STRUCTURAL AND KINEMATIC EVOLUTION OF THE LOWER CRUST
DURING CONTINENTAL EXTENSION: THE RESOLUTION ISLAND SHEAR
ZONE, FIORDLAND, NEW ZEALAND

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Abstract

Three dimensional finite strain and kinematic data from the Resolution Island Shear Zone, Fiordland, New Zealand record the progressive evolution of a lower crustal metamorphic core complex. The Resolution Island Shear Zone is a mid-Cretaceous (~114-90 Ma) extensional shear zone that juxtaposes high-pressure ($P\sim$17-19 kbar) garnet-granulite and eclogite facies orthogneiss from the lower crust against mid-crustal ($P\sim$6-8 kbar) orthogneiss and paragneiss along a low-angle upper amphibolite facies ductile normal fault.

In the lower plate of the Resolution Island Shear Zone the high-pressure garnet-granulite and eclogite facies gneissic foliations ($S_1$) are attenuated by granulite facies extensional shear zone foliations ($S_2$). Retrograde metamorphism marked by the breakdown of omphacite and garnet to amphibole and feldspar in $S_2$ foliation records the unloading of the lower plate during extension. Continued extension localized strain into weaker amphibole and feldspar-bearing lithologies. Upper amphibolite facies shear zones anastomose around rigid lenses that preserve the $S_1$ and $S_2$ fabric. Upper amphibolite facies shear zone fabrics ($S_3/L_3$) that envelop these pods display a regional-scale dome-and-basin pattern. These shear zones coalesce and form the Resolution Island Shear Zone. Coeval with the formation of the Resolution Island Shear Zone, a conjugate, southwest dipping, and lesser magnitude shear zone termed the Wet Jacket Shear Zone developed in the upper plate of the Resolution Island Shear Zone.

Three-dimensional strain analyses from $S_3/L_3$ fabric in the Resolution Island Shear Zone show prolate-shaped strain ellipsoids. Stretching axes (X) from measured finite strain ellipsoids trend northeast and southwest and are subparallel to $L_3$ mineral stretching lineations. Shortening axes (Y, Z) are subhorizontal and subvertical, respectively, and rotate through the YZ plane of the finite strain ellipsoid. This pattern reflects the dome-and-basin geometry displayed by anastomosing $S_3$ foliations and indicates the Resolution Island Shear Zone developed in the field of constriction. Three-dimensional kinematic results indicate a coaxial-dominated rotation of stretching lineations toward the X-axis in both the XZ and XY planes of the finite strain ellipsoid. Results suggest that a lower crustal metamorphic core complex developed in a constrictional strain field with components of coaxial-dominated subvertical and subhorizontal shortening.

Mid-Cretaceous (~114-90 Ma) extensional structures exposed in Fiordland, including the Resolution Island, Wet Jacket, Mount Irene and Doubtful Sound shear zones and the Paparoa metamorphic core complex allows the reconstruction of a crustal column that describes the geometry of mid-Cretaceous continental rifting of Gondwana. The overall symmetry of crustal-scale structures during continental extension suggests kinematic links between flow in the lower crust and the geometry and mode of continental extension. This result is consistent with numerical models of lithospheric rifting that predict the lower crust has a primary control on the style of continental extension.
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Chapter I: Introduction

Several factors control deformation during continental rifting including the inherited structure and rheologic stratification of the lithosphere (Buck, 1991; Hopper & Buck, 1996; Nagel & Buck, 2004), thermal conditions (Wijns et al., 2005), plate velocities and trajectories (Bennett et al., 2003), strain rate (Davis & Kusznir, 2002), and presence or absence of magma. Lithospheric models of continental extension show that among these variables the ability of the lower crust to flow has an important control on the mode and geometry of lithospheric extension (Buck, 1991; Hopper & Buck, 1998; Huismans & Beaumont, 2007; Gessner et al., 2007). Some numerical models suggest that the relative strength of the lower crust to the lithospheric mantle and upper crust controls the mode of continental extension (Buck, 1991; Hopper & Buck, 1996; Wijns et al., 2005), the symmetry or asymmetry of rifted margins (Huismans & Beaumont, 2003; 2007; Gessner et al., 2007), the development of metamorphic core complexes (Hopper & Buck 1996; Wijns et al., 2005), coupling or decoupling of the mantle and crust (Hopper & Buck, 1998), and the width of a rift zone (Behn et al., 2002). Although the effects of lower crustal strength on rifted continental lithosphere are well modeled, the kinematics and tectonic significance of flow in the lower crust remain poorly understood.

Fiordland, New Zealand preserves excellent exposures of middle and lower crustal extensional structures that formed during mid-Cretaceous (~114-90) continental rifting Gondwana (Gibson et al., 1988; Tulloch & Kimbrough, 1989; Klepeis et al., 2007). The recent discovery of deep (P~17-19 kbar; De Paoli et al., 2007) granulite and
eclogite exposures in the footwall of the Resolution Island Shear Zone indicates that the Resolution Island Shear Zone formed in the lower crust (~61-69 km depth) during the mid-Cretaceous rifting of Gondwana. This result makes the Resolution Island Shear Zones the lowermost of a suite of coeval, crustal scale extensional structures exposed in Fiordland, New Zealand (Gibson et al., 1988; Tulloch & Kimbrough, 1989; Spell et al., 2000; Scott & Cooper, 2006; Klepeis et al., 2007; Kula et al., 2007), providing an ideal setting for studying the influence of lower crustal flow on the geometry of deformation during continental extension.

In this thesis, I present field-based structural, strain, and kinematic data from the Resolution Island Shear Zone in central Fiordland, New Zealand to describe the evolution of a lower crustal extensional shear zone. I then compare this shear zone to other extensional shear zones in Fiordland to assess the geometry of deformation in the crust during continental extension. With these goals in mind I use a field-based approach to address several fundamental questions about continental extension including: 1) How do large-scale extensional shear zones evolve in the lower crust? 2) What is the three-dimensional geometry of deformation in extending lower crust? 3) Is deformation in the lower crust kinematically linked to deformation at middle and upper crustal levels, as some models predict? 4) What processes result in the formation of metamorphic core complexes during continental extension?

The results of this research have implications for understanding the evolution of large-magnitude normal faults that form metamorphic core complexes. The Resolution Island Shear Zone displaces eclogite and garnet-granulite facies rocks in the lower plate
against upper amphibolite facies metasedimentary and metavolcanic mid-crustal rocks in the upper plate along a low-angle detachment fault. Field observations document the occurrence of interconnected anastomosing amphibolite facies shear zones in the lower plate that coalesce into the Resolution Island Shear Zone. These shear zones form dome-and-basin patterns with synthetic-antithetic pairs, suggesting that both top-down-to-the northeast and top-down-to-the southwest displacements accommodated extensional uplift of the metamorphic core. Three-dimensional strain and kinematic data from the Resolution Island Shear Zone indicate that deformation occurred under the field of coaxial-dominated constriction. This style of deformation appears to be the driving mechanism for developing dome-shaped patterns in the Resolution Island Shear Zone and uplifting the lower plate along a large-magnitude normal fault.

A second interesting result of this study has implications for the structural evolution of Fiordland during mid-Cretaceous continental rifting and the geometry of rifted continental margins in general. Geochronology, thermobarometry and structural data from the Resolution Island Shear Zone and several other coeval extensional shear zones previously documented in western New Zealand are well constrained (Gibson et al., 1988; Tulloch & Kimbrough, 1989; White, 1994; Spell et al., 2000; Scott & Cooper, 2006; Klepeis et al., 2007; Kula et al., 2007). These data allowed me to reconstruct a model showing the geometry of deformation in the crust during the mid-Cretaceous continental rifting of Gondwana. The model indicates an overall symmetry of deformation in the crust. Klepeis et al. (2007) indicated that pure-shear dominated vertical shortening is an important driving mechanism for extension. Results from the
Resolution Island Shear Zone describe symmetric coaxial-dominated flow in the lower crust, suggesting that deformation in the lower crust is kinematically linked to deformation at middle and upper crustal levels. This interpretation supports some numerical models of continental extension that predict that the ability of the lower crust to flow has a significant control on the geometry of deformation in the lithosphere during continental extension (Buck, 1991; Huismans & Beaumont, 2007; Gessner et al., 2007).
Chapter II: Analytical methods

1. Field Methods

Data for this project were collected during one field season from Resolution Island, Acheron Passage, and Wet Jacket Arm in western Fiordland, New Zealand. These localities were selected because of their proximity to known structures including the Resolution Island Shear Zone, a newly discovered major extensional shear zone. Geological relationships are well exposed by wave-washed shoreline outcrops. The field excursion was conducted using an inflatable boat as a means of transportation to shoreline outcrops. Data were collected from 41 stations interspaced along the shorelines (Plate 1).

Field methods at each station included: identifying the lithologies present, measuring the attitudes of foliations, lineations, fault planes, and other structures, identifying a sequence of fabrics, identifying kinematic indicators, and collecting samples for petrographic, kinematic, and strain analyses. Field notes also included creating detailed sketches, maps, and cross sections of geological relationships.

Data collected from this field excursion resulted in a geologic map and cross section of the field area, the delineation of a sequence of geological fabrics, and a quantitative assessment of flow patterns related to the Resolution Island Shear Zone. Results are discussed in detail in the main section of my thesis (Chapter IV).
2. Laboratory Methods

2.1 Microstructural analysis

To constrain the relative timing, and sequence of metamorphic and deformational events I described grain-scale structures from 17 samples in the lower plate of the Resolution Island Shear Zone. Deformation and metamorphic textures in feldspar, hornblende and garnet phases from each of these samples allowed me to determine which of the mineral phases best records strain, define a temperature range for deformation, interpret sense of shear, and describe metamorphism that occurred during deformation.

Microstructures in deformed mineral phases that grew before or during deformation are used as kinematic indicators. In thin sections cut from surfaces oriented perpendicular to foliation and parallel to the mineral stretching lineation, kinematic indicators tell the sense of shear and rotation in a shear zone and include: asymmetric tails on rotated porphyroblasts, asymmetric shear band foliations, preferred orientations of elongate minerals, asymmetry of porphyroblasts (Passchier & Trouw, 2005). I describe kinematic indicators in the Resolution Island Shear Zone to determine the sense of shear, symmetry or asymmetry of deformation, and the bulk flow regime.

Metamorphic reactions that are coeval with deformation and do not run to completion commonly preserve replacement textures or reactions rims on mineral phases within a deformed rock. These textures include pseudomorphs, symplectite rims and coronas around porphyroclasts. Pseudomorph textures occur when one mineral phase is completely replaced by a new mineral phase during changing pressure and temperature conditions. Symplectite rims and coronas occur along the grain boundaries of an altering
mineral grain. These textures record the progress, and direction of metamorphic reactions. I described metamorphic replacement textures including symplectites, coronas, and pseudomorphs to determine the pressure and temperature conditions during deformation from samples in the lower plate of the Resolution Island Shear Zone. Results are presented in Chapter IV.

Feldspar displays a range of deformation textures that record changing deformation conditions. Describing deformation textures displayed by feldspar constrains a temperature range on the rock during deformation, or suggests that these temperatures were exceeded and that strain in the feldspar was reset (Passchier & Trouw, 2005). Deformation textures in feldspar are particularly important for describing the conditions of deformation within the feldspar-dominated lithologies from the lower crust (Kohlstedt et al., 1995).

Below ~400°C feldspar deforms by brittle fracture and cataclastic flow. Textures associated with brittle fracture can be easily assessed in thin section by the presence of poorly sorted, angular fragment in a matrix of fine grain pulverized material (Passchier & Trouw, 2005). With increasing depth in continental lithosphere an ~25°C/km geothermal gradient will generate temperatures in excess of ~400°C, that corresponds with the brittle-ductile transition (Kohlstedt et al., 1995). Under these higher temperature conditions strain in feldspar-bearing lithologies can be accommodated by the migration of dislocations through the crystal lattice mineral grains (Passchier & Trouw, 2005).

At temperatures >400°C for feldspar-dominated lithologies, dislocation creep processes in the crystal lattice of feldspar grains can accommodate strain. Deformed
feldspar grains display textures representative of different types of dislocation formation processes that occur in a variety of temperature ranges at constant strain rates. Between ~400°-500°C feldspar deforms by grain-scale processes including microfracturing, grain boundary bulging, and dislocation glide. These processes produce textures in feldspar grains including undulose extinction, deformation bands, twinning, and kink bands. Microfracture textures including bookshelf fractures are also common (Tullis & Yund, 1985; Passchier & Trouw, 2005).

At increased temperatures (450°-600°C), and constant strain-rate, recrystallization can occur by the onset of dislocation climb. Grain boundary bulging occurs and can nucleate recrystallized subgrains, and feldspar subgrains may begin to form mantle textures and fracturing becomes less prominent. With continued temperature increases (>600°C) dislocation climb and glide become highly operative processes. Subgrain rotation and grain boundary bulging become fully dominant recrystallization processes. Under these conditions (T>600°C) at low pressures, recrystallized feldspar grains can be strain free and may display clear subgrain boundaries, uniform extinction and annealing. Grain boundary migration does not occur in feldspar until very high temperatures (>850°C) and in the presence of a melt phase. Feldspar at these extreme conditions will develop strain free grains by the diffusion of dislocations across grain boundaries and display cuspate-lobate grain boundary textures (Tullis & Yund, 1985; Passchier & Trouw, 2005).

2.2 Strain analyses

I used the Rf/∅/θ method (Lisle, 1985) to measure three-dimensional shape fabrics in six samples from the Resolution Island Shear Zone. Each sample was cut from
three perpendicular planes on surfaces oriented parallel to lineation and perpendicular to foliation (XZ plane), parallel to foliation (YX plane), and perpendicular to foliation and lineation (YZ plane). Planes cut in these orientations contain the X, Y, and Z-axes of the finite strain ellipsoid for the sample (Figure 1). I used feldspar as a strain marker because it was the most ubiquitous mineral present in all of the samples and appeared to accommodate the most strain. The results of strain analyses are presented in Chapter IV.

Two-dimensional strain ellipses were measured using a computer program called SAPE (Mulchrone, 2005). I developed a method of sample preparation and data collection that streamlines the process of obtaining large amounts of data by using this program. I measured aspect ratios from a total for 1,275 feldspar grains and used these values to calculate plane-strain magnitudes for each perpendicular section (XZ, XY, and YZ planes of each sample). Using plane-strain magnitudes from each section I calculated best-fit strain ellipsoids for each sample, after Owens (1984). I present a method for preparing samples and using these programs in Appendix A.

I used the magnitudes of principal axes (X, Y, Z) from a best-fit strain ellipsoid as a proxy to describe the three-dimensional shape of deformed fabrics from the Resolution Island Shear Zone. The relative magnitudes of the principal axes of finite strain ellipsoids describe the deformation field under which rocks were deformed. If X/Y > Y/Z then rocks were deformed under the field of apparent constriction (Figure 1). In this case X is an axis of finite extension and both Y and Z are axes of finite shortening. Conversely if X/Y < Y/Z the rocks were deformed in the field of apparent flattening, Z is the axis of finite shortening and Y and X are axes of finite extension. If X/Y = Y/Z then deformation
occurred under plane strain, implying there was no finite strain along Y. Results were plotted on a Flinn diagram with ln(X/Y) plotted along the y-axis, and ln(Y/Z) along the X-axis to determine whether samples were deformed in constriction or flattening fields (Figure 1) (Flinn 1962). The results of three dimensional strain analyses are discussed in detail in Chapter IV.
Figure 1 – Strain analysis illustrations. (a) Diagram showing the relationship between the aspect ratio \(R_i\) and orientation \(\theta\) of an undeformed strain marker and the aspect ratio \(R_f\) and orientation \(\varnothing\) of that marker deformed by a strain ellipse with an aspect ratio = \(R_s\), after Lisle (1985). (b) Diagram showing the principal axes X, Y, and Z of a deformed strain ellipsoid in the field of constriction. (c) Diagram showing expected strain facies in the fields of constriction, flattening and plane-strain, after Gapais et al. (1987).
2.3 The Isocon Method

Volume loss during deformation affects the shape of a finite strain ellipsoid by removing material along the XY plane during the formation of foliations. If volume is lost by processes including dissolution, pressure solution, veining, or partial melt, then the finite strain ellipsoid will preserve a shape that is biased toward the field of apparent flattening (Davis & Reynolds, 1984). I used the Isocon method (Grant, 1986) to quantify the amount of volume change associated with deformation in the Resolution Island Shear Zone and to test the possibility of volume change altering the shape of calculated finite strain ellipsoids.

The Isocon method is a graphical method that assumes no mobility of certain elements during deformation (i.e. Al₂O₃). Plotting the weight percent concentrations of major and trace elements allows a comparison of the relative change in concentrations from an undeformed sample (X-axis) and a deformed sample (Y-axis). If the assumption of no mobility of certain elements during deformation is true, than a best-fit line to these data points through the origin defines an isocon, or line of no mass change (Grant, 1986). The slope of this line is proportional to the amount of volume change according to the equation:

\[ \frac{V^a}{V^o} = \frac{M^a}{M^o} \cdot \frac{\rho^o}{\rho^a} \]

where V is volume, M is mass, \( \rho \) is density, ‘a’ denotes the altered sample and ‘o’ denotes the host sample unaltered by deformation. The mass ratio is the inverse slope of the isocon (Grant, 1986; Barnes et al., 2004). The results of volume change calculations
are discussed in Chapter IV.

2.4 Kinematic analysis and determination of bulk flow regime

I used a graphical method developed by Gapais et al. (1987) to describe the bulk kinematics of flow in, and below the Resolution Island Shear Zone (~15 km²). This method uses outcrop-scale shear zone descriptors including mineral stretching lineations, foliations, and planes representing shear zone boundaries. Each of these data points represents a variable that describes flow in the shear zone in an X-Y-Z reference frame from the finite strain ellipsoid. Stretching lineations record the shear direction (L), foliation planes represent the shearing plane and their poles (N) record rotations or folding of the shearing plane, and planes parallel to both L and N represent the shearing plane or shear zone boundaries. Poles to the shearing plane (M) track simultaneous rotations in the shear direction (L) and rotations of the shearing plane (N) and represent rotations of the shear zone boundaries.

