1995) reported U-Pb isotopic data indicating that at or immediately before the initiation of extension in Central Fiordland, thick parts of the lower crust recorded elevated temperatures of >750–800 °C. Zircon extracted from high-temperature granulite-facies fabrics at Doubtful Sound (DS, Fig. 1) yielded an age of 107.5 ± 2.8 Ma (Gibson and Ireland, 1995). The chemistry and age of the zircon suggested that it represented a new generation of zircon growth at high temperatures. These data, and the dependence of crustal viscosity on temperature (e.g., Royden, 1996; Ellis et al., 1998), suggest that the lower crust was weak and flowed easily at the time extension initiated. However, new high-precision data reported by Flowers et al. (2005) conflict with this interpretation. U-Pb ages on titanite, rutile, and apatite reported by these authors indicate isobaric lower crustal cooling through the range 650–550 °C by 113.5–111 Ma (Table 1).

A possible resolution to these conflicting observations was presented by Klepeis et al. (2007) who suggested that the initiation of extension coincided with areas of the lower crust that were weakened by heat and magma, but the transition occurred by ca. 114 Ma, significantly earlier than previously suggested. This interpretation is based on fabric studies at Doubtful Sound and ages from high-temperature granulites obtained by Hollis et al. (2004). Nevertheless, given the uncertainties in available ages (Table 1), the time of the shift from regional contraction to regional extension could have occurred at any time in the interval 114–111 Ma. Evidence of magmatism during the interval 116–113 Ma (Table 1; Gibson and Ireland, 1998; Tulloch and Kimbrough, 2003; Hollis et al., 2004), granulite-facies metamorphism (Oliver, 1977; Oliver and Coggon, 1979; Oliver, 1980; Gibson and Ireland, 1995), and partial melting at Doubtful Sound (Hollis et al., 2004) support the interpretation of a weak lower crust at the time extension initiated.

An alternative hypothesis has been suggested on the basis of observations in Northern Fiordland. At Milford Sound (MS, Fig. 1), less than 100 km to the north of Doubtful Sound, metamorphic mineral assemblages and U-Pb isotopic data suggest that the lower crust there had cooled from temperatures of >800 °C to 650 °C prior to 111 Ma and probably as early as ca. 116 Ma (Daczko et al., 2002a; Hollis et al., 2003; Klepeis et al., 2004; Flowers et al., 2005). Even though partial melting appears to have occurred within the lower-crustal section at ca. 120 Ma, piston cylinder experiments on unmelted samples of dioritic gneiss from Pembroke Valley (Antignano, 2002) and petrological analyses (Daczko et al., 2001b) suggest that the total volume of melt probably remained low (Klepeis et al., 2003). The experiments showed that fluid-absent melting was controlled by the decomposition of hornblende ± clinzoisite to produce garnet + melt and resulted in low (≤10 volume %) melt volumes at all temperatures up to 975 °C. Low melt volumes would have helped the lower crust remain strong even as it partially melted. Heat loss, the crystallization of magma, and the efficient removal of partial melt from the lower crust are interpreted to have resulted in isobaric cooling.
TABLE 1. SELECTED MINERAL AGES FROM IGNEOUS AND METASEDIMENTARY ROCKS IN NORTHERN AND CENTRAL FIO RDLAND

<table>
<thead>
<tr>
<th>Location</th>
<th>Rock type</th>
<th>Method</th>
<th>Mineral</th>
<th>Corrected age (Ma)</th>
<th>Source</th>
<th>Interpretation</th>
</tr>
</thead>
<tbody>
<tr>
<td>Northern Fiordland</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Mt. Kepka</td>
<td>Syntectonic felsic dike</td>
<td>U-Pb, LA ICPMS</td>
<td>Zircon, r</td>
<td>119.0 ± 4.7</td>
<td>Marcotte et al. (2005)</td>
<td>ICSZ outlasted dike emplacement</td>
</tr>
<tr>
<td>Mt. Kepka</td>
<td>Syntectonic dike</td>
<td>U-Pb, LA ICPMS</td>
<td>Zircon, c</td>
<td>115 ± 3.6</td>
<td>Marcotte et al. (2005)</td>
<td>ICSZ outlasted dike emplacement</td>
</tr>
<tr>
<td>Mt. Kepka</td>
<td>Syntectonic dioritic dike</td>
<td>U-Pb, LA ICPMS</td>
<td>Zircon, r, c</td>
<td>117.4 ± 3.9</td>
<td>Marcotte et al. (2005)</td>
<td>ICSZ outlasted dike emplacement</td>
</tr>
<tr>
<td>Anchorage Cove</td>
<td>Syntectonic dioritic dike</td>
<td>U-Pb, LA ICPMS</td>
<td>Zircon, r</td>
<td>121.7 ± 4.2</td>
<td>Marcotte et al. (2005)</td>
<td>GSSZ outlasted dike emplacement</td>
</tr>
<tr>
<td>Mt. Daniel</td>
<td>WFO diorite</td>
<td>U-Pb, SHRIMP</td>
<td>Zircon, c</td>
<td>121.8 ± 1.7</td>
<td>Hollis et al. (2004)</td>
<td>WFO emplacement age</td>
</tr>
<tr>
<td>Selwyn Creek</td>
<td>Dioritic gneiss</td>
<td>U-Pb, SHRIMP</td>
<td>Zircon, r, c</td>
<td>125-115</td>
<td>Hollis et al. (2003)</td>
<td>Age of granulite-facies metamorphism</td>
</tr>
<tr>
<td>Caswell Sound</td>
<td>Monzodiorite</td>
<td>U-Pb, LA ICPMS</td>
<td>Zircon, c</td>
<td>116.8 ± 3.7</td>
<td>Klepeis et al. (2004)</td>
<td>Lower limit Caswell fold-thrust belt</td>
</tr>
<tr>
<td>Mt. Ada</td>
<td>Syntectonic dioritic dike</td>
<td>U-Pb, LA ICPMS</td>
<td>Zircon, r</td>
<td>115.7 ± 3.8</td>
<td>Klepeis et al. (2004)</td>
<td>ICSZ outlasted dike emplacement</td>
</tr>
<tr>
<td>Caswell Sound</td>
<td>Dioritic dike</td>
<td>U-Pb, LA ICPMS</td>
<td>Zircon, c</td>
<td>118.7 ± 3.8</td>
<td>Klepeis et al. (2004)</td>
<td>Lower limit Caswell fold-thrust belt</td>
</tr>
<tr>
<td>Arthur River Complex</td>
<td>Gabbroic gneiss</td>
<td>U-Pb, SHRIMP</td>
<td>Zircon, r</td>
<td>120-110</td>
<td>Hollis et al. (2003)</td>
<td>Ages of high-T metamorphism</td>
</tr>
<tr>
<td>Central Fiordland</td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>0359, 0355</td>
<td>Orthogneiss, calcisilicate</td>
<td>U-Pb, ID</td>
<td>TiHitite</td>
<td>113.4–111</td>
<td>Flowers et al. (2005)</td>
<td>Cooling through 550-650°C</td>
</tr>
<tr>
<td>Doubtful Sound</td>
<td>WFO</td>
<td>K-Ar</td>
<td>Hornblende</td>
<td>98–93</td>
<td>Gibson et al. (1988)</td>
<td>Cooling through ~500°C</td>
</tr>
<tr>
<td>0361</td>
<td>Granulite in WFO</td>
<td>U-Pb, ID</td>
<td>Rutile</td>
<td>~70</td>
<td>Flowers et al. (2005)</td>
<td>Cooling through 400-450°C</td>
</tr>
<tr>
<td>Wet Jacket</td>
<td>Diorite</td>
<td>U-Pb, ID</td>
<td>Zircon</td>
<td>116.6 ± 1.2</td>
<td>Tulloch and Kimbrough (2003)</td>
<td>Youngest known phase of the WFO</td>
</tr>
<tr>
<td>CA10</td>
<td>Granulite in WFO</td>
<td>U-Pb, SHRIMP</td>
<td>Zircon</td>
<td>114 ± 2.2</td>
<td>Hollis et al. (2004)</td>
<td>Age of granulite facies metamorphism</td>
</tr>
<tr>
<td>CA90</td>
<td>WFO diorite</td>
<td>U-Pb, SHRIMP</td>
<td>Zircon</td>
<td>115.6 ± 2.4</td>
<td>Hollis et al. (2004)</td>
<td>Crystallization age of WFO</td>
</tr>
<tr>
<td>CA39</td>
<td>Paragneiss</td>
<td>U-Pb, SHRIMP</td>
<td>Zircon, r</td>
<td>117.7 ± 2.8</td>
<td>Hollis et al. (2004)</td>
<td>Age of high-T metamorphism</td>
</tr>
<tr>
<td>Joseph Point</td>
<td>Syntectonic dioritic dike</td>
<td>U-Pb, LA ICPMS</td>
<td>Zircon, c</td>
<td>102 ± 1.8</td>
<td>Klepeis et al. (2007)</td>
<td>Age of DSSZ</td>
</tr>
<tr>
<td>0460</td>
<td>Post-extension felsic dike</td>
<td>U-Pb, LA ICPMS</td>
<td>Zircon, c</td>
<td>88.4 ± 1.2</td>
<td>King et al. (2008)</td>
<td>Upper limit of extension</td>
</tr>
<tr>
<td>Mt. Irene</td>
<td>Syntectonic dike</td>
<td>U-Pb, LA ICPMS</td>
<td>Zircon</td>
<td>107.3 ± 0.8</td>
<td>Scott and Cooper (2006)</td>
<td>Age of MISZ</td>
</tr>
</tbody>
</table>