To evaluate shear zone kinematics shear zone descriptors are plotted on a stereonet around the principal axes (X, Y, and Z) of the finite strain ellipsoid. The symmetry or asymmetry of these plots are compared to predictive diagrams provided by Gapais et al. (1987). Data are divided into groups according to shear sense: top-to-the-northeast shear zones and top-to-the-southwest shear zones. These plots are termed the L-diagram, N-diagram, and M-diagram, respectively. The view direction of each plot is rotated so that the X-axis of the finite strain ellipsoid is parallel to the east-west axis of the stereonet, and the Z-axis is parallel to the north-south axis of the stereonet. In the case of the Resolution Island Shear Zone, the X-axes of six finite strain ellipsoids are locally
parallel to the mineral stretching lineation, and the Z-axes of the strain ellipsoids are normal to foliation. This allowed me to orient the stereonet such that the mean mineral stretching lineations (L) was oriented plunging 00° toward 090°, and the mean pole to foliation (N) was oriented plunging 00° toward 000°.

Predictive diagrams for areas deformed by apparent constriction suggest that the L-diagram clusters symmetrically around X and approaches Y, and the N-diagram will form a girdle through the YZ-plane. For areas deformed by apparent flattening predictive diagrams suggest the L-diagram will form a wide girdle through the XY-plane and the N-diagram will cluster strongly around Z, and the M diagram with cluster strongly around Y (Gapais et al., 1987). The results of kinematic analyses of the Resolution Island Shear Zone are discussed in Chapter IV.
Chapter III: Literature Review

1. Tectonic History of Fiordland, New Zealand

The Alpine Fault is a major northeast-striking dextral strike-slip fault that forms the Australian-Pacific plate boundary. The Alpine Fault accommodates differential plate movements between the Pacific Plate which is subducting toward the west beneath the Australian Plate to north of New Zealand and the Australian plate which is subducting toward the east beneath the Pacific Plate to the south of New Zealand. Oblique convergence along the Alpine Fault has been active since ~25 Ma (Lamarche & Lebrun, 2000; Lebrun et al., 2003). This convergence combined with erosion exhumed deep levels of the crust in the Fiordland region of New Zealand (Figure 2).

Rocks exposed in Fiordland record evidence for at least five different tectonic settings throughout the Phanerozoic. Before the Late Cretaceous, the microcontinent containing New Zealand known as Zealandia, was located on the Pacific margin of Gondwana. Western Fiordland contains two tectonic terranes that record sedimentation along the margin of Gondwana. The Buller Terrane in Fiordland consists of siliciclastic sandstones and mudstones derived from the Gondwana craton. The Takaka Terrane comprises Cambrian – Early Devonian siliciclastic, carbonate, and volcanic rocks derived from an intra-oceanic Cambrian age island arc (Figure 2) (Mortimer, 2004).

The Tuhua Orogeny (~380-360 Ma) is the earliest known orogeny recorded in Fiordland (Gibson, 1990; Oliver, 1980; Oliver & Coggon, 1979). During this time (~360 Ma) Ordovician (~480 Ma zircon date) arc-derived metasedimentary rocks (Takaka
Terrane) were intruded by granitic batholiths that caused high temperature and low-pressure sillimanite grade \( (P = 3-5 \text{ kbar}, \text{ and } T = 650^\circ-700^\circ\text{C}) \) metamorphism (Ireland & Gibson, 1998; Gibson & Ireland, 1996). Subsequent burial of these rocks resulted in a medium pressure \( (P = 7-8 \text{ kbar}) \) overprint at \( \sim 330 \text{ Ma} \) (Ireland & Gibson, 1998).

Between the Upper Paleozoic (\( \sim 375 \text{ Ma} \)) and the Lower Mesozoic (\( \sim 110 \text{ Ma} \)) craton-facing subduction occurred beneath Zealandia. During this interval the Paleozoic Gondwana margin rocks were intruded by subduction-related magmas forming the Median Batholith, a composite of Devonian to Early Cretaceous, mafic to felsic, I-type intrusions (Mortimer, 2004). Cretaceous magmatism (125-110 Ma) resulted in the emplacement of a dioritic batholith into the lower crust known as the Western Fiordland Orthogneiss. Emplacement of the Western Fiordland Orthogneiss was contemporaneous with continued contraction along the paleo-Pacific margin of Gondwana. Cretaceous arc-magmatism and contraction thickened the crust to \( >45 \text{ km} \) from \( \sim 146-110 \text{ Ma} \) (Mortimer, 1999; Waight et al., 1998; Bradshaw, 1989). Heat from the intrusion of the Western Fiordland Orthogneiss, and continued crustal thickening, resulted in high-pressure \( (P \sim 14 \text{ kbar}) \) and high-temperature \( (T > 750^\circ\text{C}) \) granulite facies metamorphism of the Western Fiordland Orthogneiss (Davids, 1999; Clarke et al., 2000; Daczko et al., 2001; Hollis et al., 2004; Flowers et al., 2005).

By the end of the Early Cretaceous (\( \sim 110 \text{ Ma} \)) arc magmatism waned and a sufficiently thickened crust (\( >45 \text{ km} \)) began to extend (Flowers et al., 2005). Fiordland contains a well-preserved record of continental extension during the mid-Cretaceous. Extensional shear zones that cut the 114 \( \pm 2.2 \text{ Ma} \) Western Fiordland Orthogneiss in
Doubtful Sound record the earliest known evidence of continental extension in Fiordland (Klepeis et al., 2007). A cross cutting post-kinematic dike in this shear zone yields an age of ~88.4 ± 1.2 Ma and constrains the end of continental extension (Klepeis et al., 2007). During this relatively rapid phase of continental extension, metamorphic core complexes developed at several crustal levels, resulting in the excision of deep crustal roots along low-angle normal faults.

Thermobarometry, thermochronology, and structures exposed in western New Zealand record the unroofing of the middle and lower crust (Gibson et al., 1988; Gibson & Ireland, 1995). The Paparoa metamorphic core complex in the Westland-Nelson region of New Zealand records 15-20 km of unroofing of the lower plate from $P = 5.2 ± 1.1$ kbar, and $T = <500°C$ to $P = 4 ± 1$ kbar and $T ≈500°C$ to $T ≈170°C$ during the interval from 110 – 90 Ma (Tulloch and Kimbrough, 1989; White, 1994; Spell et al., 2000).

Several contemporaneous middle and lower crustal extensional shear zones are exposed in Fiordland. The Doubtful Sound Shear Zone juxtaposes high-$P$ ($P ~12-13$ kbar) lower crustal components of the Western Fiordland Orthogneiss against mid-crustal Paleozoic orthogneiss and meta-sedimentary rocks at $P ~7-9$ kbar between 114-90 Ma (Oliver & Coggon, 1979; Gibson et al., 1988; Klepeis et al., 2007). The Sisters Shear Zone, located on Stewart Island, is the youngest known extensional shear zone in New Zealand (Kula et al., 2007). This shear zone records an increased cooling rate of 20-30°C Myr$^{-1}$ of Median Batholith granites along a mylonitic extensional fault from the interval 89-82 Ma (Kula et al., 2007), marking a the final phase of continental rifting before the opening of the Tasman Sea (Laird, 1994).
By ~85 Ma continental extension successfully rifted the lithosphere and sea floor spreading commenced (Mortimer, 2004). From ~84 Ma to ~52 Ma continued sea floor spreading opened the Tasman Sea (Gaina et al., 1998). The Australian-Pacific plate boundary developed by ~45 Ma and transitioned to transpression by 25-20 Ma. The Australian-Pacific plate boundary and Alpine Fault developed as early as ~25 Ma (Sutherland et al., 2000; Lamarche & Lebrun, 2000; Lebrun et al., 2003).
Figure 2 – Geologic setting of Fiordland, New Zealand. (a) Generalized geologic terrane map of the South Island of New Zealand, after Mortimer (2004). The location of Fiordland is shown. (b) Generalized geologic map of Fiordland showing locations of major structures, after Wood (1972); Oliver & Coggon (1979); Bradshaw (1989); Daczko et al. (2002); Tulloch & Kimbrough (2003); Klepeis et al. (2004); Turnbull et al. (2005); Allibone et al. (2005); Milan et al. (2005); Klepeis et al. (2007); and King et al. (2008). The study area for this thesis is highlighted.
a) **Simplified Terrane Map of the South Island, New Zealand**

- Cretaceous Granitoid Batholiths
  - Karamia, Paparoa, and Hohonu Batholiths
  - Median Batholith

- Permian - Early Cretaceous Sedimentary Rocks
  - Pahau Terrane
  - Caples Terrane

- Permian Basalts and Ophiolites
  - Maitai Terrane (ophiolite)
  - Brook Street Terrane

- Permian-Late Jurassic Volcaniclastic Marine Sediments
  - Muruhiku Terrane

- Cambrian-Devonian Metasedimentary Rocks
  - Takaka Terrane
  - Buller Terrane

- Fiordland

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b) **Simplified Geologic Map of Fiordland, New Zealand**

### Mesozoic Arc Rocks (Median Batholith)
- Western Fiordland Orthogneiss, Breakea & Resolution Gneiss (Early Cretaceous)
- Separation Point Suite (Early Cretaceous)
- Mafic-intermediate orthogneiss (Triassic-Early Cretaceous intrusive with some Paleozoic host rock)
- Plutonic and volcanosedimentary rocks (Triassic-Early Cretaceous)

### Paleozoic Basement (Takaka Terrane)
- Undifferentiated paragneiss and orthogneiss including the Deep Cove Gneiss (Cambrian-Pennsian)

### Structures
- Tertiary transpressional shear zone (Straight River Shear Zone)
- Indecision Creek Shear Zone
- Aniata Shear Zone
- Mid-Cretaceous extensional shear zones, including Resolution Island, Wet Jacket, Doubtful Sound, Mount Irene shear zones (dot in hanging wall)
- Thrust fault
2. Continental Extension

Continental lithosphere consists of the continental crust and lithospheric mantle, separated by the Mohorovičić discontinuity. The continental crust itself can be subdivided into three seismically and compositionally different layers: the upper crust (0 - ~15 km), the middle crust (~15 - ~30 km), and the lower crust (~30 - ~50 km) (Mooney et al., 1998; Christensen & Mooney, 1995; Rudnick & Gao, 2003). Along active plate boundaries, strength contrasts between the layers of the crust and the lithospheric mantle prove to be an important influence on deforming lithosphere (Jackson, 2002; Koons et al., 2003; Beaumont et al., 2004; Wijns et al., 2005).

Here I describe several models of extension that focus on the influence of lower crustal rheology on continental extension. Generally, these models suggest that the strength of the lower crust controls many elements of rifting, including: 1) the symmetry or asymmetry of rifted margins, 2) the degree of coupling or decoupling of the lower and upper crust, and 3) localizing or delocalizing of strain throughout the lithosphere. Understanding the role of lower crustal strength on the evolution of lithospheric extension is important in the context of this thesis for two reasons: 1) the predicted results have significant implications about the structural evolution of fabrics that are exposed in nature; 2) models provide a potential framework for interpreting observed geological relationships.

2.1 Models of Continental Extension

Models developed by Buck (1991) and Hopper & Buck (1996) describe three widely accepted end-member modes of lithospheric extension: core-complex mode, wide
rift mode, and narrow rift mode (Figure 3a-c). These models assume a three-layer rheology of continental lithosphere. They describe the middle-upper crust and lithospheric mantle as brittle/plastic layers sandwiching a thin and viscous lower crust. Hopper & Buck (1996) concluded that the mode of extension depends highly on the strength of the lower crust.

Very weak and low viscosity lower crust above an equally weak mantle lithosphere favors core-complex mode extension, where one localized detachment zone accommodates strain from the uplifting metamorphosed footwall block. Wide rift mode extension occurs when a weak lower crust is sandwiched by more viscous mantle lithosphere and upper crust. In this case, crustal buoyancy forces between the weak lower crust and rigid upper and middle crust are the dominant driving force for extension, and strain is spread over several normal faults across a large area (i.e. the Basin and Range province of N. America). A strong lower crust favors narrow rift mode extension. This occurs when crustal heat flow is low, and the viscosity of the lower crust is too high to flow easily. Strain is accommodated by normal faults localized into a narrow rift zone (i.e. the East African rift) (Figure 3a-c) (Buck, 1991; Hopper & Buck, 1996).

Numerical modeling by Wijns et al. (2005) suggests that the mechanical stratification of the crust controls the mode of continental extension. Wijns et al. (2005) separates continental crust into two layers, the upper and lower crust. The strength ratio between the upper and lower crust are the primary control on type of extension. In addition to strength contrasts, Wijns et al. (2005) include in their model strain-softening processes that occur in fault zones and can result in the localizing of strain into discrete
detachment faults. When there is a low strength ratio Wijns et al. (2005) predict 
distributed faulting mode extension, where strain is evenly accommodated by faults or 
plastic deformation in the crust. When the strength ratio between upper and lower crust is 
large, they predict core complex mode extension, and large amounts of strain are 
localized in detachment fault zones.

Another numerical model by Gessner et al. (2007) tests the significance of strain 
localization in the lower crust in terms of the evolution and thermal history of 
metamorphic core complexes. They conclude that a high viscosity lower crust would 
result in higher stresses transmitted to the upper crust causing asymmetrical segmentation 
of the upper crust (similar to distributed faulting mode of Wijns et al., 2005). In the case 
of a low viscosity lower crust, less stress is transmitted to the upper crust. The model 
predicts two symmetric oppositely dipping extensional shear zones will develop and 
result in the near symmetric advection of the lower crust, forming a dome shaped 
metamorphic core complex (Gessner et al., 2007). These authors suggest that regional 
metamorphic gradients will also record the symmetry or asymmetry of lithospheric 
extension. In the case of a high viscosity lower crust and asymmetric shear zones, their 
model predicts a small lateral gradient in metamorphism across the shear zone, with a 
localized zone of retrogression in the footwall near the detachment. A large lateral 
metamorphic gradient is expected with a significant zone of retrogression in the footwall 
when the lower crust has a low viscosity. Gessner et al. (2007) conclude that the strength 
of the lower crust has significant control on the geometry, and distribution of structures 
and metamorphism exposed on the surface (Figure 3h-k).
Hopper & Buck (1998) predict two styles of decoupling between the upper and lower crust that develop during continental extension. They suggest that once rifting is initiated, gravitational forces cause the lower crust to flow toward zones of thinning, or laterally away from uplifted rift margins. In the former case, strain is localized into a vertically coupled rift zone and deformation is dominated by pure-shear. In the latter case deformation in the upper crust migrates laterally away from deformation in the upper mantle. The lower crust must deform by simple shear to accommodate the spatially different spreading zones between the upper mantle and upper crust. Hopper & Buck (1998) suggested that the latter case will cause delocalization of strain throughout the lithosphere and may be responsible for processes occurring in the Basin and Range province of North America.

One group of models investigates the role of brittle-plastic, and ductile strain softening mechanisms on deformation patterns in an extending lithosphere (Huismans & Beaumont, 2003; 2007). In this case the presence or absence of strain softening processes in both the brittle and ductile regimes of the lithosphere control the symmetry or asymmetry of the rifted lithosphere. If only frictional-plastic fault zone weakening occurs (i.e. relatively strong lower crust) the models predict an asymmetric geometry. However, if frictional plastic, and ductile strain softening processes operate (i.e. a weak lower crust) these authors suggest that a symmetric geometry will develop as conjugate frictional shear zones that form in the upper crust sole out in the lower crust and propagate beneath rift flanks (Figure 3d-g) (Huismans & Beaumont, 2003; 2007).
Figure 3 – The results of numerical models of continental extension. (a-c) Buck (1991) predicted (a) core-complex mode in thick crust with a low yield strength, (b) wide rift mode in thick crust with a low yield strength above a strong upper mantle, and (c) narrow rift mode with thin crust and high yield strengths throughout the lithosphere. (d-g) Huismans & Beaumont (2007) predicted symmetric or asymmetric rift geometries for (d) low rift velocity \( (V= 0.6 \text{ mm a}^{-1}) \), (e) high rift velocity \( (V= 100 \text{ mm a}^{-1}) \), (f) strong lower crust, and (g) weak lower crust. (h-k) Gessner et al. (2007) predicted the symmetry or asymmetry of (h) strain partitioning, (i) temperature, (j) metamorphic grade, and (k) metamorphic history for rift settings with both high and low viscosity lower crust.
2.2 Metamorphic Core Complexes

Metamorphic core complexes are geological features traditionally defined by the juxtaposition of high-grade metamorphic rocks, against low-grade metamorphic rocks along a large-magnitude displacement, low-angle brittle-ductile normal fault (Armstrong, 1982; Coney & Harms, 1984; Lister et al., 1986). In this section I describe several field-based studies of metamorphic core complexes and interpretations of the processes that control their formation. I will use the results of these studies as a reference frame for interpreting processes that resulted in the formation of lower crustal metamorphic core complexes exposed in Fiordland, New Zealand. I am especially interested in factors that contribute to the localization of deformation on a detachment surface and processes that cause synextensional folding of the detachment.

Wernicke (1992) defined metamorphic core complexes by describing a shallowly dipping (~10-30°) low-angle normal fault that propagates through deep layers of the crust (Figure 4). As the lower plate is unloaded it experiences isostatic rebound and gently folds the detachment fault forming a dome-shaped core. This model is identifiable by a uniform shear sense on opposite sides of the metamorphic core in the lower plate of the detachment. Similarly, if the upper plate is extended along two oppositely dipping low-angle normal faults, the lower plate will be symmetrically denuded, with oppositely verging shear sense indicators on both sides of the lower plate. Isostatic rebound of the denuded core folds the oppositely dipping faults into a dome (Figure 4) (Wernicke, 1992). Since the development of this classic model, researchers describe many three-dimensional processes that result in the formation of metamorphic core complexes.
On the D’Entrecasteaux Islands, Papua New Guinea continental extension and metamorphic core complexes occur in close proximity (<30 km) to an ocean-spreading center, the Woodlark rift. Two field-based studies provide dramatically different explanations for the formation of exposed domes of mylonitic and basement rocks. The first study describes a density inversion of the crust due to ~15 km of island arc ophiolites (3.1 g cm$^{-3}$) located structurally above granitic continental crust (2.7 g cm$^{-3}$) (Martinez et al., 2001). High geothermal gradients near an ocean-spreading center allow the structurally lower continental crust to flow at high temperatures (>700°C). These authors suggest breaching of the ophiolite layer due to local extensional stresses, and density inversion of the crust, allow the ductile and buoyant lower crust to be extruded vertically through the ophiolite sequence. Martinez et al. (2001) developed a numerical model of this process that shows vertical extrusion of the lower crust will allow the crust to thin enough (~10-15 km) to explain an observed zone of crustal thinning from seismic receiver functions, and form metamorphic core complexes at the site of extrusion.
Figure 4 – Models of metamorphic core complex formation. (a) Simple shear produces an asymmetric detachment fault with a uniform shear sense. (b) Pure-shear mode produces a symmetric detachment fault with opposite senses of shear on either side of the metamorphic core, after Wernicke (1992).
Other authors argue that dome-shaped culminations of mylonitic and gneissic lower crustal rocks exposed on the D’Entrecasteaux Islands are not a result of diapirism, or crustal extrusion, but rather an exhumed detachment surface that has been upwarped (Little et al., 2007). These authors present structural and kinematic data that describe the evolution of the mylonitic detachment. Structural observations indicate there are two oppositely dipping shear zones on either side of the metamorphic core. Kinematic indicators from these shear zones display opposite senses of shear. Evidence for changing deformation styles includes the transition from ductile shearing to brittle faulting, and deformation textures in quartz that indicate that rocks in the footwall were exhumed from >10 km depth. These authors argue that motion on one detachment fault with >12.5 mm/yr slip rate could rapidly excise deep crustal rocks in a 1~6 m.y. timescale. Little et al. (2007) suggest that rapid unloading of the lower plate caused isostatic doming and back-rotation of the detachment surface. After the detachment was folded it was reactivated under high pore-fluid pressure with opposite shear senses on either side of the dome (Little et al., 2007). This interpretation includes the following processes: isostatic doming of the footwall, strain softening on the detachment surface, and low-angle normal detachments.

Little et al. (2007) draw parallels between the rapid exhumation rate and doming of the footwall in the D’Entrecasteaux Islands to processes documented at oceanic core complexes. Recent work by Schroeder & John (2004) and Karson et al. (2006) document the presence of metamorphic core complexes in oceanic crust along the Mid-Atlantic Ridge. Along these oceanic core complexes, extension occurs by the hydrous alteration of
upwelling mantle dunite, peridotite, and harzburgite into amphibolite and greenschist facies shear zones. As the footwall of the shear zone becomes unroofed it is hydrated and experiences retrograde metamorphism. A localized detachment shear zone develops in the weaker lithology and begins to accommodate extension as the core complex is uplifted. Extension along these core-complex shear zones serves as the plate boundary in the absence of active magmatism (Karson et al., 2006).

Several authors conclude that metamorphic core complexes can form by progressive folding of a detachment fault as a result of isostatic rebound of the metamorphic core. Gessner et al. (2001) describe a metamorphic core complex in western Turkey that is ~100 km wide and bound by two symmetrically arranged, oppositely dipping detachment systems. The footwall of the detachments comprises the uplifted metamorphic core and is folded into a syncline. The axis of this syncline trends parallel to the strike of the detachment faults. Using apatite fission-track thermochronology these authors show that the footwall syncline developed synchronously with the detachment systems. Fission-track thermochronology sampling positions plotted in their sampling locations on a cross-section show a younging trend toward the core of the footwall. The oldest cooling dates are closest to the detachment faults. These authors present a rolling-hinge model to describe the back folding of initially steep (∼40°-60°) normal faults that ‘roll-over’ to shallow dips (0°-20°) as a result of isostatic rebound of the unloaded footwall. In this case, development of a dome-shaped metamorphic core complex is a result of isostatic rebound and consequential folding of initially high-angle normal faults.
Davis et al. (2002) report the progressive folding of a detachment during extension from a metamorphic core complex exposed in Inner Mongolia, China. These authors attribute the development of low-angle normal faults to the gravitational collapse of a recently (~4 m.y.) thickened crust. Field evidence indicates that exhumed detachment faults developed parallel to inherited low-angle, mylonitic thrust ramp-flat structures. On both sides of the dome the detachment displays consistent top-to-the south shear indicators. Davis et al. (2002) explain this geometry by interpreting an initially south dipping low-angle detachment that follows the inherited, ramp-flat geometry. Isostatic rebound of the unloaded footwall folds the detachment. The new north-dipping limb of the fault is no longer in a favorable position for extension, and strain is localized to the south-dipping limb, preserving top-to-the south kinematic indicators. This process includes isostatic doming of an unloaded footwall and inherited anisotropy of the crust that promotes a low-angle normal fault.

Several authors document the structural and kinematic evolution of metamorphic core complexes. Chauvet & Seranne (1994) and Osmundsen et al. (2003) describe a transition from extension to strike-slip faulting during the late stages of extension that produces extension normal shortening. Chauvet & Seranne (1994) use paleostress indicators, including slickenlines orientations and tension gashes to describe a rotation of the principal stresses ($\sigma_1, \sigma_2$, and $\sigma_3$) as the metamorphic core complex is exhumed. Their results suggest that during the initial phases of extension $\sigma_1$ is vertical and accounts for the gently dipping normal displacement detachment systems. As the lower plate is further exhumed toward the surface and unloaded, paleostress indicators record a rotation of $\sigma_1$
from vertical to horizontal. These authors argue that during late stages of extension the changing orientation of $\sigma_1$ to the horizontal forms folds on the detachment surface that trend parallel to the extension direction. In this case, the development of horizontal shortening during late stages of extension results in a subsequent phase of strike slip faulting. These authors conclude that rotation of the stress tensor during extension results in extension parallel folding of the detachment surface and ultimately changes the strain response from extension to left-lateral translation (Chauvet & Seranne, 1994).

Similarly, Osmundsen et al. (2003) describe the formation of asymmetric foliations and normal displacement ductile-brittle extensional shear zones to be the result of an initial phase of subvertical flattening. Isostatic rebound of the lower plate caused extension perpendicular back-folds on the detachment surface. These authors also describe extension parallel folds on the detachment surface. Extensional structures are cross cut by strike-slip structures that are oriented obliquely to the extension directions. These authors conclude that isostatic doming of the unloaded footwall can occur simultaneously with the extension parallel shortening. They suggest that the transition from extension to transtension and strike-slip faulting is a result of unloading the lower plate and changing the orientation of $\sigma_1$ from vertical to horizontal.

One study (Aerden, 1994) describes a phase of bulk coaxial vertical shortening around inferred gneiss domes in the Variscan Pyrenees. The presence of a gneiss dome in the crust is evidenced by the concentric arrangement of metamorphic isograds toward the center of the Lys Caillaouas Massif, and the Garonne Dome (Aerden, 1994). This study uses rotated porphyroblasts in the amphibolite facies schist, inferred to be the upper plate,
to describe the kinematics of gravitational collapse around the margins of the dome. Structures include a subhorizontal crenulation cleavage that cuts in an inherited subvertical schistose foliation in the core of the domes. Toward the margins of the domes, crenulation cleavages dip gently in opposite directions and rotated porphyroblasts display opposite senses of rotation. Aerden (1994) argues that initial steep foliations, and the dome structure are a result of an early phase of shortening. The later development of shallow crenulation cleavages, with opposite vergence on either side of the dome, record the gravitational collapse of the thickened crust around the rigid underlying dome. The results of kinematic analyses indicate that gravitational collapse was achieved by bulk coaxial flattening that was partitioned into zones of non-coaxial shear on opposite flanks of a rigidly behaving dome.

Fletcher & Bartley (1994) report constrictional strain in a non-coaxial shear zone from the Mojave metamorphic core complex, California. Structural observations document the existence of northeast-trending mineral stretching lineations that record the extension direction. The detachment surface is folded by two sets of folds. One set of folds have axes that trend perpendicular to the extension direction and are interpreted to be a result of isostatic doming of the metamorphic core. The second set of folds trend parallel to the extension direction and suggest a component of shortening perpendicular to extension. Finite strain analyses from footwall mylonites indicate that the detachment fault formed in the field of apparent constriction, which supports the interpretation of two shortening directions during extension. Fletcher & Bartley (1994) argue that constriction could be achieved under uniaxial stress fields in the Basin and Range. They state that
Andonian fault theory would allow continental extension in a uniaxial stress field. An initial uniform stress tensor ($\sigma_1=\sigma_2=\sigma_3$) could result in a uniaxial stress tensor ($\sigma_3<\sigma_2=\sigma_3$) if extension is imposed along $\sigma_3$. They suggest the field of constriction would explain the formation of both extension parallel, and extension perpendicular folds observed on the Mojave metamorphic core complex (Fletcher & Bartely, 1994).

The Shuswap metamorphic core complex exposed in British Columbia, Canada records the progressive evolution of a core complex and the influence of partial melt during shearing. Johnson (2006) describes the Shuswap metamorphic core complex being delimited by west dipping ductile-brittle normal faults and mylonite zones with asymmetric top-to-the west shear indicators. This detachment is separated into two segments that are offset across a 45 km wide transfer zone, where a mylonitized dome of high-grade (kyanite+sillimanite) migmatite is overprinted by brittle normal faults. Johnson (2006) concludes that the emplacement of a granite during extension caused migmitization of mid-crustal rocks and served as a strain softening process that facilitated shearing. Continued extensional shearing along this weak zone unloaded the migmatized dome causing isostatic uplift. Shearing along the detachment stopped when footwall doming folded the detachment into an orientation no longer favorable for shear. Continued regional extension caused strain to propagate further into the hanging wall, forming high-angle normal faults that root in, or cut, the mylonitic detachment fault. Johnson (2006) concludes that magmatism, nucleation of low-angle shear zones, heterogeneous extension, doming, and exhumation by developing new extensional
structures are all critical interrelated processes that result in the formation of a metamorphic core complex.

A similar study by Vanderhaeghe et al. (2003) describes the evolution of another Shuswap metamorphic core complex in terms of its cooling history using \(^{40}\text{Ar}/^{39}\text{Ar}\) dating. These authors report two phases of exhumation: from 700°C to 300°C related to crustal thinning and the development of extensional detachments in the mid-crust and a subsequent phase (~5 m.y. later) related to regional Basin and Range extensional tectonics. These authors suggest that mechanical weakening of the crust by partial melting enhanced an early phase of ductile shearing along the detachment fault. Block tilting and brittle normal faults cut this mylonitic detachment after the core has thermally relaxed (Vanderhaeghe et al., 2003).

Gneiss domes described in collisional tectonic settings (i.e. Himalayas) are also defined by the juxtaposition of high-grade metamorphic rocks from deeper levels in the crust against lower-grade metamorphic rocks near the surface along ductile normal faults. Gneiss domes are attributed to ductile flow of weak, and thickened mid-lower crust that drives orogenic collapse of thickened crust. Although gneiss domes form in collisional settings they may be compared to metamorphic core complexes because they are defined by the juxtaposition of high-grade metamorphic rocks against low-grade metamorphic rocks along normal faults. Teyssier & Whitney (2002) explain this processes as a positive feedback loop between two mechanisms that produces negatively buoyant mid-lower crust, isothermal decompression (thinning) and decompression melting. They suggest that partial melting of thickened mid-crustal rocks produces negative buoyancy that causes
isostatic doming. In turn, unloading of the footwall causes decompression melting, and enhances the negative buoyancy effect. Teyssier & Whitney (2002) suggest that tectonic thinning of the crust, which amplifies the effects of isothermal decompression, further enhances this feedback loop. They conclude that the signatures of this process are dome shaped structures cored by high-grade migmatites and they suggest that all metamorphic core complexes are affected by some component of isothermal decompression.

These field-based studies of metamorphic core complexes show that the formation of dome shaped footwall rocks exhumed from deeper crustal levels along normal faults involve a wide range of processes including: large displacement low-angle normal faulting, rotation of initially high-angle normal faults, isostatic rebound of a tectonically denuded core, strain softening mechanisms on detachment surfaces, diapirism, mid-crustal flow during contraction, or density inversion of the crust. These studies also demonstrate that the metamorphic core complexes form in a range of tectonic settings including: continental collision (gneiss domes in Tibet), transtension (Scandinavian Caledonides), orogenic collapse (Pyrenees), continental extension (Basin and Range), newly rifted continental margins (Papua New Guinea), and in settings were there may be a density inversion of the crust. The mechanisms for core complex formation and exhumation of the lower plate also depend on the antecedent geology of the range and the present tectonic setting. It is my intention that this thesis will enhance our understanding of processes that result in the formation of metamorphic core complexes in terms of the three-dimensional finite strain and kinematic history of flow from a lower crustal metamorphic core complex.
Chapter IV: Three-dimensional finite strain and kinematics of flow from a lower crustal extensional shear zone: implications for continental rifting

The Resolution Island Shear Zone is a ~1 km thick upper amphibolite facies extensional shear zone exposed in Fiordland, New Zealand that juxtaposes eclogite and garnet-granulite facies rocks in the lower plate against mid-crustal upper amphibolite facies schist and gneiss in the upper plate. Contrasting P-T conditions in the upper and lower plates suggest that the Resolution Island Shear Zone accommodated a minimum of ~29 km of vertical displacement, satisfying the criteria for a metamorphic core complex. This shear zone attenuates gneissic foliations from lower crustal components of the Western Fiordland Orthogneiss, providing a lower age limit of ~115 Ma. This age constraint indicates that the Resolution Island Shear Zone formed contemporaneously with three other extensional shear zones in Fiordland associated with the mid-Cretaceous (~114-90 Ma) rifting of Gondwana.

We present structural and three-dimensional finite strain and kinematic data that describe the progressive evolution of the Resolution Island Shear Zone and the coeval formation of a second, antithetic shear zone called the Wet Jacket Shear Zone. Gneissic banding (S₁) preserves eclogite (grt+om) and granulite (grt+pl+cpx+opx) facies metamorphic textures that record burial metamorphism. Pervasive garnet-granulite grt+hbl+pl foliations and lineations (S₂/L₂) truncate or attenuate S₁ fabric in the lower plate, marking the onset of extension. Locally, 10 cm -100 m thick penetrative upper amphibolite facies (pl+hbl) ductile shear zones (S₃/L₃) attenuate S₂/L₂ fabric. These S₃/L₃ shear zones coalesce to form the Resolution Island Shear Zone. At this stage an antithetic upper amphibolite facies shear zone formed in the upper plate of the Resolution Island Shear Zone.

S₃/L₃ shear zones in the lower plate anastomose around dismembered lenses of eclogite facies rocks, forming a dome-and-basin pattern. S₃/L₃ shear zones display both top-down-to-the northeast and top-down-to-the southwest kinematic indicators, with respectively oriented mineral stretching lineations (L₃). Three-dimensional finite strain analyses from S₃/L₃ fabric in the lower plate of the Resolution Island Shear Zone yielded prolate finite strain ellipsoids. Geochemical comparisons between the orthogneiss host and shear zone rocks suggest that deformation occurred with <10% volume loss. Bulk kinematic analyses at the scale of the lower plate (~15 km²) show coaxial-dominated rotations of lines toward the stretching direction (NE-SW) of the finite strain ellipsoid. We determined that the formation of the Resolution Island Shear Zone, a major core-complex-style ductile detachment fault occurred in the field of coaxial-dominated constriction. The kinematics of deformation in the Resolution Island Shear Zone are symmetric to three coeval previously described shear zones in Fiordland. This suggests a general kinematic coupling of deformation between the lower and upper levels of the crust.
1. Introduction

The ability of the lower crust to flow has an important control on the mode, geometry, and partitioning of strain within the lithosphere during continental rifting (Buck, 1991; Hopper & Buck, 1996; Wijns et al., 2005). In situations where the lower crust can flow easily, continental extension can result in the formation of metamorphic core complexes. These structures typically are characterized by dome-shaped uplifts of high-grade rocks in the footwall juxtaposed against lower-grade rocks in the hanging wall along a low-angle normal fault (Armstrong, 1982; Coney & Harms, 1984; Lister et al., 1986; Wernicke, 1992).

Both field-based and numerical models of continental extension have established that the relative strengths of different layers in the crust are a primary control on the development of several aspects of metamorphic core complexes. These include the symmetry of crustal shear zones, the partitioning of strain, and the general geometry of the deformation (Hopper & Buck, 1996; Wijns et al., 2005; Johnson, 2006; Huismans & Beaumont, 2003; 2007; Gessner et al., 2007). In addition, processes that result in large-magnitude, dome-shaped normal faults may include isostatic doming of the unloaded lower plate (Davis et al., 2002), strain softening mechanisms within the detachment fault (Johnson, 2006), high extensional strains (Huismans & Beaumont, 2007), and hot or weak lower crust (Vanderderhaeghe et al., 2003; Johnson, 2006). Nevertheless, the kinematics of flow in the lower crust, and the possible kinematic linkage of deformation in the lower crust to deformation at higher crustal levels during continental extension, remain poorly understood.
Western New Zealand provides excellent exposures of metamorphic core complexes that formed during the mid-Cretaceous (114-90 Ma) continental rifting of Gondwana (Gibson et al., 1988; Tulloch & Kimbrough, 1989; Spell et al., 2000; Gibson et al., 1996; 1998; Klepeis et al., 2007; Kula et al., 2007). Mid-Cretaceous extensional structures in western New Zealand including the Paparoa metamorphic core complex (Tulloch & Kimbrough, 1989), a metamorphic core complex exposed in Doubtful Sound (Gibson et al., 1988), the Mount Irene shear zone (Scott & Cooper, 2006), and the Resolution Island Shear Zone (Klepeis et al., 2007) record the contemporaneous evolution of different levels of the crust during continental extension (discussed in Section 2). The recent discovery of high-pressure eclogite facies rocks ($P = 17$-$19$ kbar; De Paoli et al., 2007) in the lower plate of the Resolution Island Shear Zone make the Resolution Island Shear Zone the deepest know extensional structure in western New Zealand associated with the Mid-Cretaceous breakup of Gondwana. These constraints provide an ideal field setting for testing the relationship between deformation in the lower crust to deformation at higher crustal levels during continental extension.