*Sample locations are indicated in Figure 2. ID—Isotope dilution; ICSZ—Indecision Creek shear zone; GSSZ—George Sound shear zone; WFO—Western Fiordland Orthogneiss; DSSZ—Doubtful Sound shear zone; MISZ—Mt. Irene shear zone. r indicates zircon rim age; c indicates zircon core age.
zones of deformation were thickening or thinning in different
ticity vectors, average degrees of non-coaxiality, and whether
boundaries, bulk shear directions, orientation of the average vor-
superposed deformations, including the orientation of shear zone
a number of different kinematic and geometric parameters for
during exhumation. To overcome this problem, we evaluated
heterogeneity and the possibility of at least some reorientation
contractional versus extensional tectonics because of the extreme
localities, especially in zones of flat-lying shear zones, shear
ctions (Oliver, 1977; J.Y. Bradshaw, 1989a, 1989b; Gibson and
Ireland, 1995; Clarke et al., 2000; Daczko et al., 2001a, 2001b,
2002a, 2002b; Hollis et al., 2004), Northern and Central Fiord-
land record divergent tectonic histories. Nevertheless, because
they reflect observations made in different parts of Fiordland, it
is possible that elements of both hypotheses are correct. This pos-
sibility is explored in detail in the “Discussion” of this article.

**APPROACH TO THE ANALYSIS OF MID-LOWER
CRUSTAL DEFORMATION**

To resolve the conflicting interpretations described in the
previous section, we measured kinematic patterns in shear zones
and superposed ductile fabrics in both Northern and Central
Fiordland that evolved through the transition from contraction
to extension during the interval 126–90 Ma. Our goal was to
determine which deformations in each locality, if any, displayed
patterns that could be related to each tectonic regime and how
the effects of extremely heterogeneous physical conditions also
may have influenced the patterns. This approach relied on linking
kinematic data with previously published age determinations
and thermobarometry from key localities. Salient age determinations
are summarized in Table 1. Key localities in Northern and Cen-
tral Fiordland are shown in Figures 2A and 2D, respectively. We
also present textural and microstructural data from lower-crustal
fabrics that help us link the results of previously published ther-
mbobarometry to specific fabrics in lower-crustal shear zones.

Part of the problem with analyses of lower-crustal deformation
is that both Northern and Central Fiordland display arrays
of flat, moderately dipping, and steep fabrics (Fig. 2). In most
localities, especially in zones of flat-lying shear zones, shear
zone geometry and sense of shear are insufficient to distinguish
contractional versus extensional tectonics because of the extreme
heterogeneity and the possibility of at least some reorientation
during exhumation. To overcome this problem, we evaluated
a number of different kinematic and geometric parameters for
superposed deformations, including the orientation of shear zone
boundaries, bulk shear directions, orientation of the average vor-
ticity vectors, average degrees of non-coaxiality, and whether
zones of deformation were thickening or thinning in different
planes of observation (i.e., sectional kinematic vorticity num-
bers). Here we use the terms thinning and thickening in the sense
that the boundaries of the deforming zone either moved toward
or away from each other, respectively (Lin et al., 1998). Estimates
of mean kinematic vorticity \( \left( W_m \right) \) provided us with a means of
comparing the relative contributions of pure shear and simple
shear for distinctive deformations. Absolute values of \( W_m \) vary
from 0 to 1 and represent a nonlinear ratio between coaxial and
non-coaxial components of deformation. Low numbers are highly
coaxial and high numbers are mostly non-coaxial. This approach
is a convenient way of determining whether a shear zone was
thinning (positive \( W_m \) values) or thickening (negative \( W_m \) values)
in different planes. These and other parameters, commonly used
in studies of ductile deformation, are defined below in the context
of our measurements. We also refer the reader to other publica-
tions for additional information (e.g., Lister and Williams, 1983;
Passchier, 1987; Passchier and Urai, 1988; Simpson and DePaor,
1993; Tikoff and Fossen, 1995; Klepeis et al., 1999; Jiang et al.,
2001; Daczko et al., 2001a; Bailey and Eyster, 2003; Law et al.,
2004; Bailey et al., 2007).

To determine kinematic patterns for each deformation we
identified local and regional strain gradients and determined
how the geometry of structures varied across them. Three of the
techniques we describe involved determining: angular changes
in deformed vein and dike sets; rotation histories of asymmetric
porphyroclasts; and progressive changes in the orientation of
structural elements such as foliations, fold hinge lines and mineral
stretching lineations. For some of the deformations, kinematic
patterns have been reported in previous publications. In these
cases, we provide a brief summary of salient results for compari-
son with other deformation events. Our study represents the first
comprehensive comparison of all kinematic data from Fiordland.
In all cases, our reference frame was the upper and lower con-
tacts of the Western Fiordland Orthogneiss and the orientation
of a regional compositional layering that parallels these contacts
throughout the western side of the orogen. Thermobarometric
data suggest that this layering originally formed in an approxi-
mately horizontal orientation and subsequently has been tilted
(see also Klepeis et al., 2004, 2007). The P-T determinations also
indicate that paleodepths increase from north to south across this
layering and allow us to reconstruct paleohorizontal. In addition,
most of the techniques we used, including analyses of kinematic
vorticity, are insensitive to reorientation and tilting.

**NORTHERN FIORDLAND**

**Geometry, Age, and Sequence of Structures**

In Northern Fiordland (Fig. 2A), four sets of superposed shear
zones record “snapshots” of evolving strain patterns during
cooling from peak supra-solidus conditions (\( P = 14 \text{kbar} \), \( T >
800 \text{ °C} \)) to upper amphibolite facies conditions (\( P = 14 \text{kbar} \), \( T =
650 \text{ °C} \)). Structures in these shear zones allowed us to reconstruct
changing kinematic patterns in the lower crust that accompanied
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cooling during the period 126–111 Ma. Klepeis et al. (2004) reconstructed the regional sequence of deformation, which we summarize here.