In this paper, we present field-based data describing the flow of material in the footwall of a lower crustal (~61-69 km depth) extensional shear zone, called the Resolution Island Shear Zone, and a newly discovered shear zone termed the Wet Jacket Shear Zone from a ~30 km transect in Fiordland, New Zealand. We describe flow in the lower crust by: 1) determining the relative timing and geometry of deformation in upper and lower plates of the Resolution Island Shear Zone; 2) determining the significance of lower crust fabrics in terms of a three-dimensional finite strain history; and 3)
determining the kinematic evolution of flow from a 15 km$^2$ exposure of the lower plate and the Resolution Island Shear Zone. Using these data we compare the geometry and kinematic evolution of the Resolution Island Shear Zone and Wet Jacket Shear Zone with coeval shear zones that formed at higher levels in the crust, and discuss the possible kinematic linkages of the lower crust with the upper crust during continental extension. Our results reveal the progressive evolution of fabrics in the lower crust that culminated in the formation of a lower crustal metamorphic core complex. Dome-and-basin patterns displayed by the Resolution Island Shear Zone, and three dimensional finite strain and kinematic analyses from the lower plate indicate that coaxial-dominated constrictional strain played an important role in the development of extensional structures in the lower crust during the mid-Cretaceous rifting of Gondwana. We suggest that extending by coaxial-dominated constriction in the lower crust can result in the formation of metamorphic core complexes at several crustal levels and a generally symmetric pattern of deformation in the crust during continental rifting.

2. Geologic background

Western New Zealand records several orogenic pulses since the mid-Paleozoic. Emplacement of granitic batholiths into the crust during the Tuhua orogeny (~360 Ma) caused high temperature and low-pressure sillimanite grade ($P = 3-5$ kbar, and $T = 650^\circ$-$700^\circ$C) metamorphism of Lower Paleozoic metasedimentary host rock (Oliver & Coggon, 1979; Oliver, 1980; Gibson, 1990; Gibson & Ireland, 1996; Ireland & Gibson,
Tectonic thickening resulted in a kyanite grade medium pressure (7-8 kbar) metamorphic overprint during the Carboniferous (~330 Ma) (Ireland & Gibson, 1998). An Early Cretaceous (~146-110 Ma) record of extensive magmatism and crustal thickening overprints Paleozoic structures, metamorphic textures, and intrusions (Oliver, 1980; Gibson et al., 1996; Ireland & Gibson, 1998; Mortimer et al., 1999). Subduction-related arc-magmatism and contraction thickened the crust to >45 km and emplaced gabbroic and dioritic melts broadly known as the Median Batholith into the crust from ~146 to ~110 Ma (Bradshaw, 1989; Waight et al., 1998; Mortimer et al., 1999; Clarke et al., 2000). Tectonic burial of this batholith accompanied granulite facies metamorphism at $T \geq 750^\circ$ C, and $P \sim 14$ kbar in the widespread Western Fiordland Orthogneiss, a result of continued contraction along the Pacific-Gondwana margin during the mid-Cretaceous (Figure 5) (Clarke et al., 2000; Daczko et al., 2001; Flowers et al., 2005; Hollis et al., 2004).

Thermobarometry, thermochronology, and structural observations from several extensional shear zones in western New Zealand record the unroofing of the middle and lower crust from the period (~114-90 Ma). During this time New Zealand rifted away from the margin of Gondwana (Gibson et al., 1988; Tulloch and Kimbrough, 1989; Muir et al., 1994; Gibson & Ireland, 1995; Spell et al., 2000; Scott & Cooper, 2006; Kula et al., 2007; Klepeis et al., 2007). Several metamorphic core complexes formed at this time at different levels of the crust that are now exposed in western New Zealand. In the Westland-Nelson region of New Zealand the Paparoa metamorphic core complex records 15-20 km of unroofing of the lower plate from $P = 5.2 \pm 1.1 - 4 \pm 1$ kbar (White, 1994).
and $T<500^\circ$C and $T\approx500^\circ$C to $T\approx170^\circ$C during the interval from 110 – 90 Ma. (Figure 5) (Tulloch & Kimbrough, 1989; Spell et al., 2000). In western Fiordland the Mount Irene Shear Zone (Scott & Cooper, 2006) and the Doubtful Sound Shear Zone (Gibson et al., 1988; Klepeis et al., 2007) juxtapose high-$P$ ($P\sim12-13$ kbar) lower crustal components of the Western Fiordland Orthogneiss against mid-crustal Paleozoic orthogneiss and meta-sedimentary rocks at $P\sim6-9$ kbar between 114-90 Ma (Scott & Cooper, 2006; Klepeis et al., 2007). The Sisters Shear Zone exposed on Stewart Island records an increased cooling rate of 20-30°C of Median batholith granites along a mylonitic extensional fault during the interval 89-82 Ma, marking the final phase of continental rifting before the opening of the Tasman Sea (Kula et al., 2007). Rifting of the continental lithosphere completed and transitioned to sea floor spreading in the Tasman Sea from 84-52 Ma (Gaina et al., 1998).

A late Tertiary north-northeast striking steeply dipping oblique transpressional fault known as the Straight River Shear Zone overprints contractional and extensional structures in western Fiordland (King et al., 2008). Active transpressional faults in Fiordland are related to activity along the Alpine fault since the Late Tertiary (~25 Ma) (Lamarche & Lebrun, 2000; Lebrun et al., 2003; King et al., 2008).
Figure 5 – Geologic map of Fiordland. (a) Locality map of New Zealand showing the location of Fiordland, the Alpine Fault and the Paparoa Metamorphic core complex. (b) Generalized geologic map of Fiordland showing locations of mid-Cretaceous extensional structures, the pressure conditions in the footwall of these structures, and the time period during which they were active, after Wood (1972); Oliver & Coggon (1979); Bradshaw (1989); Daczko et al. (2002); Tulloch & Kimbrough (2003); Klepeis et al. (2004); Turnbull et al. (2005); Allibone et al. (2005); Milan et al. (2005); Scott & Cooper (2006); Klepeis et al. (2007); Kula et al. (2007); and King et al. (2008). Thermobarometry and geochronology data are from: ¹Gibson et al. (1988); ²Tulloch & Kimbrough (1989); ³White (1994); ⁴Davids (1999); ⁵Spell et al. (2000); ⁶Scott & Cooper (2006); ⁷Klepeis et al. (2007); ⁸Kula et al. (2007); and ⁹De Paoli et al. (2007). The boxed area shows the field setting for this study.
Mesozoic Arc Rocks (Median Batholith)
- Western Fiordland Orthogneiss, Breaksea & Resolution Gneiss (Early Cretaceous)
- Separation Point Suite (Early Cretaceous)
- Mafic-intermediate orthogneiss (Triassic-Early Cretaceous intrusives with some Paleozaic host rock)
- Plutonic and volcanosedimentary rocks (Triassic-Early Cretaceous)

Paleozoic Basement (Takaka Terrane)
- Undifferentiated paragneiss and orthogneiss including the Deep Cove Gneiss (Cambrian-Permian)

Structures
- Tertiary transpressional shear zone (Straight River Shear Zone)
- Indecision Creek Shear Zone
- Anita Shear Zone
- Mid-Cretaceous extensional shear zones, including Resolution Island, Wet Jacket, Doubtful Sound, Mount Irene and Sissers shear zones (dot in hanging wall)
- Thrust fault
3. Field relationships

3.1 General structure

We determined the structure of a ~30 km transect in western Fiordland across the north shore of Resolution Island and Wet Jacket Arm. There are two general groups of rocks exposed along this transect, including: 1) Undifferentiated Devonian greenschist to upper amphibolite facies paragneiss know as the Deep Cove Gneiss (Oliver & Coggon, 1979), and 2) Mesozoic (123-115 Ma) dioritic to gabbroic upper amphibolite, granulite and eclogite facies orthogneiss, including the Western Fiordland Orthogneiss, the Supper Cove Orthogneiss, the Breaksea Gneiss, and the Resolution Gneiss (Hollis et al., 2004; Milan et al., 2007). Figure 6 shows a geologic map of the field area on Resolution Island and Wet Jacket Arm.

Exposed in Wet Jacket Arm, the Deep Cove Gneiss consists of undifferentiated Paleozoic greenschist to upper amphibolite facies paragneiss (grt+pl+ep+bt+qtz; mineral abbreviations follow the scheme of Kretz, 1983), and mafic orthogneiss that contain garnet leucosome and migmatite pods, schist, and marble. Amphibolite and dioritic dikes intrude the Deep Cove Gneiss and are commonly stretched and dismembered. The Western Fiordland Orthogneiss and Supper Cove Orthogneiss are dioritic to gabbroic, upper amphibolite to granulite facies (grt+pl+hbl+cpx+bt+ep+czo) coarse-grained orthgneiss with pegmatite intrusions. The Western Fiordland Orthogneiss and Supper Cove Orthogneiss are associated with the Early Cretaceous (115-128 Ma) emplacement of the Median Batholith (Tulloch & Kimbrough, 2003; Hollis et al., 2004; Milan et al., 2007).
The north shore of Resolution Island exposes two main rock units, the Breaksea Gneiss and the Resolution Gneiss. The Breaksea Gneiss is an upper amphibolite (grt+pl+hbl±ep+px) and garnet-granulite (grt+pl+px±opx+om+hbl) to eclogite (grt+pl+opx+cpx) facies orthogneiss. The Resolution Gneiss is an upper amphibolite facies (pl+hbl±grt+bt) mafic orthogneiss (Figure 6). Both the Resolution Gneiss and the Breaksea Gneiss contain 1 m² - 50 m² lenses or dikes of mafic eclogite facies (om+grt) orthogneiss that occur as dismembered lenses or boudins. The Breaksea Gneiss and Resolution Gneiss located on Resolution Island are interpreted as structurally lower components of the Western Fiordland Orthogneiss based on zircon and U/Pb lead age data, field relations, and whole rock geochemistry (Milan et al., 2007).

Two newly discovered north-striking ductile shear zones termed the Resolution Island Shear Zone and the Wet Jacket Shear Zone cut Paleozoic and Mesozoic age rocks (Figures 6 & 7). These shear zones are distinguished by the localized (~1-2 km thick) attenuation and transposition of older pervasive gneissic foliations. The Resolution Island Shear Zone (defined below) consists of anastomosing upper amphibolite facies shear bands. These shear bands form synthetic and antithetic pairs that dip moderately to the east and west and record top-to-the northeast and top-down-to-the southwest shear senses, respectively. The Wet Jacket Shear Zone dips gently to the southwest with displays top-down-to-the southwest shear sense (Figure 6). For simplicity, we determined that structures along this transect can be divided into four domains: I) the Resolution Island Shear Zone, II) the lower plate of the Resolution Island Shear Zone, III) the upper plate of the Resolution Island Shear Zone, and IV) the Wet Jacket Shear Zone.
**Figure 6** – Geologic map of the field area in Fiordland, New Zealand. (a) Locality map showing the major lithologic units and the locations of (b) Resolution Island and (c) Wet Jacket Arm map areas. (d-h) Equal-area lower hemisphere stereonet projections of (d) contoured poles to foliation from $S_2$ (hollow dots) and $S_3$ (solid dots) fabric on Resolution Island, (e) plagioclase and feldspar stretching lineations $L_2$ (hollow dots) and $L_3$ (solid dots) from Resolution Island, (f) mineral stretching lineations $L_2$ (hollow dots) and $L_3$ (solid dots) from Wet Jacket Arm, (g) contoured poles to $S_2$ foliation (hollow dots) in Wet Jacket Arm, and (h) cylindrical best-fit to folded $S_2$ foliations (hollow dots) in the Wet Jacket shear zone. Solid dots are poles to $S_3$ proto-mylonite zones and solid diamonds are $F_3$ fold axes.
Lithologic Units
- Western Fiordland Orthogneiss
dioritic orthogneiss
(p = biot + grt + al + cpx + pl)
- Breaksea Gneiss
interlayered granulite
and eclogite facies gneiss
(grt + ssp + cpx + om + hbl)
- Resolution Gneiss
amphibolite facies gneiss
(pl + hbl + biot + cpx + ca)
- Supper Cove Orthogneiss
dioritic orthogneiss,
(p = biot + al + grt + cpx)
- Deep Cove Gneiss
undifferentiated paragneiss,
orthogneiss schist, and marble

Geologic Symbols
- foliations
- folds
- lineations
- contacts
- faults and shear zones
- Straight River Shear Zone
- Resolution Island Shear Zone
- Wet Jacket Shear Zone
Figure 7 – Geologic cross sections from Resolution Island and Wet Jacket Arm. (a) Composite A-A’ and B-B’ transects from Resolution Island showing cross cutting relationships in the lower plate, and the Resolution Island Shear Zone, drawn parallel to transport direction (location in Figure 6b). (b) C-C’ transect from Resolution Island showing dome-and-basin geometry in the lower plate perpendicular to Figure 7a (location in Figure 6b). (c) Wet Jacket Arm D-D’ transect showing the upper plate of the Resolution Island Shear Zone and the Wet Jacket Shear Zone (location in Figure 6c). See text for details. Mineral abbreviations are after Kretz (1983).
**Tertiary**

- Straight River Shear Zone

**Mid-Cretaceous**

- pl + hbl ± bt (Breaksea and Resolution Gneiss)
- grt + pl + hbl ± cpx + opx (Breaksea & Resolution Gneiss)
- om + gt ± di + opx + ky (eclogite zone)
- Western Fiordland Orthogneiss

**Paleozoic**

- Deep Cove Gneiss
- hbl + czo + bt vein-filled fractures S₃
- migmatite zone

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VE = 1:1
3.2 The Resolution Island Shear Zone (Domain I)

The Resolution Island Shear Zone is marked by the occurrence of upper amphibolite facies shear zone foliations in a ~0.5-1 km thick zone. Fine-grained (1-5 mm) foliations defined by aligned plagioclase and amphibole grains dominantly dip 30°-65° toward the east and truncate or attenuate older gneissic fabrics. Plagioclase and amphibole mineral stretching lineations plunge toward the northeast and southwest (Figure 6c). Transposition of shear zone foliations indicate top-down-to-the northeast normal displacements, but also include antithetic top-down-to-the southwest shear bands (Figure 7a). Within the shear zone, foliations anastomose around the boundaries of rigid lenses defined by eclogite and amphibolite facies rocks. Wrapping of foliations and lineations around competent lenses produces a folded or dome-shaped shear zone geometry, and mineral stretching lineations that plunge toward both the northeast and southwest (Figure 6a, d, e). The Resolution Island Shear Zone juxtaposes upper amphibolite, garnet-granulite and eclogite facies gneiss (the Breaksea Gneiss and the Resolution Gneiss) from the lower plate against greenschist and amphibolite facies paragneiss (the Deep Cove Gneiss) of the upper plate (Figures 6 & 7).

3.3 Metamorphic core and lower plate of the Resolution Island Shear Zone (Domain II)

We determined that there are four groups of structural fabrics distinguishable on the basis of cross cutting relationships, mineral assemblage, and spatial distribution in the lower plate of the Resolution Island Shear Zone. Magmatic layering and igneous contacts preserved in the lower plate of the Resolution Island Shear Zone are recrystallized and overprinted by at least three phases of foliations and lineations. Dismembered dikes or
lenses of eclogite facies orthogneiss with the assemblage grt+om±pl+cpx+opx represent igneous contacts and magmatic layering. Mafic dike (grt+om±opx+cpx) swarms are folded. Inside eclogite zones compositional layering defines eclogite – garnet-granulite facies foliations with the assemblage grt+cpx+opx±pl (S$_1$).

At the boundaries of eclogite facies lenses and dikes, omphacite and clinopyroxene foliations (S$_1$) are truncated, folded or transposed by recrystallized plagioclase and hornblende ± garnet gneissic foliations (S$_2$) (Figure 6b, d, & 8). Mineral stretching lineations (L$_2$) defined by stretched, recrystallized feldspar aggregates and parallel hornblende grains trend toward the northeast and southwest (Figure 6b, e). Garnet-granulite and upper amphibolite facies foliations in the Breaksea Gneiss and Resolution Gneiss comprise a pervasive gneissic foliation (S$_2$) in the lower plate of the Resolution Island Shear Zone.

Locally, fine-grained (<2mm) ~2cm – ~10m thick upper amphibolite (pl+hbl) facies ductile shear zones transpose upper amphibolite (pl+hbl), garnet-granulite (grt+pl+hbl±cpx) and eclogite (om+grt) facies foliations (S$_1$ & S$_2$) and forms part of the Resolution Island Shear Zone. The parallel alignment of plagioclase and hornblende defines a penetrative foliation (S$_3$) within these shear zones. Alignment of recrystallized plagioclase and hornblende minerals forms strong northeast- and southwest-trending mineral stretching lineations (L$_3$) (Figure 6b, d, e). In some places where S$_2$/L$_2$ garnet-granulite fabrics (grt+pl+hbl) are transposed into S$_3$/L$_3$ shear zones, a retrograde metamorphic gradient is observed. Hornblende and feldspar rims occur on garnet and...
clinopyroxene grains in penetrative $S_3$ shear zones and commonly have completely replaced garnet and clinopyroxene grains (Section 4; Figures 8, 9 & 12).

**Figure 8** – Field relationship sketches. (a, b) Profile sketches showing the relationships between $S_1$, $S_2$ and $S_3$ foliations. Dismembered lenses and dikes displaying eclogite facies gneissic foliations ($S_1$) are truncated by pervasive $S_2$ and $S_3$ foliations. $S_2$ foliations are localized forming penetrative shear zones ($S_3$) that anastomose around the boundaries of rigid mafic zones. $S_3$ shear zones display top-to-the northeast and top-to-the southwest shear sense. (c) Photograph of an eclogite facies (grt+om+cpx±opx) dike that is folded around $S_2$ foliations. Rigid eclogite zones do not display $S_2$ foliations (pen for scale). (d) Photograph of $S_2$ foliations attenuated by a penetrative amphibolite facies $S_3$ shear zone with top-down-to the southwest shear sense (pen for scale).
Figure 9 – Sketches showing the evolution of fabrics in the lower plate of the Resolution Island Shear Zone. (a) In westernmost localities, centimeter-scale $S_3/L_3$ shear zones attenuate $S_2/L_2$ fabric. (b) Following a transect toward the east, $S_3/L_3$ shear zones become meter-scale and increase in occurrence. (c) Mafic lenses that display $S_1$ gneissic foliations are dismembered and deflect $S_3/L_3$ shear zones. (d) Within the Resolution Island Shear Zone anastomosing $S_3/L_3$ fabrics coalesce and display top-down-to-the northeast shear-sense.
Garnet-granulite (grt+pl+hbl±cpx, Breaksea Gneiss) and upper amphibolite (pl+hbl, Resolution Gneiss) facies gneissic foliations (S$_2$) are gently folded forming a series of antiforms and synforms in the lower plate of the Resolution Island Shear Zone. Fold axes trend both northeast-southwest, and north-south forming doubly plunging folds. Interference patterns between antiforms and synforms form broad map-scale (~100 m$^2$ – 1 km$^2$) dome-and-basin structures in the lower plate of the Resolution Island Shear Zone (Figures 6b, d, & 7a).

At the outcrop scale shear zone foliations (S$_3$) envelop rigid pods (1 m$^2$ – 50 m$^2$) of eclogite and garnet-granulite that display S$_1$, and S$_2$/L$_2$ fabrics (Figure 10). Rigid pods do not display S$_3$/L$_3$ shear zone recrystallization textures and instead deflect S$_3$/L$_3$ shear zone foliations and lineations around boundaries of these rigid zones. Eclogite lenses and dikes are commonly cross cut by brittle fractures filled with lower amphibolite facies (hbl+ep+czo) veins. Lower amphibolite facies veins in brittle fractures (hbl+czo+ep+bt) cross cut S$_1$/S$_2$ fabrics within rigid eclogite pods but do not cross cut S$_3$ shear zones and owing to their mineral assemblage (hbl+ep), they are grouped with S$_3$ fabric. In plan view S$_3$ shear zone foliations trace a circle around mafic pods and in profile view S$_3$ shear zone foliations transpose S$_1$ and S$_2$/L$_2$ gneissic fabrics forming both top-down-to-the northeast and top-down-to-the southwest shear bands (Figure 8, 9, 10). Hornblende and plagioclase shear zone mineral stretching lineations (L$_3$) follow margins of domes around mafic pods and dominantly plunge towards the northeast and southwest. These structures are associated with respectively oriented S$_3$ shear bands on the northeast and southwest margins of domes (Figure 8, 9, & 10). This pattern of ductile shear zone fabrics
anastomosing around relatively rigid lenses creates a dome-and-basin pattern in the shear zone fabric of an outcrop.

Minor upper amphibolite facies shear zones (S₃) increase in occurrence, thickness from west to east until localized in the Resolution Island Shear Zone (Figure 7 & 9). In the westernmost part of the section, pervasive S₂/L₂ fabric is transposed into <2m scale northeast and southwest ductile shear zones (S₃) that are apparently discontinuous (Figure 9a, b). Following a transect toward the east, meter-scale amphibolite facies shear zones (S₃) enlarge to <10m thick and become increasingly abundant. Towards the central and eastern localities, the S₃ fabric becomes pervasive and forms anastomosing arrays around mafic (eclogite and garnet-granulite) pods truncating or attenuating S₁ and S₂ fabrics. Farther eastward, the shapes of S₃ fabrics preserve dome-and-basin geometries. However, amphibolite facies mineral assemblages replace eclogite or granite-granulite mineral assemblages in rigid pods. In eastern localities all S₃/L₃ fabric is localized into the Resolution Island Shear Zone (S₃) forming a ~0.8 km thick low-angle (~30°) ductile normal fault. This trend defines an increasing trend in the occurrence of S₃/L₃ ductile normal-sense shear zones toward the northeast into the Resolution Island Shear Zone (Figure 9).
Figure 10 – Form map of the Resolution Island Shear Zone. (a) Structural form map showing dome-and-basin patterns displayed by $S_3/L_3$ shear zones. Mineral (pl+hbl) stretching lineation ($L_3$) traces (long-dashed lines) show northeast- and southwest-trending transport directions. Foliation planes ($S_3$) anastomose around the margins of domes and basins. (b) Photograph of $S_3/L_3$ amphibolite facies shear zone dome-and-basin pattern. This dome is located between localities 15f, and 15a in (a). Sample localities 15a-f show the orientations of finite strain ellipses from the XY plane of each sample’s finite strain ellipsoid (discussed in section 5.1).
3.4 Upper plate of the Resolution Island Shear Zone (Domain III)
We distinguished three groups of fabrics in the upper plate of the Resolution Island Shear Zone. Within the Western Fiordland Orthogneiss the alignment of plagioclase, biotite, and hornblende ± garnet defines a composite coarse-grained granulite facies foliation. Igneous banding and gneissic foliations preserved in the Western Fiordland Orthogneiss and within the boundaries of dioritic and gabbroic boudins or dikes that intrude the Deep Cove Gneiss define a composite $S_1$ fabric. The western contact between the Western Fiordland Orthogneiss and the Deep Cove Gneiss is an intrusive contact that locally displays a migmatitic textures including plagioclase leucosome. The Deep Cove Gneiss displays composite upper amphibolite facies gneissic foliations defined by the alignment of biotite, quartz, plagioclase ± garnet ($S_2$) that truncates composite $S_1$ foliations preserved in dismembered lenses or dikes. Gneissic foliations ($S_2$) in the Deep Cove Gneiss form a broad north-northeast trending syncline (Figure 6c, g & 7c). Recrystallized alignment of biotite, hornblende and plagioclase forms northeast- and southwest-trending $L_2$ mineral lineations (Figure 6f). Steep, brittle-ductile strike-slip faults strike north-northeast and cut through all igneous and metamorphic fabrics in the upper plate of the Resolution Island Shear Zone (King et al., 2008).

3.5 Wet Jacket Shear Zone (Domain IV)
The Wet Jacket Shear Zone is a localized, ~2 km thick southwest-dipping zone of ductile fabrics. At the eastern contact of the Western Fiordland Orthogneiss and the Deep Cove Gneiss pervasive gneissic foliations ($S_2$) are refolded into tight recumbent
southwest vergent folds ($F_3$) that define the boundary of the Wet Jacket Shear Zone. Within the shear zone gneissic foliations ($S_2$) are transposed and tightly folded with asymmetric top-down-to-the southwest sense of shear ($F_3$). Tight recumbent $F_3$ folds plunge gently toward the northwest and axial planes ($S_3$) dip gently toward the northeast (Figure 6h). Locally mylonitic top-to-the southwest shear bands occur within the shear zone. Quartz, biotite, and hornblende mineral stretching lineations within Wet Jacket Shear Zone plunge moderately toward the southwest and define $L_3$ (Figure 6f).

Plagioclase grains in the Wet Jacket Shear Zone are micro-faulted and rotated with top-down-to-the southwest shear sense. Asymmetric shear indicators localized in the Wet Jacket Shear Zone include: top-to-the southwest normal displacement C-S fabric in biotite, C’ shear bands in biotite, asymmetric dismembered amphibolite gneissic pods, oblique inclined foliations, and localized (<10 cm) mylonite zones (Figure 11).
**Figure 11** – The Wet Jacket Shear Zone. (a) Form map of a field station in the Wet Jacket Shear Zone. The shear zone is defined by the tight folding (F₃) of pervasive S₂ foliations. Boudins that display S₁ foliations show top-down-to-the southwest asymmetry. (b) Profile view of the Wet Jacket Shear Zone showing the tight folding of S₂ foliation by F₃ folds. Line of section is not shown on map in (a). (c) Schematic block diagram illustrating the relationship between tight F₃ folds and top-down-to-the southwest protomylonite zones (S₃ – dashed lines).
4. Metamorphism

Metamorphic textures preserved in the lower plate of the Resolution Island Shear Zone include feldspar and hornblende pseudomorphs, symplectite rims and coronas. These textures record a retrograde metamorphic pathway from eclogite and garnet-granulite facies mineral assemblages (grt+pl+hbl±ep±cpx) to upper amphibolite facies mineral assemblages (pl+hbl±grt+bt) that are preserved inside penetrative S3/L3 shear zones. In the S2/L2 fabric from the Breaksea gneiss, pyroxene and garnet grains display feldspar and hornblende symplectite rims. Within the boundaries of S3/L3 shear zones pyroxene and garnet pseudomorphs of hornblende and biotite are common. Brittle fractures in eclogite lenses are filled with hornblende and epidote, indicating that they formed at upper amphibolite facies conditions (Figure 12). Pyroxene and garnet replacement textures of plagioclase, hornblende and biotite displayed by eclogite and garnet-granulite facies rocks in S3/L3 shear zones indicate that S3/L3 shear zones formed under upper amphibolite facies conditions (T~500-700 & P~7-12 kbar) (De Paoli et al., 2007).

Thermobarometry conducted by De Paoli et al. (2007) from the lower plate of the Resolution Island Shear Zone indicates that the Resolution Island Shear Zone is the deepest known extensional shear zone in western Fiordland. Samples taken from eclogite facies (om-grt-rt & om-grt-opx), and omphacite-garnet-granulite (om-grt-pl-ky-qtz-rt) rocks exposed within the boundaries of eclogite zones (S1) record metamorphism at P = 17-19 kbar and T= 850°-960°C (De Paoli, 2007). In the Breaksea Gneiss diopside, kyanite, spinel and anorthite symplectite rims on omphacite crystals record garnet-
granulite facies metamorphism ($S_2$) at $P = 12-14$ Kbar, and $T = 650^\circ-780^\circ$ C (De Paoli, 2007). Within penetrative $S_3$ shear zones, breakdown of garnet and omphacite minerals to hornblende, plagioclase, and biotite records medium-$P$ and medium-$T$ epidote-amphibolite facies metamorphism. At amphibolite facies conditions ($P \approx 7-12$ kbar, $T \approx 500^\circ-700^\circ$C) omphacite, rutile, kyanite are consumed, garnet is replaced by hornblende and biotite pseudomorphs, and plagioclase is recrystallized (De Paoli et al., 2007).

Metamorphic conditions in the upper plate of the Resolution Island Shear Zone are also well established (Davids, 1999). Thermobarometry from a sample of Western Fiordland Orthogneiss collected from Wet Jacket Arm suggest garnet-granulite facies metamorphism overprinted igneous pyroxene and hornblende mineral assemblages at $P > 7-9$ kbar and $T \approx 800^\circ$C (Davids, 1999). Similar work on a sample of Supper Cove Orthogneiss reports metamorphism associated with syntectonic intrusion of mid-Cretaceous plutons at $P \approx 6-8$ kbar, and $T \approx 725^\circ$C (Davids, 1999). Davids also reports a localized southwest verging high strain zone with southwest trending mineral stretching lineations where biotite is replaced by chlorite at temperatures $\approx 540-600^\circ$C. Based on the sampling location, and the agreement between the reported temperature ($540-600^\circ$C) and our observations that indicated quartz, biotite and chlorite were dynamically recrystallized, but feldspar grains displayed brittle fractures, we deduce that these temperature constraints describe the Wet Jacket Shear Zone. The large contrast in $P-T$ conditions between the lower and upper plates implies the Resolution Island Shear Zone accommodated significant vertical motion (calculated in Section 6.4).
**Figure 12**—Metamorphic textures in the lower plate. (a) Inside eclogite zones, garnet and omphacite-bearing foliations (S₁) are cross cut by brittle fractures (S₃) filled with amphibole, epidote, and clinozoisite veins. (b) Garnet grains from the S₂/L₂ fabric in the Breaksea Gneiss display clinopyroxene and plagioclase symplectite rims. Note the recrystallized textures in plagioclase grains, these reflect high temperature (>650° C) feldspar creep processes that operated during deformation (Tullis & Yund, 1985). (c) A sample of Breaksea Gneiss deformed by an S₃/L₃ shear zone displays a pyroxene pseudomorph replaced by hornblende and biotite.
a) $S_1$ fabric in eclogite zone

b) Breaksea Gneiss from $S_2/L_2$ fabric

c) Breaksea Gneiss inside $S_3/L_3$ shear zone
5. Strain and kinematic analyses

Structural data show a sequence of fabrics (S1-S3) that evolved in the lower plate of the Resolution Island Shear Zone and culminated in the formation of the Resolution Island Shear Zone. Orientations of L2 and L3 mineral stretching lineations suggest that the dominant transport direction was northeast-southwest (Figure 6e). Broad, gently plunging synforms and antiforms form large-scale dome-and-basin patterns throughout the lower plate of the Resolution Island Shear Zone (Figures 6b, d, & 7a, b). Similarly, at outcrop scales penetrative S3 shear zone foliations anastomose around apparently rigid lenses of material unaffected by S3, and form unique dome-and-basin patterns that are localized into the Resolution Island Shear Zone (Figure 10a, b).

5.1 R∅/θ Method

We used the R∅/θ method (Lisle, 1985) to measure feldspar shape fabrics in six samples from the Resolution Island Shear Zone. We collected samples from a well-constrained exposure of S3/L3 fabric in the Resolution Island Shear Zone (Figure 10). We did not attempt to measure true strains in the shear zone. Instead, we used the strain measurements to determine if anastomosing foliations and lineations (S3/L3) observed in the field track finite strain directions, and determine the type of shape change (i.e. constriction, flattening or plane strain) associated with the dome-and-basin patterns observed in S3/L3 fabric.

Feldspar aggregates were chosen for strain markers because they are apparently the weakest, and most ubiquitous, mineral phase in the Resolution Island Shear Zone. Other phases present, such as amphibole, garnet and pyroxene were considered for strain
analysis but eliminated on the basis that they clearly experienced syntectonic metamorphism. These minerals also occur as unstrained grains or pseudomorphs, suggesting their shapes do not fully reflect $S_3/L_3$ strain shapes. As the weakest phase in the lower crust (Kohlstedt et al., 1995) feldspar preserves the best record of strain conditions during the development of the Resolution Island Shear Zone.

Three perpendicular block sections were cut from each hand sample. Block sections were scanned at high resolution and feldspar grain boundaries were traced with a computer graphics program called $SAPE$ (Mulchrone et al., 2005). This program calculates $R_f$ and $\varnothing$ values from traced grain boundaries in each section (Figure 13a-d). Computer automation of this processes with $SAPE$ allowed us to collect a minimum of 37 grains per sample for best statistical results. We used the $R_f/\varnothing/\theta$ curve fitting method to obtain $R_s$ values with the lowest $\chi^2$ for each section (Lisle, 1985). Using plane strain results from three faces, we calculated 5-6 best-fit strain ellipsoids for each sample with the computer package $Strain3D$ (Yonkee, 1988). This program automates the method for calculating best-fit strain ellipsoids presented by Owens (1984). The magnitudes of the principal axes from the calculated best-fit strain ellipsoids were considered proxies for describing the shape of strain ellipsoids and determine the type of strain that occurred during $S_3/L_3$ deformation.

Two-dimensional strain results on surfaces oriented perpendicular to foliation and parallel to lineation (X-Z plane) show strain ratios ranging from $R_s = 2.55$ to $R_s = 7.00$ (Figure 13d). Strain ratios measured on surfaces oriented parallel to foliation (X-Y plane) range from $R_s = 1.90$ to $R_s = 3.20$. On surfaces oriented perpendicular to foliation and
lineation (Y-Z plane) strain ratios range from $R_s = 1.25$ to $R_s = 2.00$. These results are under estimates of true strains because of the extent of recrystallization of feldspar aggregates. However, our results show that the measurements are sensitive to small variations in strain shapes from different sampling localities.

Calculated strain axes from best-fit strain ellipsoid solutions cluster tightly; this indicates an excellent fit for each sample (Figure 13e-j). Results show long axis (X) orientations trend toward the northeast and southwest and have mean strains that range from 2.08-2.77 signifying extension along the X-axis. Z-axes (poles to foliation) have mean strain magnitudes that range from 0.45 to 0.76, and Y-axis magnitudes range from 0.70-0.91 reflecting shortening along both the Y and Z-axes. Both Y and Z-axes rotate from subvertical to subhorizontal (northwest-southeast) and this pattern records the dome-and-basin geometry of the sampling locality (i.e. Figure 13g & j). Results for all ellipsoids show two shortening axes (Y, and Z), and one prominent extension axis (X) defining prolate strain ellipsoids.

To qualitatively test the shapes and orientations of best-fit results, we plotted strain ellipses that represent the $R_s$ value for each of the sample’s best-fit X-Y planes (parallel to foliation) in their sampling locations according to the trend of the X-axis (Figure 10a). We compared the calculated orientations of X and Z axes with field measurements of mineral stretching lineations and foliations ($S_3/L_3$). Long axes (X) of the best-fit ellipsoids are subparallel to stretching lineations, and short axes (Z) are subparallel to poles to foliation in each sampling locality (Figures 10a & 13e-j). Irregular magnitudes and orientations of best-fit solutions from locality 15e reflect this sample
location in strain shadow adjacent to rigid mafic pods. Consistency between the orientations and shapes of strain results and field observations indicate the best-fit ellipsoids accurately describe $S_3/L_3$ strain shapes. Figure 10a also shows that strain results are sensitive enough to record variation in the orientations and shapes of strain ellipsoids from different regions of the dome-and-basin pattern.