Between ca. 126 Ma and ca. 120 Ma, arc-related magmatism resulted in the emplacement of a >10-km-thick batholith, the Western Fiordland Orthogneiss. The oldest and hottest shear zones (T>800 °C) in Northern Fiordland record supra-solidus deformation at the margins of intrusive sheets that make up this batholith. One of the best exposures of these shear zones occurs at Mount Daniel (Fig. 3) where a zone of penetrative magmatic foliations separates the Western Fiordland Orthogneiss above from the Milford Gneiss below. This shear zone is defined by successive sheets of intrusive rocks that were deformed at supra-solidus conditions during emplacement of the Western Fiordland Orthogneiss. Textures that record flow in a semi-molten state are described in detail by Klepeis and Clarke (2004).

Another layer-parallel shear zone forms a ductile mid-crustal fold-thrust belt at the top of the Western Fiordland Orthogneiss at Caswell Sound (Figs. 2A and 2C). This belt was first identified by Daczko et al. (2002a). Both the Mt. Daniel and Caswell shear zones are generally less than 1 km thick and record sub-horizontal, layer-parallel shortening at high angles to the northeast-southwest trend of the Early Cretaceous magmatic arc. Deformation in the Caswell shear zone also accompanied emplacement of the Western Fiordland Orthogneiss and localized within its contact aureole, where strength contrasts between hot, magmatic material and cooler host rock occurred (Fig. 3C). Similar structures formed during arc magmatism on Stewart Island...

Figure 2. Simplified geologic maps of northern (A) and central (D) Fiordland. Two structural profiles are shown for each area: (B) and (C) for northern Fiordland and (E) and (F) for central Fiordland. Letters refer to key localities discussed in the text. Bold dashed arrows in (A) refer to regional-scale strain gradients along the margins of the Indecision Creek shear zone. Dashed lines in cross sections are the traces of foliation. Abbreviations are as follows: AC—Anchorage Cove; CS—Caswell Sound; JP—Joseph Point; MA—Mt. Ada; MD—Mt. Daniel; MI—Mt. Irene; MK—Mt. Kepka; P—Pembroke Valley; SC—Supper Cove; SH—Steep Hill; SWC—Selwyn Creek; W—Worsley; WJ—Wet Jacket. Numbered abbreviations in (B) refer to samples listed in Table 1, patterns are same as in Figure 1.
Evolution of the middle and lower crust during the transition from contraction to extension in Fiordland, New Zealand

The input of heat accompanying emplacement of the Western Fiordland Orthogneiss resulted in the partial melting of the lower crust at temperatures of 750°–850° C (Clarke et al., 2000; Daczko et al., 2001b). The effects of this thermal pulse are reflected in the occurrence of migmatite and granulite-facies mineral assemblages within and up to 10 km below the batholith. Thin metamorphic rims around Paleozoic and Mesozoic zircon cores from below the batholith have yielded ages of ca. 120 Ma, suggesting that they reflect a period of partial melting and granulite-facies metamorphism that coincided with the emplacement of the batholith (Tulloch et al., 2000; Hollis et al., 2003). During this period of high-temperature metamorphism, a series of steep, thin (1–3 m) shear zones formed within and below the batholith (Figs. 4A and 4B). Analyses of the mineral assemblages that define foliations in the shear zones, including garnet, pyroxene, hornblende, plagioclase, rutile, and quartz, suggest that they record deformation at transitional granulite-facies conditions of \( \approx 670 \, ^\circ \text{C} \) and \( \approx 14 \, \text{kb} \) (Daczko et al., 2001a). These shear zones cut and recrystallize older foliations in migmatitic gneiss (Figs. 4A and 4B). Most contain mylonitic foliations defined by dynamically recrystallized plagioclase and stretched garnet, hornblende and clinozoisite aggregates. A sinistral set is dominant and strikes northeast. A second, subordinate set is dextral and strikes northwest. Both sets display gently plunging, hornblende mineral lineations and abundant asymmetric sense-of-shear indicators. They are preserved best in Pembroke Valley (P, Fig. 2A) north of Milford Sound (Daczko et al., 2001a).

Cross-cutting the steep shear zones in Pembroke Valley are a series of gently dipping, layer-parallel shear zones. Each of these shear zones contains a central zone of 7–10-m-thick mylonite (Fig. 5A). Connecting each central zone of mylonite are thin (<1 m thick), curved shear bands that dip to the southeast and northwest (Fig. 5B). These shear bands envelop imbricated, asymmetric pods of mafic-intermediate orthogneiss. Relative displacement between groups of pods is indicated by the disparate orientations of foliation planes across mylonitic boundaries. Mineral lineations within the shear bands display similar trends but steeper plunges than the lineations that occur in the central mylonite zones. Boudinaged mafic layers and stretched garnet porphyroblasts show that these mineral lineations represent true stretching directions. The sense of shear displayed by these mylonitic zones and shear bands is dominantly top-to-the-northwest in a direction normal to the trend of the magmatic arc. Daczko et al. (2001a) interpreted these shear zones as forming part of a lower-crustal thrust duplex. This deformation also occurred at transitional granulite-facies conditions of \( \approx 670 \, ^\circ \text{C} \) and \( \approx 14 \, \text{kb} \) (Daczko et al., 2001a).

Following development of the gently dipping, duplex-style shear zones, two near vertical 10–15-km-wide zones of penetrative deformation, called the Indecision Creek and George Sound shear zones (Figs. 2A–2C), formed within the lower-crustal section. The structural elements that define the shear zones and the framework of the kinematic analysis described in this study are described by Marcotte et al. (2005). Both shear zones cut granulite-facies metamorphic mineral assemblages (ca. 120 Ma) in the Western Fiordland Orthogneiss and its granulite-facies host rock. Mineral assemblages, U-Pb determinations and cross-cutting relationships indicate that the shear zones developed before the shift to extension and before decompression and exhumation of the lower crust (Daczko et al., 2002c; Marcotte et al.,
2005). Their formation reflects a progressive widening and thickening of deformation zones that accompanied lower-crustal cooling, following emplacement of the Western Fiordland Orthogneiss batholith. The George Sound shear zone cuts up-section and merges with the mid-crustal Caswell fold-thrust belt. U-Pb data on zircon from the fold-thrust belt (Klepeis et al., 2004) and the Indecision Creek and George Sound shear zones (Marcotte et al., 2005) indicate that these structures all evolved during the same 119–111 Ma interval (Table 1).

The conditions of deformation that accompanied the formation of the steep shear zones during the interval 119–111 Ma were very heterogeneous, reflecting localized magmatic activity, some partial melting and rapidly changing lower-crustal temperatures. Kyanite- and paragonite-bearing assemblages indicate that the Indecision Creek shear zone records isobaric (P = 14 kb) cooling from >750 °C to ≈650 °C (Daczko et al., 2002c). The internal structure of the shear zones is complex and includes multiple generations of foliations and folds (Fig. 6). Much of this complexity reflects a close relationship between deformation and the pre- and syn-tectonic emplacement of dikes displaying variable orientations (Marcotte et al., 2005). In some places, such as Selwyn Creek (SK, Fig. 2A) and Mt. Kepka (MK, Fig. 2A), garnet-bearing leucosome and numerous syntectonic dikes provide evidence of local partial melting, transient high temperatures and the transport of magma.