The ratios of strain axes define the shape of the finite strain ellipsoid recorded and are illustrated on a Flinn diagram (Figure 13k). The plane strain field ($R_1/R_2 = R_2/R_3$) falls along a line through the origin with a slope $= 1$. Assuming no volume change during deformation (evaluated in Section 5.2), data that plots above this line deformed under constrictional strains ($R_1/R_2 > R_2/R_3$), and data that plots below deformed under flattening strains ($R_1/R_2 < R_2/R_3$) (Flinn, 1962). All samples have prolate ellipsoid geometries. The $R_1/R_2 > R_2/R_3$ suggests that the $S_3/L_3$ dome-and-basin pattern in the Resolution Island Shear Zone footwall reflect constrictional strains (Figure 13k).
Figure 13 – Finite strain results. (a) Diagram showing the relationship between the aspect ratio (R_i) and orientation (θ) of an undeformed strain marker, and the aspect ratio (R_f) and orientation (∅) of that marker after being deformed by a strain magnitude = R_s (after Lisle, 1985). (b) Photograph of deformed feldspar aggregates that were used for strain markers from a sample of an S_3/L_3 shear zone. (c) An example θ-curve used for finding a best-fit finite strain magnitude. (d) Table of plane-strain results from the XZ, XY, and YZ planes of six deformed samples displaying the number of feldspar grains measured (n), the harmonic mean of R_f, the vector mean of θ, results of the initial symmetry test (I_{sym}), the measured strain ratio (R_s), and the χ^2 value. See Lisle (1985) for an explanation of these terms. (e-j) Orientations and magnitudes (labeled) of principal axes of the finite strain ellipsoids for samples 0715a-f, respectively. See text for details. (k) Flinn diagram showing that all of the samples plot in the field of apparent constriction (after Flinn, 1962). The inset Flinn diagram shows the position of the plane-strain line for a given amount of volume loss during deformation. The shaded box on the constant volume line represents an error margin of ±10% volume change (discussed in Section 5.2).
### X-Z Plane Strain Results

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### X-Y Plane Strain Results

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---

### Diagrams

- Various diagrams illustrating strain results and analysis.
- Graph showing log($S_1$) vs. log($S_2/S_3$) with different markers indicating different samples.
- Scatter plot with axes labeled Constriction and Flattening.

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5.2 Volume change

Volume loss during deformation can remove material from a sample effectively altering the shape of the finite strain ellipsoid (Twiss and Moore, 1992). Structures commonly associated with deformed rocks including cleavage and/or selvages can form by removing material along the XY plane of a strain ellipsoid. If this process occurs, then the results of a three-dimensional finite strain analysis may not reflect true finite strains. For example, Twiss & Moore (1992) suggest that volume loss during deformation changes the position of the plane strain line along the $R_2/R_3$ axis of a Flinn diagram towards the right proportional to quantity of volume lost (Figure 13k).

Volume loss during deformation is commonly associated with a flux of fluids through rock during deformation in the middle and upper crust. It can be distinguished by the presence of veins, cleavage, selvages, and/or leucosome. Although our field observations provided little evidence for volume loss by pressure solution cleavage or the presence of leucosome, we wanted to further test our assumption of no volume change. To test the importance of volume change on our shape fabric calculation we used the Isocon method to quantify volume change during $S_3/L_3$ deformation (Grant, 1986; Srivastava et al., 1990; Barnes et al., 2004; Bailey et al., 2007).

The Isocon method is a graphical method that assumes no mobility of certain elements (i.e. $Al_2O_3$) during deformation. Plotting the weight percent concentrations of major and trace elements allows a comparison of the relative change in concentrations from an undeformed sample (X-axis) and a deformed sample (Y-axis). If the assumption of no mobility of certain elements during deformation is true, than a best-fit line to these data points through the origin defines an isocon, or line of no mass change (Grant, 1986).
The slope of this line is proportional to the amount of volume change according to the equation:

\[
\frac{V_a}{V_o} = \frac{M_a}{M_o} \cdot \frac{\rho_a}{\rho_o}
\]  

(1)

where \( V \) is volume, \( M \) is mass, \( \rho \) is density, ‘\( a \)’ denotes the altered sample (\( S_3/L_3 \) shear zone) and ‘\( o \)’ denotes the host sample unaltered by \( S_3/L_3 \). The mass ratio is the inverse slope of the isocon (Grant, 1986; Barnes et al., 2004).

We compared the relative concentrations of major (\( \text{SiO}_2, \text{TiO}_2, \text{Al}_2\text{O}_3, \text{Fe}_2\text{O}_3, \text{MgO}, \text{CaO}, \text{Na}_2\text{O}_3 \)) and trace (\( \text{Cr, Ni, Y, Zr} \)) elements from two samples of Resolution Gneiss from within, and outside of the Resolution Island Shear Zone, respectively, using X-ray fluorescence data obtained from the New Zealand Institute for Geological and Nuclear Science PetLab database (http://pet.gns.cri.nz). Concentrations in weight percent (major elements) and parts-per-million (trace elements) were plotted on an isocon diagram (Figure 14). Major elements plot very close to the line of no mass change (i.e. line with slope = 1) indicating little mass transfer during deformation in the shear zone. This result supports our petrographic observations of amphibolite facies replacement textures including hornblende, biotite, and plagioclase symplectite rims and pseudomorphs on pyroxene and garnet grains inside \( S_3/L_3 \) shear zones. These textures suggest that most elements were likely to be consumed by major mineral phases associated with amphibolite facies metamorphism in the Resolution Island Shear Zone. A linear best fit through \( \text{TiO}_2, \text{Al}_2\text{O}_3, \text{Fe}_2\text{O}_3, \text{MgO}, \text{CaO} \) yields a slope = 1.016 \((R^2 = 0.985)\) that equates to a 1.16% volume loss. A linear best fit through \( \text{SiO}_2, \text{TiO}_2, \text{Al}_2\text{O}_3, \text{Fe}_2\text{O}_3, \)
MgO, CaO, and Na$_2$O$_3$ yields a slope = 0.965 ($R^2 = 0.990$) indicating 8.59% volume gain (Figure 14). We deduced that scatter in trace element concentrations is a result of sampling locality and do not reflect significant mass transfer during deformation.

Twiss & Moores (1992) conclude that as much as 20% volume loss will shift the plane-strain line marginally to the right (Figure 13k). Our results indicate a range from 1.16% volume loss (when SiO$_2$ and Na$_2$O$_3$ are considered mobile elements) to 8.59% volume gain (if SiO$_2$ and Na$_2$O$_3$ are considered immobile). Considering an error margin of 0 ± 10% volume change during deformation we deduced that the amount of volume change associated with the formation of the Resolution Island Shear Zone is too small to significantly move apparent prolate strains to the oblate strain field. Therefore we conclude that measured prolate-shaped finite strain ellipsoids accurately reflect constrictional strains during deformation (Figure 13k).
Figure 14 – Volume change calculations. (a) Isocon diagram showing the relative concentrations of SiO$_2$, TiO$_2$, Al$_2$O$_3$, Fe$_2$O$_3$, MgO, CaO, Na$_2$O$_3$, Cr, Ni, Y, Zr from an undeformed (x-axis) and deformed (y-axis) sample of Resolution gneiss. The slopes of two potential isocons are shown. See text for details. (b) Table of X-ray fluorescence data from the New Zealand Institute of Geological and Nuclear Science PetLab data-base. Volume change calculations are shown for each of the plotted samples. Densities were measured using a gram-balance and a graduated cylinder. The density average of 7 deformed samples and 4 undeformed samples were used to calculate volume, see text for details.
### Shear Zone in the Resolution Gneiss

**Weight % (major) ppm (trace)**

**Resolution Gneiss Host Weight % (major) ppm (trace)**

![Graph showing data points and regression lines](image)

#### Assume no mobility of Al-Ti-Ca-Fe-Mg (weight % concentrations)

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<th>RNG host</th>
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#### Assume no mobility of Si-Ti-Al-Fe-Mg-Ca-Na (weight % concentrations)

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5.3 Shear sense indicators in the lower plate of the Resolution Island Shear Zone

Kinematic indicators at the outcrop and thin section scales from $S_3/L_3$ fabric include the following: rotated garnet and plagioclase porphyroblasts with asymmetric recrystallized tails, C-S and C’ shear bands, and asymmetric hornblende and plagioclase fish. Thin section scale shear sense indicators on surfaces oriented perpendicular to foliation and parallel to lineation recorded both top-down-to-the northeast, and top-down-to-the southwest normal shear senses. In outcrop, $S_3$ fabrics that dip towards the northeast have top-down-to-the northeast normal shear sense, and surfaces that dip to the southwest have top-down-to-the southwest normal shear sense (Figures 8 & 9). Associated plagioclase and hornblende stretching lineations ($L_3$) plunge to the northeast and southwest, respectively. Apparently symmetrical orientations of top-down-to-the northeast and top-down-to-the southwest normal shears suggest coaxial-dominated deformation during northeast-southwest extension.

5.4 Regional Kinematic analysis

We conducted a bulk kinematic analysis of the entire lower-plate (~15 km$^2$) of the Resolution Island Shear Zone using a graphical method developed by Gapais et al. (1987). This method uses three shear zone descriptors: mineral stretching lineations (L), poles to shear zone foliations (N), and calculated poles to the LN plane (M) to evaluate the symmetry of shear zone patterns. Data are divided into two groups according to shear sense and plotted on an equal area stereonet in the X-Y-Z strain axis reference frame (Figure 15). Once plotted, we assessed the symmetry of opposite shear senses about principal strain axes from the Resolution Island Shear Zone.
Our strain results show that mineral stretching lineations in the Resolution Island Shear Zone ($L_3$) are parallel to the extension direction (northeast-southwest) (Figure 10 & 13). Poles to foliation ($N$) in high strain zones approximate the shearing plane, and $M$-poles track rotations of the plane-of-shear, or shear zone boundaries, and are dependent on both shear direction ($L$) and the shearing plane ($N$) (Figure 15d). Coaxial-dominated deformation produces a symmetric plot about principal strain axes, and non-coaxial-dominated deformation produces an asymmetric plot about principal axes (Gapais et al., 1987).

Predictive diagrams for rocks deformed in constriction fields suggest that the $L$-diagram clusters symmetrically around $X$ and approaches $Y$, and the $N$-diagram will from a girdle through the $YZ$ plane. For areas deformed in flattening fields, predictive diagrams suggest the $L$-diagram will form a wide girdle through the $XY$ plane and the $N$-diagram will cluster strongly around $Z$ (Gapais et al., 1987).

Stretching lineations and three dimensional strain analyses from the Resolution Island Shear Zone indicate a subhorizontal, northeast-southwest-trending maximum finite extension direction ($X$), and two finite shortening axes ($Y$ and $Z$) that rotate from subvertical to subhorizontal. Plots of $L$, $N$, and $M$ cluster around these axes showing that the mineral stretching lineations ($L$), poles to shearing planes ($M$), and poles to planes of shear ($N$) reflect the orientations of principal axes from the finite strain ellipsoid.

Results show that mineral stretching lineations ($L$-diagram) from northeast dipping (black dots) and southwest-dipping (hollow dots) shear zones cluster symmetrically around the $X$-axis and form a girdle pattern in the $XY$ plane (Figure 15). A
girdle of L in the XY plane records both clockwise and anticlockwise rotations of lineations towards the X-stretching axis, indicating a dominant component of pure shear strain in the XY plane as a result of shortening along the Y-axis. Similarly, in the XZ plane the symmetrical distribution of L around X reflects both clockwise and anticlockwise rotations of L with increased strain towards X recording pure shear dominated strain the XZ plane (Figures 15e-g). Both clockwise and anticlockwise rotations of stretching lineations, with increased strain in the XZ and XY planes, indicates a coaxial rotation of stretching lineations towards the direction of maximum finite extension (X-axis).

Poles to S₃ foliation (N-diagram) cluster around the maximum shortening axis (Z) and form a girdle through YZ plane (Figure X). This girdle records a rotation of the maximum shortening directions through the YZ plane, and reflects the prolate or dome-and-basin geometry of S₃/L₃ shear zones (Figure 15b). The symmetry of poles to foliation from both northeast and southwest-dipping shear zones around the Z-axis indicates a component of coaxial shortening along the Z-axis. M-poles track rotations of the shear zone boundaries (plane-of-shear) as N rotates through the YZ plane, and L rotates through the XZ, and XY planes. M-poles scatter symmetrically about the Y-axis as the plane of shear rotates with the shearing plane and mineral stretching lineation. These patterns reflect the rotations of shear zone boundaries, and foliation planes as the planes of shear, and shearing planes rotate around L and N. Coaxial rotations of L and N forms anastomosing, dome-and-basin patterns in S₃ fabric from the Resolution Island Shear Zone.
Results of bulk scale kinematic analyses from the Resolution Island Shear Zone are symmetric about principal strain axes and indicate a coaxial-dominated kinematic history. Coaxial shortening is accommodated along the Y and Z-axes, a pattern that is predicted for areas of constrictional strain and supports our prolate-shaped strain ellipsoid results. These patterns reflect anastomosing, dome-and-basin shear zone geometries around rigid pods that are bounded by top-down-to-the northeast and top-down-to-the southwest extensional shear zones.
Figure 15 – Kinematic analysis results. (a-c) Lower hemisphere equal-area stereonet projections showing the distribution of L, N, and M directions from S3/L3 shear bands in the lower plate of the Resolution Island Shear Zone. Note the view direction is set looking down the Y-axis that plunges 00° toward 117°. (d) Diagram illustrating the relationships between L, N, and M directions and the shearing plane (locally parallel to the foliation plane) and plane of shear (locally parallel to the shear zone boundaries). (e) Diagram showing the rotations of poles to foliation (N) through the YZ plane of the finite strain ellipsoid. (f) Diagram showing the pure-shear rotations of mineral stretching lineations (L) toward the extension axis (X) of the finite strain ellipsoid in the XZ plane, and (g) in the XY plane. See text for details.
6. Discussion

6.1 Fabrics in the lower plate of the Resolution Island Shear Zone

Fabrics preserved in the lower plate record a sequence of deformation that culminated in the formation of the Resolution Island Shear Zone. Recrystallized igneous layering and gneissic banding in eclogite and garnet-granulite zones in the lower plate define composite $S_1$ foliations. These foliations are truncated and attenuated by two phases of younger recrystallized fabrics ($S_2/L_2$ & $S_3/L_3$). The $S_2$ foliation is folded into broad, doubly plunging folds that form a dome-and-basin pattern throughout the lower plate of the shear zone (Figure 6b, d, & 7a, b).

Locally, penetrative upper amphibolite facies extensional shear zones ($S_3$) transpose and attenuate $S_1$ and $S_2/L_2$ fabric. Amphibolite facies $S_3/L_3$ shear zones anastomose around competent lenses of eclogite and garnet-granulite that preserve $S_1$ and $S_2$ foliations, forming a dome-and-basin pattern that resembles the gentle folds of the $S_2$ foliations. Poles to the $S_2$ and $S_3$ foliation form a girdle through the vertical axis of a lower hemisphere equal area stereonet recording symmetrically shaped dome-and-basin geometries of $S_2$ and $S_3$ fabrics (Figure 6b, d, 7a, b, & 10). Mineral lineations from $L_2$ and $L_3$ are parallel to each other, and trend to the northeast and southwest. On the basis of similarities in the dome-shaped folds of $S_2$ and $S_3$ fabric, the parallelism of $L_2$ and $L_3$ mineral stretching lineations, and the attenuation of $S_2/L_2$ fabric into localized $S_3/L_3$ shear zones, we determined that the $S_2/L_2$ and $S_3/L_3$ fabrics record the progressive evolution of extension in the lowermost crust. At this time, $S_3/L_3$ shear zones coalesced to form the
Resolution Island Shear Zone, a large-scale (~1 km thick) lower crustal ductile shear zone.

Calculated pressures and temperatures from eclogite facies to upper amphibolite facies metamorphic fabrics in the lower plate of the Resolution Island Shear Zone record changing metamorphic conditions. High pressures and temperatures \( (P = 17-19 \text{ kbar}, T = 850^\circ-960^\circ \text{C}) \) recorded by early \( S_1 \) eclogite facies rocks suggest that the Breaksea Gneiss and Resolution Gneiss resided in the lowermost crust before the onset of continental extension (De Paoli \textit{et al.}, 2007). \( P = 12-14 \text{ kbar}, \) and \( T = 650^\circ-780^\circ \text{C} \) coincide with the formation pervasive \( S_2 \) garnet-granulite foliations, suggesting that significant vertical displacement accompanied extensional unloading (calculated in Section 6.4) (De Paoli \textit{et al.}, 2007). Coinciding with ductile normal faulting the breakdown of garnet and pyroxene in \( S_2/L_2 \) fabric outside the shear zone to plagioclase and amphibole inside \( S_3/L_3 \) shear zones defines a retrograde path. Prevalence of amphibolite facies \( S_3/L_3 \) shear zones indicates that \( S_3/L_3 \) extension caused further uplift and unloading of the lower plate to upper amphibolite facies conditions \( (P = 7-12 \text{ kbar}, T = 500^\circ-700^\circ \text{C}) \) (De Paoli \textit{et al.}, 2007). This transition from eclogite facies conditions \( (P = 17-19 \text{ kbar}) \) in the lower plate to the formation of a lower-crustal scale shear zone at upper amphibolite facies conditions \( (P = 7-12 \text{ kbar}) \) indicates that extensional unloading was accommodated by the Resolution Island Shear Zone.

6.2 Fabrics in the upper plate of the Resolution Island Shear Zone

The upper plate of the Resolution Island Shear Zone also records an evolution of fabrics and changing \( P-T \) conditions. Composite, recrystallized igneous and gneissic
foliations in the Western Fiordland Orthogneiss and Supper Cove Orthogneiss define the S$_1$ foliation planes. This group of foliations is truncated by a composite upper amphibolite facies gneissic foliation (S$_2$) in the Deep Cove Gneiss that is gently folded into a northeast-trending syncline (Figures 6c, g, & 7c). At the eastern contact of the Deep Cove Gneiss and the Western Fiordland Orthogneiss, S$_2$ foliations are folded by tight northwest-plunging F$_3$ folds that define the Wet Jacket Shear Zone. Mylonitic top-down-to-the southwest shear bands (S$_3$) occur in the Wet Jacket Shear Zone. The appearance of a southwest-plunging mineral lineation in quartz, biotite, and hornblende defines L$_3$ lineations (Figures 6c, f). Asymmetric top-down-to-the southwest kinematic indicators record northeast-southwest extension along the Wet Jacket Shear Zone (Figures 6c, f-h, 7c, & 11).

Davids (1999) reported prograde metamorphism in the Western Fiordland Orthogneiss at $P > 7.9$ kbar and $T \sim 800\, ^\circ$C. We interpret these results as part of the S$_1$ composite granulite facies gneissic and igneous foliations in the Western Fiordland Orthogneiss. Upper amphibolite facies (pl-hbl-bt-qtz±grt) foliations in the Deep Cove Gneiss (S$_2$) truncate S$_1$ granulite facies foliations and suggest that the upper plate was uplifted to upper amphibolite facies conditions ($P = 6.8$ kbar, $T = 500\, - 650\, ^\circ$C) by the time S$_2$/L$_2$ fabrics developed. This interpretation is in agreement with a $P \sim 6.8$ kbar, and $T \sim 725\, ^\circ$C retrograde metamorphic overprint in the Supper Cover Orthogneiss reported by Davids (1999). Within the Wet Jacket Shear Zone, L$_3$ mineral stretching lineations are defined by the recrystallization of biotite, hornblende and quartz. Temperature conditions of $T \approx 540$-600$\, ^\circ$C of biotite altering to chlorite, reported by Davids (1999) from the Wet
Jacket Shear Zone, suggest that $S_3/L_3$ deformation occurred at upper amphibolite facies conditions. Changing pressures in the upper plate during the development of $S_1$-$S_3$ fabric suggest that these fabrics formed during extensional unloading.

6.3 Correlation of fabrics in the lower crust

We correlated deformation sequences in the upper and lower plate of the Resolution Island Shear Zone based on cross cutting relationships, $P$-$T$ conditions of extensional fabrics, and absolute age constraints. Both shear zones attenuate $S_1$ gneissic fabrics from the Western Fiordland Orthogneiss, the Breaksea Gneiss and the Resolution Gneiss. This provides a lower age limit for deformation $\sim 115$ Ma in both shear zones, after the emplacement and metamorphism of these orthogneisses (123-115 Ma) (Hollis et al., 2005; Milan et al., 2007).

Mineral stretching lineations ($L_2$ & $L_3$) from the upper and lower plates of the Resolution Island, and Wet Jacket shear zones are subparallel and trend northeast and southwest. This suggests that the overall tectonic stretching direction remained northeast-southwest during the formation of these shear zones. Shear zone foliations ($S_3$) from the Wet Jacket shear zone dip to the southwest with a top-down-to-the southwest sense of shear. The orientation and sense of shear in the Wet Jacket Shear Zone resembles that in southwest dipping $S_3/L_3$ shear zones from the lower plate of the Resolution Island Shear Zone. Kinematic indicators inside the Resolution Island Shear Zone are down-dip and dominantly plunge toward the east-northeast. Kinematic indicators inside the Wet Jacket Shear Zone plunge down-dip toward the southwest, suggesting that the two shear zones form a synthetic-antithetic pair.
Eclogite facies metamorphism in the Breaksea and Resolution gneisses occurred at \( P \sim 17\text{-}19 \text{ kbar} \) recorded by \( S_1 \) fabrics from the lower plate of the Resolution Island Shear Zone (De Paoli et al., 2007). Granulite facies metamorphism for the Western Fiordland Orthogneiss and Deep Cove Gneiss occurred at \( P > 7\text{-}9 \text{ kbar} \), as recorded by \( S_1 \) composite foliations in the Western Fiordland Orthogneiss (Davids, 1999). We interpret these conditions as the result of emplacement and subsequent metamorphism of the Western Fiordland Orthogneiss, Breaksea Gneiss, and Resolution Gneiss with the Breaksea Gneiss and Resolution Gneiss residing at structurally lower crustal levels than the Western Fiordland Orthogneiss. \( S_1 \) fabrics record the maximum pressure and temperature conditions recorded in the upper and lower plate and are cut by younger extensional shear zones so we consider these fabrics to be pre-extension (\( D_1 \)). Table 1 summarizes the relative sequence of deformational events in the upper and lower plates of the Resolution Island Shear Zone.

High temperatures \((T > 700°C)\) at deep levels in the crust allowed feldspar slip processes to activate in the Breaksea Gneiss and Resolution Gneiss (Figure 12), and the lower crust began to flow under regional extensional tectonic stresses. Changing metamorphic conditions \((P = 12\text{-}14 \text{ kbar}; T = 650\text{-}780°C; \) De Paoli et al., 2007) during the formation of the \( S_2/L_2 \) fabric indicates crustal thinning and marks the onset of continental extension (\( D_2 \)) (Table 1). At this time, upper amphibolite facies gneissic foliations (\( S_2 \)), and a northeast-southwest-trending mineral stretching lineation (\( L_2 \)) developed in the upper plate of the Resolution Island Shear Zone.
Continued regional extension localized strain into S₃/L₃ upper amphibolite facies shear zones in the lower plate. Anastomosing S₃/L₃ shear zones coalesced into the ~1 km thick upper amphibolite facies Resolution Island Shear Zone. At this stage, in the upper plate extension resulted in top-down-to-the southwest asymmetric folding of S₂/L₂ fabric that defines the Wet Jacket Shear Zone. Within the Wet Jacket Shear Zone 1-10 cm-scale protomylonitic shear zones and a southwest-plunging mineral stretching lineation forms S₃/L₃ foliations and lineations. The Wet Jacket Shear Zone and the Resolution Island Shear Zone form a synthetic-antithetic pair. We interpret this phase as the dominant phase of extension when the metamorphic core of the Resolution Island Shear Zone was uplifted to mid-crustal (P ~6-8 kbar; Davids, 1999) levels (D₃) (Table 1).
Table 1 - Correlation of mid-Cretaceous fabrics in the upper and lower plates of the Resolution Island Shear Zone.

<table>
<thead>
<tr>
<th>Tectonic Interpretation</th>
<th>Lower plate of Resolution Island Shear Zone</th>
<th>Upper plate of Resolution Island Shear Zone</th>
<th>Metamorphism</th>
</tr>
</thead>
<tbody>
<tr>
<td>D₃</td>
<td>S₃ Upper amphibolite facies (bt + hbl + pl ± czo) foliations of the Resolution Island Shear Zone. Interpreted to include hydrous fractures (hbl + czo) in eclogite pods that cut S₂ and are wrapped by S₃.</td>
<td>S₃ Tight F₃ folding of S₂ fabric in the Wet Jacket Shear Zone and formation of top-down-to-the SW mylonite zones.</td>
<td>Lower Plate - Amphibolite facies metamorphism and breakdown of peak minerals, grt replaced by hbl+pl, om+rt+ky are consumed, formation of czo+hbl veins. Upper Plate - Deformation temperatures of T ~540°-600° in the Wet Jacket Shear Zone, biotite altering to chlorite.</td>
</tr>
<tr>
<td>D₁</td>
<td>S₁ Includes (a) igneous banding, cumulate layering and (b) recrystallized igneous foliations defined by grt ± opx ± omhp ± hbl in eclogite facies pods in the Breaksea Gneiss.</td>
<td>S₁ Includes igneous banding and recrystallized igneous foliations defined by grt + cpx ± opx ± hbl assemblages in the Western Fiordland Orthogneiss.</td>
<td>Lower Plate - Omphacite garnet granulate metamorphism om-grt-ky-gtz-rt &amp; om-grt-rt &amp; om-grt-opx (Breaksea Gneiss) P ~17-19 kbar &amp; T ~850°-960° C. Upper Plate - Granulite facies metamorphism grt-cpx-opx-hbl (Western Fiordland Orthogneiss) P &gt;9 kbar.</td>
</tr>
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6.4 Estimation of minimum displacement on the Resolution Island Shear Zone

We estimate minimum displacements along the Resolution Island Shear Zone using contrasting pressure conditions in the upper and lower plate that occurred prior to D₂. We assume an average crustal density of 2,800 kg/m³, equivalent to intermediate composition igneous rocks (Robinson & Çoruh, 1988). Eclogite zones displaying S₁ fabric in the lower plate of the Resolution Island Shear Zone record P ≈ 17-19 kbar (De Paoli et al., 2007). This suggests that the lower plate resided at depths ~61-69 km depth before extension. Maximum metamorphic pressures in S₁ fabric from the Western Fiordland Orthogneiss are P ≥ 7-9 kbar (Davids, 1999) suggesting that the upper plate of the Resolution Island Shear Zone resided at shallower crustal depths (~25-32 km) prior to displacement. Using these calculations as starting conditions we estimate that the Resolution Island Shear Zone accommodated a minimum of ~29 km of vertical displacement. Assuming a constant dip of 40° (Figure 7a) this equates to ~45 km of slip along the Resolution Island Shear Zone. This amount of vertical displacement (~29 km) is sufficient to juxtapose lower crustal eclogites (P = 17-19 kbar, ~61 km) against mid-crustal orthogneisses and schist (P ~ 7-9 kbar, ~32 km) in the upper plate. This results also agrees with the pattern of retrograde metamorphism observed in the lower plate from eclogite facies conditions in S₁ fabrics to upper amphibolite facies conditions in S₃/L₃ shear zones. Although these calculations are estimates, the relationships suggest that the magnitude of vertical uplift of the metamorphic core and the displacement along the shear zone resemble those described in Cordilleran metamorphic core complexes (Wernicke, 1992; Johnson, 2006).
6.5 Flow patterns in the lower crust

We suggest that dome-and-basin patterns in the lower plate of the Resolution Island Shear Zone formed as a result of coaxial-dominated constriction during D₃. Mineral stretching lineations (L₃) record northeast-southwest-directed extension in the shear zone and wrapping of S₃ foliation around rigid mafic pods that preserve older S₁ and S₂ foliations forms dome-and-basin patterns (Figure 10). These patterns are present in localized penetrative D₃ shear zones, and in broader map scale folds indicating that this geometry is a scale independent pattern throughout the lower plate of the Resolution Island Shear Zone (Figure 6b & 10). We suggest that dome-and-basin patterns in map scale and outcrop scale S₂/L₂ and S₃/L₃ fabrics indicate two directions of shortening coeval with northeast-southwest extension during the development of the Resolution Island Shear Zone.

Three-dimensional finite strain measurements of S₃/L₃ fabric from the lower plate of the Resolution Island Shear Zone indicate prolate shaped finite strain ellipsoids and confirm that the Resolution Island Shear Zone formed under constriction. Northeast-southwest extension directions (X) recorded by finite strain ellipsoids are sub-parallel to mineral stretching lineations (Figure 10 & 13). Shortening axes (Y & Z) from finite strain ellipsoids indicate both vertical and horizontal (northwest-southeast) components of shortening (Figure 13). Finite strain results that show prolate-shaped strain ellipsoids confirm that deforming by constriction is a mechanism for producing dome-and-basin patterns in the D₃ extensional shear zones. This indicates that deforming by constriction can explain the origin of dome-and-basin patterns in the lower plate of the Resolution Island Shear Zone.
Our kinematic results indicate a type of coaxial-dominated extensional deformation in the Resolution Island Shear Zone. This result is significant in the field of constriction because it indicates that principal strain axes from the finite strain ellipsoid closely approximate the instantaneous strain axes during extension. Pure-shear dominated strain in both the XZ and XY planes of the finite strain ellipsoid was accommodated by simultaneous shortening along the Y and Z-axes and extension along the X-axis. This implies that the three-dimensional kinematics associated with the formation of the Resolution Island Shear Zone formed by constriction, as indicated by our finite strain results. Coaxial-dominated vertical shortening along Z can explain the formation of a symmetrical pattern of top-down-to-the northeast and top-down-to-the southwest shear sense indicators in the Resolution Island Shear Zone, and northeast and southwest-plunging mineral stretching lineations (L$_3$). Simultaneous coaxial shortening along Y can cause extension parallel folding that produces the dome-and-basin patterns documented in S$_3$/L$_3$ shear zones.

6.6 Comparison with other extensional shear zones in western New Zealand

Previous work in western New Zealand documents the existence of at least three other major extensional shear zones attributed to the mid-Cretaceous continental rifting of Gondwana including the Doubtful Sound Shear Zone, the Mount Irene Shear Zone, and the Paparoa metamorphic core complex (Figure 5). The Doubtful Sound shear zone is an upper amphibolite facies ductile shear zone exposed along Doubtful Sound and in Crooked Arm, Fiordland. This shear zone juxtaposes granulite facies ($P>12$ kbar) Western Fiordland Orthogneiss in the lower plate against medium pressure ($P=5-9$ kbar)
metasedimentary rocks in the upper plate (Oliver, 1979; 1980; Gibson et al., 1988; Klepeis et al., 2007). Gibson et al. (1988) first interpreted the lower plate of Doubtful Sound Shear Zone as part of an extensional metamorphic core complex that formed prior to Late Cretaceous rifting of the Tasman Sea (~85 Ma). Klepeis et al. (2007) reported that extension along the Doubtful Sound Shear Zone occurred in two phases, the first active from 114-111 Ma at $P > 12$ kbar, and the second from 111-90 Ma at $P \sim 9-7$ kbar. Klepeis et al. (2007) determined that the Doubtful Sound Shear Zone resides structurally above, and to the north of, the Resolution Island Shear Zone. These authors report pure-shear-dominated vertical shortening associated with northeast-southwest-directed extension, similar to that of the Resolution Island Shear Zone.

Scott & Cooper (2006) describe the Mount Irene Shear Zone in eastern Fiordland as a low-angle extensional mylonitic shear zone active from ~111-108 Ma (Figure 5). Thermobarometry from the upper plate of the Mount Irene Shear Zone records pressure conditions $P \sim 8.5$ kbar in the lower plate and $P \sim 5.9$ kbar in the upper plate during the final stages of extension (Scott & Cooper, 2006). These authors interpret that mid-Cretaceous (~111-108) extension along the Mount Irene Shear Zone contributed to the uplift of the Western Fiordland Orthogneiss in the lower plate from $P \sim 13$ kbar to mid-crustal levels at $P \sim 8.5$ kbar. Scott & Cooper (2006) suggested that uplift and extensional unloading of the lower plate of the Mount Irene Shear Zone is characteristic of a metamorphic core complex. These data indicate that the Mount Irene Shear Zone formed in the middle crust contemporaneously with the Resolution Island Shear Zone.
The Paparoa metamorphic core complex resides in the Westland-Nelson region of New Zealand (Figure 5). During the mid-Cretaceous, before ~480 km of dextral displacement along the Alpine Fault since the Tertiary, the Westland-Nelson region of New-Zealand and the Paparoa metamorphic core complex were contiguous with Fiordland. Tulloch & Kimbrough (1989) describe north-northeast- and south-southwest-trending mineral stretching lineations on two oppositely dipping ductile detachment faults with top-down-to-the northeast and top-down-to-the southwest shear senses, respectively. Spell et al. (2000) reported $^{40}\text{Ar}/^{39}\text{Ar}$ geochronology from the metamorphic core that indicate rapid cooling rates of up to 110°C Myr$^{-1}$ from $T \approx 500°$ to $T \approx 170°$ C between 110-90 Ma. White (1994) reported pressure conditions in the metamorphic core ($P = 5.2 \pm 1.1 – 4.1 \pm 1$ kbar) and suggested that the lower plate of the Paparoa core complex resided at 15-20 km depth prior to extension. Three-dimensional finite strain and kinematic analyses from the Paparoa metamorphic core complex indicate that strain was accommodated by pure-shear-dominated vertical shortening, and is similar in style to that recorded by the Doubtful Sound Shear Zone (Klepeis et al., 2007).

The absolute age of Late Cretaceous continental extension in western Fiordland is well constrained. Both the Resolution Island and the Wet Jacket shear zones attenuate gneissic fabrics from the Western Fiordland Orthogneiss and its components (i.e. the Breaksea Gneiss and the Resolution Gneiss). This provides a lower age limit for deformation (~115 Ma) in the lower crust after the emplacement of the Western Fiordland Orthogneiss between 123-115 Ma (Hollis et al., 2004). A post-extensional
cross cutting pegmatite dike from the Doubtful Sound Shear Zone that yields an age ~90 Ma marks the end of continental extension in western Fiordland (Klepeis et al., 2007).

It is possible to reconstruct the structure of the crust to qualitatively assess the geometry of deformation that occurred during continental rifting. We reconstructed a mid-Cretaceous crustal column using cross sections of western New Zealand extensional structures and peak pressure conditions from the lower plates of these structures prior to extension. By doing so, we evaluate the symmetry of extensional shear zones to determine if the geometry of deformation at crustal scales are consistent with coaxial-dominated constrictional finite strain regimes measured in the lower crust (Figure 16). Generally, the style of extension remains symmetric throughout all levels of the crust, displaying synthetic-antithetic extensional shear zone pairs. This observation agrees with results of pure-shear-dominated vertical shortening reported for both the Paparoa metamorphic core complex and the Doubtful Sound Shear Zone (Klepeis et al., 2007). Scott & Cooper (2006) describe folds on the Mount Irene Shear Zone that trend parallel to the extension direction and a dome shaped geometry of the shear zone surface. This pattern in the Mount Irene Shear Zone suggests folding both parallel to and perpendicular to the extension direction (northeast-southwest). Our results indicate that extension in the lower crust is dominated by coaxial flow, vertical shortening, horizontal extension, and constriction. Symmetry of coeval extensional shear zones and kinematic results that indicate coaxial-dominated extension at lower (Resolution Island Shear Zone), middle (Doubtful Sound and Mount Irene shear zones), and upper (Paparoa metamorphic core complex) crustal levels suggest that lower and upper levels of the crust were
kinematically linked during mid-Cretaceous continental extension. This implies that flow patterns in the lower crust may be transmitted through the crust and control the evolution of extensional shear zones during continental rifting.