Microstructures in dioritic gneiss provide additional information on the conditions of deformation during this critical period immediately before the transition to extension in Northern Fiordland. In the high strain zones, large porphyroclasts of hornblende are surrounded by thin, commonly asymmetric mantles of small recrystallized grains (Fig. 7A). The mantles are stretched parallel to the foliation and grade into a fine-grained matrix consisting
Figure 5. (A) Photograph of the Pembroke thrust fault. Photo by Stephen Marcotte. Note people in hanging wall for scale. Basal shear zone is one of several sub-parallel thrust zones that form a vertically stacked sequence. White lines are vertical transects along which structures were measured. (B) Vertical profiles showing the geometry of garnet-granulite-facies foliations within and above the basal shear zone. Shaded regions are pegmatite dikes. (C) Equal area lower hemisphere stereographic projections showing orientation of foliation (poles) planes and hornblende mineral lineations in the vertical transects. Structures allow the vorticity normal section (VNS), the vorticity vector and the shear zone boundaries (SZB) to be defined. (D) and (E) Histograms showing angle between foliation planes and the basal shear zone in the VNS plane. Data show a flow regime dominated by pure shear (~90%) and involving thickening in a vertical plane. (F) Sketch showing geometry of thrust zone, orientation of the VNS, and the shear zone boundaries. Sketch shows expected orientation of the X-attractor in a vertically lengthening shear zone.
mostly of recrystallized plagioclase (Fig. 7B). The cores of hornblende porphyroclasts display bent crystal lattices surrounded by bands of recrystallized grains, many of which display cuspate-lobate grain boundaries (Fig. 7B, white arrows). These grain boundary shapes suggest that grain boundary migration was an important process that controlled the recrystallization of hornblende. In addition, the lack of neoblastic amphibole growth and the mixing of recrystallized hornblende and plagioclase grains in the mantle and tails of hornblende porphyroclasts (Fig. 7A) suggest that some diffusion-accommodated grain-boundary sliding also occurred (Gower and Simpson, 1992). In the matrix, deformation was accommodated mainly by dynamic recrystallization of plagioclase crystals. Plagioclase displays evidence of grain size reduction (Fig. 7C), cuspate-lobate grain boundaries, and core-mantle structures (Fig. 7D), indicating high grain boundary mobility.

These textures suggest that deformation in the Indecision Creek shear zone occurred by recrystallization-accommodated dislocation creep in plagioclase (Tullis and Yund, 1985; Ji and Mainprice, 1990) and the recrystallization of hornblende, with some grain boundary sliding. Experimentally deformed aggregates of amphibole and plagioclase (Hacker and Christie, 1990) display amphibole microstructures at temperatures of ≥650 °C that are similar to those we observed. Evidence of rotational recrystallization in plagioclase and recovery by dislocation climb suggest deformation at temperatures of >550 °C (Olsen and Kohlstedt, 1985; Pryer, 1993), which are consistent with those obtained using thermobarometric techniques (Daczko et al., 2002c).

**Estimates of Mean Kinematic Vorticity in a Horizontal Plane**

Patterns of deformed and rotated veins adjacent to the steep shear zone pairs in Pembroke Valley reveal kinematic patterns in a horizontal (layer-parallel) plane over an area of 0.75 km². Daczko et al. (2001a) used sets of near vertical veins as strain markers to infer that the shear zone pairs record a two-dimensional type of flow characterized by subhorizontal (layer-parallel) shortening and stretching with little displacement recorded in the vertical (layer-perpendicular) plane. The monoclinic rotation history of over 250 veins along the margins of shear zones allowed these authors to define fields of back rotation and forward rotation associated with the deformation (Fig. 4C). The dominant rotation sense was sinistral. Back rotation was indicated by the dextral rotation of a vein where it entered and was reoriented by a sinistral shear zone (see figure 6 in Daczko et al., 2001a). This history also allows the
determination of the orientation of the vorticity vector and vorticity normal section (VNS, Fig. 4E). The orientation and width of the fields of back rotation (Fig. 4D) provide an estimate of average sectional kinematic vorticity numbers ($W_m$) of $0.63 < W_m < 0.73$ for the horizontal plane (Daczko et al., 2001a). Daczko et al. (2001a) obtained similar results from measures of foliation traces and an estimate of the orientation of instantaneous strain axes at specific sites. This estimation involved measures of the orientation of >50 oblique grain-shape foliations preserved inside low strain pods within the margins of shear zones (see figure 8 of Daczko et al., 2001a). Wallis (1995) provides a full description of this technique. The similarity among the average and instantaneous measures of sectional kinematic vorticity suggests that the deformation was approximately steady at the scale of observation ($0.75$ km$^2$). These patterns show that deformation in a horizontal plane involved subhorizontal arc-normal (northwest-southeast) shortening and arc-parallel (northeast-southwest) stretching (Fig. 4E) with ~43%–53% pure shear.

**Estimates of Mean Kinematic Vorticity in a Vertical Plane**

The geometry of structures that define the gently dipping, duplex-style shear zones in Pembroke Valley (Fig. 5A) allowed the estimation of the kinematics of deformation in a vertical (layer-perpendicular) plane. Above each basal shear zone, mineral lineations vary smoothly within a single plane that strikes to the northwest ($131^\circ$) and dips steeply to the southwest ($84^\circ$; Fig. 5B). This pattern defines the vorticity normal section (VNS) and allows estimation of the average orientation of the vorticity vector ($\omega$) for the deformation (Fig. 5C). The vorticity vector is constrained to lie normal to the VNS for steady-state deformations. In addition, virtually all asymmetric microstructures such as recrystallized tails on porphyroclasts were observed in the VNS plane. The bulk shear direction for the shear zones is given by the intersection between the VNS and the shear zone boundaries. This direction parallels hornblende mineral lineations within the basal mylonite zones (Fig. 5C). The boundaries of the shear
zones parallel the thick zones of mylonite that underlie rotated foliation planes. These relationships establish that the deformation involved displacements in an interconnected network of flat and steep shear zones that displays monoclinic symmetry. Similar mid-crustal geometries have been described by Karlstrom and Williams (2006).

Structural relationships in areas where the vertical stacking of asymmetric pods are most abundant are distinct from those in areas that lack the antiformal stacks. Each pod contains a curved, oblique grain-shape foliation defined by coarse, elongate aggregates of garnet, clinopyroxene, plagioclase, hornblende, and clinzoisite. The angle between these internal foliations and the surrounding mylonitic foliations is at a maximum near the centers of each pod (Fig. 5B). The oblique foliations are either truncated by or smoothly merge into parallelism with the enveloping mylonites. The smooth deflections are accompanied by a reduction in the grain size of plagioclase and hornblende and an increase in the degree of grain elongation. This grain size reduction was accomplished by the dynamic recrystallization of plagioclase and the breaking apart of hornblende grains. The acute angles ($\theta$) between the traces of the oblique foliations and the shear zone boundaries display angles as high as $\theta = 70-90^\circ$ in areas of antiformal stacking (Fig. 5D). In contrast, areas that lack stacked lenses display maximum angles of $\theta = 30-43^\circ$ and the spread of data is more homogeneous (Fig. 5E). The unusually steep angles of $\theta = 70-90^\circ$ only occur in areas of stacked pods and suggest that the deformation in these zones was dominated by pure shear and stretching at high angles to the shear zone boundaries. Mineral lineations in these stretching pods are steeply plunging within the VNS (Fig. 5C). The formation of fabrics oriented nearly $90^\circ$ from the shear zone boundary is expected in zones of pure shear thickening (Teyssier and Tikoff, 1999).

If the areas outside the zones of thickened pods reflect the geometry of structures prior to the formation of the antiformal stacks then a comparison between these zones provide a means of estimating a mean kinematic vorticity number for the vertical plane. This assumption appears reasonable because the areas that lack antiformal stacks occur adjacent to areas that display the pile-ups. In addition, areas of stacked pods display ultramylonite textures indicating that these areas are highly strained. If correct, then the increase in angle ($\dot{\theta}$) from outside to inside the zones of thickened pods suggests that layer-parallel shortening and vertical thickening resulted in a progressive steepening of foliation planes. This interpretation agrees with the qualitative evidence of vertical thickening within a flow regime locally dominated by pure shear (i.e., less than 20% simple shear).

For a thickening shear zone undergoing steady flow, numerical models predict that the directions of maximum principal stretch (X) lie within or close to the trace of the VNS (Jiang and Williams, 1998; Lin et al., 1998; Jiang et al., 2001). This relationship holds true for triclinic as well as monoclinic shear zones. For steady-state deformations (Fig. 5F), the stable end orientation of the X-axis parallels an extensional flow apophysis (or the X-attractor following the definition of Passchier, 1987). This apophysis is inclined relative to the shear zone boundary in a thickening shear zone (Jiang and Williams, 1998; Teyssier and Tikoff, 1999). With increasing strain the X-direction will rotate and converge on the extensional flow apophysis ($\dot{A}$, Fig. 5F).

An estimate of the orientation of the extensional flow apophysis in these shear zones can be obtained if the trace of the rotating foliations approximately tracked the direction of maximum principal stretch (see descriptions by Wallis, 1995, and Daczko et al., 2001a). Even if the foliations did not track the finite strains perfectly they should approach the stable end direction if strains were high enough. Using this approach, the highest angles ($\dot{\theta}$) between traces of oblique mylonitic foliations and the shear zone boundaries in the VNS provide a minimum estimate of the sectional kinematic vorticity number ($W_m$). The range $\dot{\theta} = 70-80^\circ$ suggests $-0.17 < W_m < -0.34$ on the scale of individual 20-m-thick antiformal stacks. The minus sign indicates thickening (after Simpson and DePaor, 1993).

Three-Dimensional Strain and Kinematic Vorticity Patterns at the Regional Scale

Kinematic data from Pembroke Valley suggest that deformation prior to ca. 111 Ma, at least locally, involved components of horizontal shortening and vertical thickening and >50% pure shear at the scale of several square kilometers. We tested this result at the scale of the entire lower-crustal section by measuring variations in the orientation of structural elements across two regional strain gradients. To use this approach, we assumed that the foliations and mineral lineations formed together and approximately tracked finite strain directions during deformation. We evaluate these assumptions later in this section.

One of the best exposed regional strain gradients occurs in a 5-km-wide zone adjacent to the western boundary of the Indecision Creek shear zone (SH to MA in Fig. 2A). From west to east across this zone (Marginal domain in Fig. 8A), a positive strain gradient is defined by an increase in fold tightness (decrease in interlimb angle) and a decrease in the angles between deformed dikes and vein sets (strain data reported by Marcotte et al., 2005). Within the central parts of the shear zone, these folds and dikes are almost completely transposed parallel to foliation (Fig. 6). These geometric changes reflect an increase in the component of shortening from west to east across the section. Another, narrower (1–2-km-wide) strain gradient occurs from east to west across a second marginal domain adjacent to the eastern boundary (SC, Fig. 2A). Within these marginal domains, a steep foliation is developed parallel to the axial planes of the tightening folds. Hornblende mineral lineations appear together with this steep foliation. This fabric cross-cuts the 126–120 Ma Western Fiordland Orthogneiss and has been dated as having evolved during the interval of 119–111 Ma (Marcotte et al., 2005). These relationships justify our assumption that the foliations and mineral lineations we measured formed during the same approximate time interval.
A comparison of structures in the marginal and central domains allowed us to determine the kinematics of flow within the Indecision Creek shear zone. The boundaries of the shear zone are well defined and nearly vertical (Fig. 8A). These boundaries separate the domains and parallel the steep western margin of the Darran igneous suite (Fig. 2A). With increasing strain across the two marginal domains (black dashed arrows in Figs. 2A and 2B), foliation planes steepen and rotate counter-clockwise into parallelism with the vertical boundaries. For a steep shear zone accommodating a component of horizontal shortening perpendicular to its boundaries (i.e., a horizontally thinning shear zone), the spread of foliation poles defines the VNS and allows estimation of the vorticity vector \( \omega \) (Lin et al., 1998; Jiang et al., 2001; Jones et al., 2004). This relationship is true for triclinic and monoclinic zones of steady-state deformation and is independent of any volume change. Each of the two marginal domains shows a spread of poles to foliation that define two gently-dipping planes on an equal-area stereoplot (Fig. 8A). These two dipping planes correspond to the western and eastern marginal domains, respectively. The intersection of the VNS and the shear zone boundary gives nearly horizontal bulk shear directions (Jiang et al., 2001). The counter-clockwise sense of rotation indicates a component of sinistral strike-slip displacement parallel to the shear direction.

The hornblende mineral lineations change orientation from a few degrees from horizontal in the marginal domains to near vertical in the central domain (Fig. 8A). These lineations neither perfectly parallel the vorticity vector \( \omega \) nor lie in the VNS. These patterns indicate that the deformation was dominated by pure shear because the farther the lineations plot away from the shear direction, the higher the percentage of pure shear relative to simple shear (Lin et al., 1998; Jiang et al., 2001). This pattern also implies that the marginal domains record a higher percentage of simple shear than the central domains because the lineations plot closer to the VNS in the latter. This result means that the margins of the shear zone were more efficient at maximizing offset relative to strain. It also indicates that the deformation deviated from plane strain and may have involved a significant component of sub-horizontal (layer-parallel) flattening, especially in the shear zone center. The migration of lineations toward the vertical dip-line of the shear zone with increasing strain (black arrows in...
Fig. 5B) indicates that the zone was thickening in a down-dip or vertical direction. These patterns are diagnostic of triclinic transpression and are independent of any change in volume (Lin et al., 1998; Jiang et al., 2001; Czeck and Hudleston, 2003).

These results are in excellent agreement with the results from Pembroke Valley and also are compatible with interpretations of transpression. To test the interpretation and the assumption of steady-state flow, we compared our data to forward models of homogeneous triclinic deformation (original models presented by Lin et al., 1998, and Jiang et al., 2001). The models predict the progressive changes in the orientation of finite strain axes for various flow types. One model that approximates the patterns illustrated by the Indecision Creek shear zone involves components of horizontal shortening, vertical thickening and subhorizontal sinistral shear parallel to a steep shear zone boundary (Fig. 8B) (after figure 5b of Jiang et al., 2001). Nevertheless, the natural patterns display significant deviations from the model results. First, structural patterns within Fiordland indicate that the deformation was far from homogeneous. Structures are complex in part due to the heterogeneity of the strain and also due to compositional variations created by numerous superposed dikes and sills that form part of the shear zone fabric. Second, the spread of poles to foliations is greater than the predicted ≤ 45° for transpression. The large spread we measured probably reflects folding and flattening within the marginal domain. Nevertheless, despite these complexities, the patterns we observed are compatible with deformation with components of vertical thickening, subhorizontal, arc-normal shortening, and sinistral arc-parallel translation (Fig. 8C). The result of this comparison suggests that, although the deformation probably was not truly steady state, the approximation is good enough to allow us to recognize the bulk kinematics of flow at the regional scale.

CENTRAL FIORDLAND

Geometry, Age, and Sequence of Structures

In Central Fiordland (Fig. 2D), the Doubtful Sound shear zone is composed of splays of upper-amphibolite-facies mylonite up to several hundred meters thick (Figs. 2E and 2F). The shear zone was originally interpreted as a ductile thrust fault by Oliver and Coggon (1979) and Oliver (1980) and later was reinterpreted as an extensional detachment fault by Gibson et al. (1988). Zones of high strain are mostly sub-horizontal with anastomosing branches that dip gently to the northeast and southwest (Fig. 9A). Curved zones of mylonite envelop asymmetric pods that preserve older gneissic fabrics and garnet-granulite mineral assemblages. These pods typically vary in size from <1 m to >30 m in horizontal length. One of the largest high strain zones separates two formations of amphibolite and granitoids that comprise an overlying Paleozoic cover sequence. North of Doubtful Sound the cover sequence mostly dips gently to the north and northeast. Upper-amphibolite-facies mylonite zones up to 500 m thick also cut deeply inside the batholith (Fig. 2F). Foliations are defined by retrogressive biotite and hornblende that formed from the hydration of pyroxene. The branching nature of high strain zones in dioritic gneiss is similar to shear zones exposed in Pembroke Valley.

In the foot wall of the Doubtful Sound shear zone at Crooked Arm (CA10, Fig. 2D) strands of upper-amphibolite-facies mylonite (Figs. 9C and 9D) cut obliquely across an older granulite-facies fabric (Oliver, 1977; Oliver and Coggon, 1979; Klepeis et al., 2007). In these zones, garnet-granulite-facies foliations are transposed parallel to upper-amphibolite-facies foliation planes (Figs. 9C and 9D). Penetrative northeast-plunging mineral lineations defined by aggregates of plagioclase, pyroxene and amphibole are well developed on foliation planes in zones of both granulite-facies mylonite (L\text{\text{gg}}-S\text{\text{gg}}, Fig. 9D) and zones of upper-amphibolite-facies mylonite (L\text{\text{gg}}-S\text{\text{gg}}, Fig. 9E). The regional-scale geometry of these two fabrics, including sense-of-shear indicators, is similar (Klepeis et al., 2007).

Hollis et al. (2004) obtained U-Pb ages on zircon (Fig. 2D; Table 1) from Crooked Arm that suggest granulite-facies metamorphism occurred at 114 ± 2.2 Ma in this region. This age is in agreement with the ages of the youngest part of the Western Fiordland Orthogneiss (116–113 Ma; Table 1) obtained by Tulloch and Kimbrough (2003) and Hollis et al. (2004). Klepeis et al. (2007) obtained U-Pb ages on zircon from syntectonic dikes that indicate deformation in the shear zone occurred through 102 ± 1.8 Ma. This age determination is consistent with U-Pb thermochronology on titanite obtained by Flowers et al. (2005) from the same area (Table 1). These latter ages indicate cooling through 650–550 °C during the interval 113.4–111 Ma.

Microstructures and Changing Conditions of Deformation

Microstructures in the superposed fabrics from the hanging wall and foot wall of the Doubtful Sound shear zone provide a record of changing conditions within the lower crust, including its exhumation, during the period 114–90 Ma. Outside (above and below) the Doubtful Sound shear zone, grain sizes in undeformed diorite of the Western Fiordland Orthogneiss are relatively large, with most plagioclase grains falling within the range of 200 μm–1 mm. A lattice preferred orientation (LPO) in plagioclase is visible. Plagioclase grains display deformation twins that taper toward grain boundaries and grain boundary bulging is commonly observed (Fig. 10A). Hornblende and biotite are clustered into aggregates that define a weak foliation (Fig. 10E) interpreted to reflect flow under partially molten conditions.

Within one kilometer of the Doubtful Sound shear zone, in its foot wall at Crooked Arm (CA10, Fig. 2D), garnet-granulite mineral assemblages record dehydration of host rock assemblages at temperatures of >700 °C and pressures of 12 kb (Oliver, 1977; Gibson and Ireland, 1995; Hollis et al., 2004). These exposures of sheared garnet granulite within Crooked Arm are composed of plagioclase, clinopyroxene, orthopyroxene, and garnet with
Figure 9. (A) Vertical profile of the Doubtful Sound shear zone exposed at Joseph Point. Note gently-dipping, oblate-shaped lozenges defined by foliation ($S_{sz}$). (B) Photograph of a mylonitic part of the Doubtful Sound shear zone. Light colored layers contain elongate feldspar porphyroclasts in a fine-grained biotite- and hornblende-bearing matrix. These layers could represent deformed dikes in a highly sheared part of the Western Fiordland Orthogneiss. (C) Two vertical profiles of a high-strain branch of the Doubtful Sound shear zone at Crooked Arm. High strain zone is defined by a penetrative upper-amphibolite-facies foliation ($S_{SSZ}$) that transposes a garnet-granulite-facies foliation ($S_{GG}$) in the foot wall and hanging wall (indicated by thick rectangles). (D) Equal area, lower hemisphere, stereographic projections showing orientation of garnet-granulite-facies foliations and plagioclase + hornblende mineral lineations at Crooked Arm. (E) Stereoplot showing orientation of upper-amphibolite-facies foliations and plagioclase + hornblende mineral lineations ($L_{SSZ}$) within the Doubtful Sound shear zone. (F) Sketch showing predicted geometry of field of back rotation in a vertically thinning, horizontally stretching shear zone. (G) Hyperbolic plot showing distribution of forward rotated and backward rotated feldspar clasts in the Doubtful Sound shear zone. Data show a vertically thinning flow regime with approximately equal components of pure shear and simple shear. Plot combines data from five sites (from Klepeis et al., 2007).
Figure 10. Photographs of microstructures from within the hanging wall and foot wall of the Doubtful Sound shear zone. All images are from thin sections of surfaces oriented perpendicular to foliation and parallel to lineation. Foliation planes in all images parallel the bottoms of the images unless otherwise indicated. (A) Photomicrographs of the dioritic part of the Western Fiordland Orthogneiss from the hanging wall of the Doubtful Sound shear zone at Bradshaw Sound. Image (crossed polars) shows plagioclase deformation twins as well as evidence for grain boundary bulging and grain boundary migration in plagioclase (arrows). (B) Photomicrograph of granulite-facies fabric in the foot wall of the Doubtful Sound shear zone in Crooked Arm. At the bottom of this image (crossed polars) is a deformed aggregate of clinopyroxene, orthopyroxene, garnet, and quartz. The plagioclase in the center of the image is almost completely recrystallized. (C) Photomicrograph (crossed polars) of upper-amphibolite-facies foliation within the Doubtful Sound shear zone at Joseph Point. Image shows an asymmetric hornblende fish (top-to-left, or -northeast, shear sense) surrounded by a matrix of dynamically recrystallized plagioclase. Grain boundary migration leading to recrystallization occurs along plagioclase grain boundaries. (D) Photomicrograph (crossed polars) showing subgrain formation within a plagioclase grain and recrystallization along a plagioclase grain boundary. Compare to Figure 10B. (E) Scanned image of a thin section (45 cm long) showing the macroscopic texture of dioritic orthogneiss (Western Fiordland Orthogneiss) below the Doubtful Sound shear zone. Note the relatively large grain size, weak foliation, and clustering of hornblende and biotite. (F) Scanned image of thin section at the same scale as in Figure 10E showing texture of dioritic orthogneiss (Western Fiordland Orthogneiss) within the shear zone. Note the extreme grain-size reduction and the anti-clustered distribution of hornblende and biotite.
small amounts of quartz and epidote. In most places within the samples, feldspar is completely recrystallized with sizes ranging from 100 to 500 µm in diameter (Fig. 10B). Recrystallization occurred through both grain boundary migration and subgrain rotation. A strong LPO is present. Many plagioclase grains intersect at 120° triple junctions. No growth twins were observed in plagioclase grains, but flame perthite and other deformation twin textures are common. A shape preferred orientation (SPO) in plagioclase defines a weak foliation oblique to the dominant foliation (S_{oo}). Orthopyroxene and clinopyroxene grains form asymmetric fish within a plagioclase matrix. These sense-of-shear indicators show both top-to-the-northeast (Fig. 10C) and -southwest senses of shear similar to shear indicators in the upper-amphibolite-facies strands of the Doubtful Sound shear zone.

Areas of the Western Fiordland Orthogneiss that are deformed by the Doubtful Sound shear zone display mineral assemblages that include plagioclase, hornblende, clinzoisite, K-feldspar and biotite with minor quartz and clinopyroxene. The composition is similar to that of samples outside the shear zone except there is a much greater percentage of hydrous phases. Plagioclase grain sizes are smaller within the shear zone than outside (Fig. 10F) with most grains only 50–100 µm and some up to 1 mm in diameter. Plagioclase shows exsolution lamellae and a well-developed SPO aligned with a penetrative foliation (S_{oo}). The LPO appears visually weaker than outside the shear zone. Biotite and hornblende are aligned within the foliation. Plagioclase shows recrystallization through grain boundary migration (Fig. 10D). Klepeis et al. (2007) report deformation at upper-amphibolite-facies conditions of 550–650 °C and 7–9 kb.

On the basis of these microstructures, we interpret the relative importance of different deformation mechanisms as temperature and fluid conditions changed in the lower crust. Samples from outside the Doubtful Sound shear zone display features expected to result from dislocation creep in plagioclase. However, not all grains were recrystallized in these samples as evidenced by the presence of growth twins. This may suggest that high-grade metamorphic conditions did not last long enough to cause complete recrystallization everywhere within the Western Fiordland Orthogneiss. In addition, triple junctions of plagioclase grains are evidence for annealing. This process appears to have occurred after deformation through dynamic recrystallization ceased but while temperatures remained high. This pattern of a relatively short residence time at depth is consistent with evidence of rapid isobaric cooling during the interval 114–111 Ma (Flowers et al. 2005).

Sheared samples of garnet granulite from Crooked Arm display evidence for dislocation creep and recrystallization through grain boundary migration and subgrain rotation. The presence of a LPO suggests that diffusion creep was not an important process either during or after the time that dislocation creep was active. In contrast, samples from within the Doubtful Sound shear zone display textural evidence of diffusion creep assisted by the presence of hydrous phases. The LPO within these deformed samples is weaker than in samples outside the shear zone. Diffusion creep does not produce a LPO and it can weaken a preexisting LPO. Biotite and hornblende grains form cuspate intergrowths, with plagioclase parallel to the shear zone foliation. These patterns of the weak phases within a deformed sample are indicative of the activity of diffusion-creep-accommodated grain-boundary sliding (Gower and Simpson, 1992; Kruse and Stunitz, 1999). This evidence strongly suggests that fluids played an important role in controlling deformation mechanisms within the amphibolite-facies Doubtful Sound shear zone exposures. The activity of diffusion creep also may have been facilitated by the smaller grain size that resulted from recrystallization through dislocation creep. These features indicate that changes in temperature and fluid conditions occurred over a short (3–4 Ma) period of time after extension initiated. During the period 114–111 Ma, temperatures dropped from >800 °C to 650–550 °C and the availability and importance of water in deformation became greater. The increased availability of water after ca. 111 Ma appears to have promoted a localization of strain into narrow (<1 km) zones during extension.

Strain Patterns and Estimates of Kinematic Vorticity

Klepeis et al. (2007) used measurements of shear zone boundaries, stretching lineations, different types of shear indicators, and the X-direction of finite strain ellipsoids to identify the VNS for both granulite-facies (L_{oo}-S_{oo}) and upper-amphibolite-facies (L_{oo}-S_{oo}) fabrics. Foliation and mineral stretching lineations in these superposed fabrics are nearly parallel and the VNS in both cases is approximately vertical (Figs. 9D and 9E). The bulk shear direction parallels the intersection between the shear zone boundary and the VNS. The parallelism among this bulk shear direction, mineral lineations and X-directions of 3-D finite strain calculations reported by Klepeis et al. (2007) indicate compatibility among the observations. In general, their data suggest that although both oblate and prolate strain ellipsoid shapes are represented, the bulk deformation involved mostly flattening strains.

Sense-of-shear indicators in the Doubtful Sound shear zone are best preserved in zones of upper-amphibolite-facies mylonite. In these areas, asymmetric tails of amphibole and quartz around plagioclase grains, hornblende and biotite fish (Fig. 10C), S-C fabric, and asymmetric tails around rotated garnet porphyroblasts yield top-down-to-the-northeast and -southwest senses of shear. Asymmetries associated with these kinematic indicators are best developed on surfaces oriented perpendicular to foliation and parallel to northeast-plunging stretching lineations. Rotated porphyroclast systems show σ-type, δ-type, and β-type tails that can be used to quantify the ductile flow field in the shear zone (Fig. 9F). Klepeis et al. (2007) obtained estimates of sectional kinematic vorticity (W_{oo}) for the vertical plane by applying the porphyroclast hyperbolic distribution method of Simpson and DePaor (1993, 1997) in four sites within Crooked Arm. At each site over 30 clasts were measured at multiple length scales (thin section to a 10–20 m² outcrop) to obtain spatially averaged estimates of kinematic vorticity. For each shear zone the range of
DISCUSSION

The Origin of Heterogeneity in the Middle and Lower Crust

Our analysis of superposed shear zones in Northern Fiordland indicates that most of the kinematic parameters used to describe ductile flow in the middle and lower crust changed repeatedly during a short 8–10 Ma time interval in the Cretaceous. Shear zone boundaries alternated between gently dipping (layer-parallel) and sub-vertical (layer-perpendicular) (Fig. 11A). The degree of non-coaxiality, kinematic partitioning and strain symmetry also were highly variable. Some of the spatial variability and transience of the deformations appears to reflect changes in the location and geometry of strength contrasts as the temperature, structure, and composition of the lower crust changed. For example, the cooling and crystallization of the Western Fiordland Orthogneiss batholith at ca. 120 Ma was accompanied by a change in the style of deformation in the lower crust. Deformation that was focused along the contacts of the batholith and lower crust cooled. Microstructures from the Doubtful Sound region indicate that the presence of magma and/or partial melt, temperature and fluid availability strongly influenced patterns of strain partitioning and the mechanisms that accommodated ductile flow. During the period 114–111 Ma, as extensional deformation became dominant, temperatures dropped from >800 °C to 650–550 °C and the availability and importance of water in deformation became greater. The increased availability of water appears to have controlled the localization of strain into narrow (<1 km) zones during extension. These changes strongly suggest that flow in the lower crust was intrinsically unsteady and non-uniform and strongly influenced by local boundary conditions. The variability in kinematic patterns also indicates that shear zone orientation and sense of shear alone are not diagnostic of tectonic regime.

Despite this high degree of variability, the data also show that some patterns remained constant through time. All of the lower-crustal deformations in Northern Fiordland record bulk horizontal (layer-parallel) shortening and vertical (layer-perpendicular) thickening (Fig. 11A). This result is independent of both the scale of observation and the orientation of the shear zone boundaries. In addition, all deformations involved between 50% and 90% pure shear (see Tikoff and Fossen, 1995, and Bailey et al., 2007 for discussion of general shear), in a manner consistent with a regime of lithospheric thickening and contraction. These results contrast with the deformations in Central Fiordland, which record flow involving vertical thinning, subhorizontal stretching and 40%–50% pure shear (Fig. 11B). The patterns suggest that transient, non-uniform flow histories can be assessed by comparing data from successive and sufficiently small time intervals and mesoscale observations following the approach we have outlined in this chapter.

The results of our analyses also demonstrate when and how patterns of lower-crustal flow changed during the transition from crustal thickening to crustal extension at about ca. 114 (and before ca. 111 Ma). During this transitional period, shear zones in northern Fiordland were characterized by diffuse, vertically thickening zones recording between 50% and 90% pure shear deformation. In contrast, lower-crustal shear zones that formed in Central Fiordland record focused, vertically thinning, horizontally stretching flows with approximately equal components of pure and simple shear. These patterns correspond to the stages of regional contraction and regional extension, respectively. Within the limits of precision in available ages, the timing of this transition appears diachronous. Our analysis of kinematic patterns in the lower crust combined with previously published geochronology indicate that the shift to extension in the Doubtful Sound region occurred by ca. 114 Ma, significantly earlier than implied by previous studies (ca. 108 Ma according to Gibson et al., 1988). In Northern Fiordland, the timing of the shift from regional contraction to regional extension is uncertain because no major extensional shear zones occur in this area. The limits of uncertainty on age determinations obtained from syn-tectonic dikes in the Indecision Creek shear zone indicate that contraction in the Milford region occurred during the interval 119–111 Ma (Marcotte et al., 2005).

Length and Time Scales of Changing Flow Patterns during the Transition from Contraction to Extension

A comparison of metamorphic and geochronologic data from the northern and central parts of Fiordland indicate that the lower crust of this orogen was characterized by extremely heterogeneous and transient thermal structures on length scales of less than 100 km. U-Pb geochronology indicates that the lower crustal section at Milford Sound cooled from >800 °C at ca. 120 Ma to ~650 °C by ca. 116 Ma (Klepeis et al., 2004), while remaining at...
Figure 11. Diagram showing the diachronous evolution of lower-crustal structures in Northern (A) and Central (B) Fiordland. Upper diagram shows the evolution of contractional fabrics in Northern Fiordland from magmatic flow to high-temperature deformation at the garnet-granulite facies to cooler deformation at the upper-amphibolite facies. This section records contractional deformation and magmatism in the lower crust until at least ca. 111 Ma. (B) Evolution of extensional fabrics in Central Fiordland from magmatic flow to high-temperature deformation at the garnet-granulite facies to cooler deformation at the upper-amphibolite facies. This section records the initiation of extension at ca. 114 Ma. The block labeled ‘Tectonic Shift’ represents the time the tectonic regime changed from regional contraction to regional extension. This age is taken to be ca. 114 Ma but uncertainties in published ages indicate the shift could have occurred anytime during the interval 114–111 Ma. Abbreviations are as follows: CS—Caswell Sound; ICSZ—Indecision Creek shear zone; MD—Mount Daniel; PV—Pembroke Valley; SZB—shear zone boundary; VNS—vorticity normal section (for location see Fig. 2). See text for discussion.
high pressures (12–14 kb). Less than 100 km south of Milford Sound, however, U-Pb geochronology indicates that peak conditions involving temperatures of >800 °C persisted at Doubtful Sound until ca. 114 Ma (Table 1; Klepeis et al., 2007). The length and time scales of this variability imply that any single locality provides a part of the history of the lower crust that may not be representative of the whole section. It also suggests that elements of both of the hypotheses outlined earlier may explain how and why extension initiated following a period of crustal thickening.

In contrast to the first hypothesis, the geologic history of the Milford Sound region indicates that weakening of lower-crustal rocks by heating, magmatism and partial melting does not always promote horizontal flow that leads to crustal thinning and lithospheric extension. In this region, the peak of granulite-facies metamorphism in the lower crust at ca. 120 Ma was accompanied by magmatism and partial melting but did not develop thinning flows nor major extensional shear zones. Instead the data clearly show that, in this area, these events were accompanied by contractional deformation. However, the history of deformation in Central Fiordland partially supports the first hypothesis. This latter region record the development of major extensional shear zones that, at least initially, localized into areas of the lower crust that were weakened by magma and heat at ca. 114 Ma. These relationships indicate that a prerequisite for localizing extension in the lower crust is crustal weakening but that the overall style of the flow is controlled by the tectonic regime.

Elements of the second hypothesis, where a strong lower crust and a weak lower crust controlled extension, also appear to apply to Fiordland. Although extensional deformation in Central Fiordland was localized into areas that were weakened by magma and heat, this situation was transient. These hot, weak zones had declined and the lower crust had cooled through 650–550 °C by 113.5–111 Ma (Flowers et al., 2005). After ca. 111 Ma, extension was localized into the middle crust where quartz- and mica-dominated lithologies appear to have localized deformation (Klepeis et al., 2007). This result may explain why Northern Fiordland, which preserves the lower but not the middle crust, lacks evidence of extension; extension preferentially was partitioned into a weak middle crust after ca. 111 Ma. This process, whereby the transient nature of magma-induced hot spots in the deep crust controls strain partitioning, may help explain diachronous patterns of deformation and along-strike variability in arc structure, including the development of highly localized zones of extension and contraction.

CONCLUSIONS

Kinematic analyses conducted on superposed shear zones at multiple scales in Fiordland reveal extremely heterogeneous patterns of ductile flow and strain partitioning in the middle and lower crust. In Northern Fiordland, most of the parameters used to describe ductile deformation in shear zones were highly variable as both local conditions and plate motions changed. Over an ca. 8 Ma period (119–111 Ma), kinematic patterns were strongly influenced by local variations in crustal structure, magmatic activity, composition, temperature, and rheology. In Central Fiordland, microstructures indicate that the grain-scale mechanisms that accommodated ductile flow also changed rapidly over a short 3–4 Ma interval as crustal compositions and fluid availability changed. The results imply that over periods of several million years flow in the lower crust of orogens is characterized by non-uniform, non-steady flow fields that reflect changing local boundary conditions.

The results also indicate that some kinematic parameters in lower-crustal shear zones were sensitive to changes in far-field plate boundary dynamics. All shear zones that formed during the 119–111 Ma period in Northern Fiordland record horizontal (layer-parallel) shortening and vertical (layer-perpendicular) thickening with 50%–90% pure shear. In contrast, all shear zones that formed during the 114–90 Ma period in Central Fiordland record vertical thinning and sub-horizontal stretching, with approximately equal amounts of pure shear and simple shear. These patterns are correlative with regional contraction and regional extension, respectively, and suggest that the transition to extension may have been diachronous over length scales of less than 100 km within the limits of precision in available ages. They also indicate that the orientations of shear zones and sense-of-shear indicators alone cannot be used to uniquely define different tectonic regimes in the middle and lower crust. A careful choice of length and time scales is important for interpreting the significance of deformation patterns in ancient orogenic belts where plate boundary conditions may be poorly defined. In New Zealand, these scales are ~100 km and ca. 10 Ma, respectively.

Areas of discord between published models of extension and observations in Fiordland center on the length and time scales of hot, weak zones in the lower crust and how such zones influenced the development of extensional structures. Metamorphic and geochronologic data collected from near Milford Sound suggest that part of the lower crust experienced rapid cooling from temperatures of between 850 °C and 750 °C to 700–650 °C by ca. 116 Ma following a period of mafic-intermediate magmatism and lower-crustal melting at ca. 120 Ma. This episode of lower-crustal cooling predated the tectonic shift from contraction to extension. However, less than 100 km to the south of Milford Sound, U-Pb zircon ages and metamorphic data suggest that magmatic activity there occurred later and that lower-crustal temperatures remained high (>800 °C) until ca. 114 Ma. At this latter locality, the lower crust was still hot when the shift to extension occurred. These observations indicate that the thermal structure and rheological transitions linked to magmatism and the partial melting of lower crust were spatially very heterogeneous and transient. Localized hot spots in the lower crust created by heterogeneous patterns of magmatic activity affected deformation patterns during cycles of extension and contraction within the arc. When the regional tectonic regime shifted at ca. 114 Ma, large (1-km-thick) extensional shear zones preferentially developed in hot, weak zones that thinned the lower crust. In contrast, areas of the lower crust that were relatively cool
at the time of the shift did not develop extensional structures and preserve older contractional shear zones. This example illustrates how the development of localized hot spots due to magmatism and metamorphism in the lower crust strongly influences deformation partitioning within arc crust.

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