This interpretation is consistent with numerical models that predict that the ability of the lower crust to flow may control the geometry of deformation in rifting continental lithosphere (Hopper & Buck 1998; Wijns et al., 2005; Huismans & Beaumont, 2003; 2007; Gessner et al., 2007). These models predict the formation of symmetric rift geometries and metamorphic core complexes in extensional settings dominated by low viscosity lower crust, high rift velocities, and strain softening mechanisms. These models also predict a general kinematic coupling of different layers of the crust during extension. The overall symmetry of deformation in the crust and the presence of metamorphic core complexes displayed in western New Zealand resemble the results of these models. Our results also indicate that styles of deformation in the lower crust can be transmitted to deformation at higher levels of the crust.
**Figure 16** – Cartoon demonstrating the geometry of deformation at four different levels of the crust during the mid-Cretaceous (~115-90 Ma) rifting of Gondwana. The diagram is constructed using pressure conditions documented in the lower plate of each shear zone. The reconstruction suggests several first-order similarities in the style of deformation at all levels of the crust. These include the occurrence of northeast- and southwest-dipping synthetic-antithetic extensional shear zone pairs, the generally symmetric geometry of shear zone orientations and kinematics, and the occurrence of extension parallel to and perpendicular to fold axes that create dome-and-basin patterns in the lower plate of the Resolution Island, Doubtful Sound and Mount Irene shear zones, as well as the Paparoa metamorphic core complex. The bottom diagram illustrates the coaxial-dominated kinematics and dome-and-basin geometry of the Resolution Island Shear Zone. The Wet Jacket Shear Zone resides in the upper plate, and is antithetic to, the Resolution Island Shear Zone. Jagged lines between diagrams mark section breaks between shear zones (see Figure 5 for the map locations of each shear zone). Data controlling the reconstruction of the Doubtful Sound Shear Zone were taken from cross sections drawn by Klepeis *et al.* (2007). Data controlling the geometry of the Mount Irene Shear Zone are after Scott & Cooper (2006). Data for the Paparoa metamorphic core complex are after Tulloch & Kimbrough (1989) and Klepeis *et al.* (2007). Thermobarometry data collected at each of these locations are after \(^1\)White (1994); \(^2\)Scott & Cooper (2006); \(^3\)Klepeis *et al.* (2007); and \(^4\)De Paoli *et al.* (2007).
6.7 Implications for the development of lower crust metamorphic core complexes

We suggest that the presence of a coaxial-dominated constrictional strain field is sufficient to explain observed characteristics of metamorphic core complexes, including the presence of folds on the detachment surface that trend both parallel to and perpendicular to the extension direction and the dome-shaped geometry of many metamorphic core complexes. Both top-down-to-the northeast and top-down-to-the southwest kinematic indicators and the symmetric geometry of the Resolution Island and Wet Jacket shear zones suggest that a component of vertical shortening controlled the development of these structures. Our kinematic and strain results that indicate the Resolution Island Shear Zone formed under coaxial-dominated deformation and vertical shortening. However, kinematic and strain results also show that the Resolution Island Shear Zone formed under constriction, and that coaxial shortening occurred perpendicular to the dominant extension direction. This result is compatible with the observation of dome-and-basin patterns formed by doubly plunging synforms and antiforms in the lower plate of the Resolution Island Shear Zone.

Metamorphic core complexes are commonly described as being dome-shaped (Anderson, 1982; Wernicke, 1992; Johnson, 2006). In the Mohave metamorphic core complex, California and in core complexes exposed in the Scandinavian Caledonides, authors have reported the development of folds with axes that trend both perpendicular to and parallel to the extension direction (Fletcher & Bartley, 1994; Chauvet & Seranne, 1994; Osmundsen et al., 2003). Folds that trend perpendicular to the extension direction are commonly associated with the isostatic uplift of the unloaded footwall block (i.e. Fletcher & Bartley, 1994). However, there is some debate about whether the formation of
folds that trend subparallel to the extension direction formed during extension or record changing strain conditions during deformation. Fletcher & Bartley (1994) measured constrictional finite strains from the Mojave metamorphic core complex detachment fault in California and document the existence of extension-parallel folds on the detachment surface. These authors attribute the extension-parallel folds on the detachment surface to the constrictional strain field and suggest that they formed during extension. They argue that an extensional uniaxial stress field would result in constrictional strains that produce the observed fold geometries and strain results. They also use paleostress indicators to support this interpretation. Our results indicate that constrictional finite strains during the development of the Resolution Island Shear Zone formed a dome-and-basin pattern in the $S_3/L_3$ shear zones. Our results are compatible with those of Fletcher & Bartley (1994). We suggest that extension under a constrictional regime may be one way to form dome shaped patterns in metamorphic core complexes and account for both extension parallel and perpendicular folds on the detachment surface.

Metamorphic core complexes reported in the Scandinavian Caledonides provide similar evidence for both extension perpendicular and extension parallel folding of the detachment surface (Chauvet & Seranne, 1994; Osmundsen et al., 2003). Chauvet & Seranne (1994) use paleostress indicators to track the rotation of $\sigma_1$ from subvertical to subhorizontal as the footwall block is unloaded. In these examples, extension parallel folding is attributed to the transition from an extensional tectonic regime to one of transtension and supported by the presence strike-slip faults that cross cut the detachment. Although it may be possible to rotate $\sigma_1$ to horizontal if the footwall block is
uplifted toward the surface, we determined that \( \sigma_1 \) likely would remain vertical due to the overburden at the deep crustal levels (~61-69 km) of the Resolution Island Shear Zone. Furthermore, we determined that flow during D\( _3 \) was coaxial-dominated implying no rotation of the instantaneous strain axes during deformation. We suggest that dome-and-basin patterns in the Resolution Island Shear Zone are a result of a coaxial-dominated constrictional deformation and not rotations of the overall principal stresses.

**7. Conclusions**

Field-based results describe the progressive evolution (D\( _2 \)-D\( _3 \)) of extensional shear zones in lower crust that culminated in the formation of a 1-km thick ductile detachment fault called the Resolution Island Shear Zone. This shear zone juxtaposes high-pressure (\( P \sim 17-19 \) kbar) eclogite and garnet-granulite facies rocks exposed in the lower plate against upper amphibolite and granulite facies (\( P > 7-9 \) kbar) paragneiss and orthogneiss in the upper plate, resulting in the formation of a lower crustal metamorphic core complex. We estimate that the Resolution Island Shear Zone accommodated ~29 km of vertical displacement. An antithetic extensional shear zone called the Wet Jacket Shear Zone formed contemporaneously in the upper plate of the Resolution Island Shear Zone during D\( _3 \) extension. Coaxial-dominated vertical shortening during extension resulted in the symmetric conjugate-style geometry of these shear zones.

Three-dimensional finite strain analyses from the Resolution Island Shear Zone indicate extension occurred in the field of finite constriction. Geochemical calculations from shear zone rocks show <10% volume change during deformation. Bulk kinematic analyses from \( S_3/L_3 \) fabric in the lower plate (~15 km\(^2\)) describe coaxial-dominated flow
of material during the formation of the Resolution Island Shear Zone. Orientations of
S_3/L_3 fabric and calculated finite strain ellipsoids indicate northeast-southwest finite
extension coeval with subvertical and subhorizontal (northwest-southeast) shortening.
Deformation under the field of coaxial-dominated finite constriction resulted in the
formation of dome-and-basin patterns displayed by S_2/L_2 and S_3/L_3 fabrics. This strain
field was characterized by both top-down-to-the northeast and top-down-to-the southwest
kinematic indicators in the upper and lower plates of the Resolution Island Shear Zone.

Deformation under the field of coaxial-dominated constriction is an efficient
mechanism for creating symmetric dome-shaped metamorphic core complexes. The field
of constriction involves two directions of shortening perpendicular to the extension
direction that can explain the dome-shaped geometry of metamorphic core complexes
and the origin of the fold interference patterns characteristic of metamorphic core
complex detachment surfaces.

The Resolution Island Shear Zone and Wet Jacket Shear Zone both attenuate
gneissic fabrics from ~115 Ma plutons indicating that they formed approximately
contemporaneous with at least three other extensional shear zones in western New
Zealand during the mid-Cretaceous rifting of Gondwana (~114-90 Ma). High-pressure (P
~17-19 kbar) eclogite and garnet-granulite facies rocks exposed in the lower plate of the
Resolution Island Shear Zone indicate that this shear zone is the deepest of all extensional
shear zones exposed in Fiordland. A crustal-scale reconstruction of the Resolution Island
Shear Zone and three other extensional shear zones suggests similar coaxial-dominated
kinematic styles and a symmetric geometry of deformation characterized the rifting of
continental crust. The apparent kinematic links between deformation in the lower crust and deformation at higher crustal levels resembles the results of numerical models that predict the ability of the lower crust to flow is a primary control on the evolution of a rifted margin.
Chapter V: Discussion

1. Thesis summary

Extensional shear zones and metamorphic core complexes that formed during the mid-Cretaceous rifting of Gondwana at upper, middle, and deep levels of the crust are well exposed in Fiordland, New Zealand. This unique window into different levels of the crust provides an ideal field setting for understanding how the crust behaves during continental extension. The recent discovery of high-pressure ($P \sim 17-19$ kbar; De Paoli et al., 2007) eclogite and granulite facies rocks in the footwall of the Resolution Island Shear Zone make this shear zone the deepest known structure associated with the mid-Cretaceous rifting of Gondwana. The goals of this research were to describe the three-dimensional evolution of the Resolution Island Shear Zone, compare the Resolution Island Shear Zone with other contemporaneous extensional shear zones in Fiordland, and to determine the processes that resulted in the formation of a lower crustal metamorphic core complex.

In this thesis I have presented structural, strain and kinematic data that describe the three-dimensional evolution of fabric in the Resolution Island Shear Zone. High-pressure ($P \sim 17-19$ kbar; De Paoli et al., 2007) gneissic foliations associated with peak metamorphism of the Breaksea and Resolution Gneiss are truncated and attenuated by two phases of extensional fabric. The onset of extension ($D_2$) in the lower plate of the Resolution Island Shear Zone is marked by pervasive garnet-granulite facies foliations that truncate gneissic banding. As extension continued these fabrics are localized into
penetrative 1 m to 100 m thick upper amphibolite facies ductile shear zones. Penetrative extensional shear zones anastomose around the boundaries of rigid lenses that preserve earlier fabric and form a dome-and-basin pattern in the lower plate. Toward the northeastern side of Resolution Island penetrative upper amphibolite facies shear zones coalesce into the Resolution Island Shear Zone forming a low-angle ductile detachment fault (Figure 6 & 7a). At this stage (D₃), the Wet Jacket Shear Zone formed in the upper plate antithetic to the Resolution Island Shear Zone. During D₃, the Resolution Island Shear Zone accommodated approximately 45 km of displacement resulting in 29 km of vertical uplift and the juxtaposition of high-grade eclogite and granulite facies gneiss from the lower crust against mid-crustal upper amphibolite facies rocks. Extension along the Resolution Island Shear Zone resulted in the formation of a metamorphic core complex.

One major result of this study is a determination of the three-dimensional finite strain and kinematics of flow from the Resolution Island Shear Zone. Finite strain results indicate that the Resolution Island Shear Zone formed by constriction. This result agrees with field observations of dome-and-basin shaped shear zone geometries and suggests that two directions of shortening (Y & Z) were coeval with horizontal stretching (X). Coaxial-dominated kinematics measured in the Resolution Island Shear Zone show the symmetric rotations of mineral stretching lineations away from the shortening axes of the finite strain ellipsoid (Y & Z) toward the extension direction (X). Coaxial-dominated flow in the field of constriction resulted in a large-magnitude normal fault and a dome-shaped metamorphic core complex. My research suggests that the deformation in the
field of constriction is a sufficient mechanism for forming dome-shaped detachment faults. This result compliments other field-based studies of metamorphic core complexes by providing a mechanism for explaining the formation of dome-shaped structures and fold axes on detachment surfaces that form parallel to the extension direction.

Understanding the kinematics and geometry of deformation in the Resolution Island and Wet Jacket shear zones has implications for the regional geologic setting in Fiordland. Thermobarometry by De Paoli et al. (2007) indicates that the Resolution Island Shear Zone is the lowest of several extensional shear zones in Fiordland attributed to the rifting of Gondwana. I compared the apparent kinematic coupling of deformation in upper and lower levels of the crust during this phase of continental extension by drawing a schematic reconstruction of four extensional shear zones exposed in Fiordland. The reconstruction shows a general symmetry of shear zones and suggests that deformation in the lower crust was kinematically linked to deformation at higher levels in the crust. The kinematics and geometry of deformation in the lower crust appear to be transmitted to coeval shear zones at higher crustal levels and resulted in the formation of metamorphic core complexes at several different levels of the crust. This interpretation is consistent with the results of numerical models by Huismans & Beaumont (2007) and Gessner et al. (2007) that predict a symmetric geometry of deformation throughout lithosphere will occur during continental extension in settings where the lower crust can flow. This interpretation is also consistent with the results of models by Buck (1991), Hopper & Buck (1996), and Wijns et al. (2005) that predict the formation of
metamorphic core complexes during continental extension when the lower crust is weak and able to flow.

2. Future Work

My work suggests that it is possible to form large-magnitude normal faults associated with metamorphic core complexes by coaxial-dominated constriction. Assuming no volume change, in the realm of coaxial-dominated constriction, extension along X can be accommodated by coaxial shortening along both Y and Z so that each contribute to the extension factor along X. Contrasting, in plane-strain, extension along X is only proportional to shortening along Z. In the flattening field shortening along Z can be accommodated by extension along both X and Y, thus reducing the total magnitude of extension along X. Our kinematic results document that the pure-shear extensions of X in both XY, and XZ planes was accommodated by shortening along Y and Z, respectively. If it is possible to form a large-magnitude extensional shear zone by crustal thinning in field of coaxial-dominated constriction then it is tempting to suggest that the dome-shaped geometries, large magnitude displacements, and evidence for vertical and horizontal shortening displayed by some metamorphic core complexes may be a result of a constrictional strain field. I suggest conducting three-dimensional strain and kinematic analyses in other field settings to test this hypothesis.

This research describes a general kinematic coupling of deformation between the lower crust and upper crust during mid-Cretaceous rifting of Gondwana. Further work in Fiordland could test this interpretation by providing three-dimensional strain data from the Doubtful Sound, and Mount Irene shear zones and the Paparoa metamorphic core
complex. This type of study could test whether finite strains remained linked throughout
the crust during continental extension or if apparent finite strain fields changed at
different levels of the crust. One direction for approaching this task could be to quantify
volume change associated with deformation in each of these shear zones. Fluid fluxes
associated with pressure-solution cleavage, or percolation of meteoric fluids common in
the middle and upper crust alter the shape of finite strain markers. If this occurs then one
might expect the field of deformation to change from apparent constriction in the lower
crust to apparent flattening in the upper crust as a result of volume change by the removal
of material during deformation.


Appendix A

Methods for using SAPE and Strain3D

In this section I explain a streamlined process for simplifying the measurement of a large quantity of strain markers to produce a complete and statistically significant result. I employed four separate computer programs and used them in conjunction to automate the majority of a typically tedious data collection process.

Calculating strain ellipses

I used the program Semi-Automated Parameter Extraction for strain analysis (SAPE) to generate best-fit ellipses for each grain, and calculate the aspect ratio \((R_f)\) and orientation \((\varphi)\) of each ellipse (Mulchrone et al., 2005). SAPE can generate a database of the magnitudes and orientations of \(R_f\), \(\varphi\), and the principal axes of a strain ellipses from a bitmap image. The bitmap image is produced by tracing grain boundaries with any computer illustration program (Adobe Illustrator™, or Adobe Photoshop™ work well).

Below I describe a stepwise process for using SAPE:

1) Obtain a copy of the above reference and software (free by contacting Kieran Mulchrone), read thoroughly!

2) Choose a sampling strategy. Decide which strained objects to measure, what questions to answer, and which scale will best provide the necessary information to answer these questions. For example, it may be better to scan a
cut slab from a sample to measure strain markers, than to look at deformed grains in a thin section. This depends on the size of the strained objects.

3) Generate a digital image of the sections to measure. If using a thin section, a photomicrograph works well, be sure to use proper light (i.e., cross polarized or plane polarized) to highlight the mineral phases to measure. It is very important to ensure that the photomicrograph is of the appropriate scale to capture a sufficient sample of strained objects. If a cut section from a hand sample is the best way to measure strained object, scanning the hand sample with a high-resolution scanner works well. Use a wet cloth (water works best) and wipe down the cut section to remove dust and to contrast the different minerals. Place the wetted sample onto a glass plate (a grain mount plate works well) to keep the sample from drying out and to avoid scratching the scanner glass. Scan the sample at a high resolution (600 dpi+) so that the image can be enlarged without being pixilated. Mineral oil also works for dampening the samples to contrast the colors, however it tends to darken the sample, making it difficult to see some strain markers. Also, the oil will sink into the pore space of the rock and become impossible to remove.

4) Import the digital image into a vector program (Adobe Illustrator™). Rotate the image so that the mean stretching axis (x-axis) is closest to horizontal. The SAPE program measures ∅ angles to a horizontal reference line, so it is
important to align the samples best as possible to this orientation. Zoom in close to each of the strain markers and ensure that they can be identified and easily traced. Using the pencil tool, trace all of the strain markers in the object. If possible attempt to collect at least 50 markers per section, per sample (for more information see Lisle, 1985). Mulchrone et al. (2005) suggests 150 markers are required. When finished tracing grains, highlight all tracings and set the line color to black, and 2 point width. Be sure all grain tracings are closed, and that they do not intersect. Also, be sure only the new tracings are visible on the screen (i.e. there should be no border, or text, SAPE will assume that every image is a strain marker). Hide the original image, and export the tracings to a black and white bitmap file (.bmp) with a one-bit pixel depth.

5) Open SAPE (must use a PC with a Windows™ operating system) and load the new bitmap file. The image should display on the screen, if it does not display be sure the file is a simple black and white 1-pixel bitmap. Under the process menu select “set parameters.” This window allows the selection of a percentage of samples that will be analyzed (according to area), and how SAPE will interpret the line widths. I have found that it is best to leave the bottom tail at 1% because SAPE will often generate several small erroneous ellipses between adjacent grains. Set the top tail at 0% if certain that all data should be included (this can be adjusted later). The shrink and grow numbers
refer to the way that SAPE will modify the bitmap lines to interpret ellipses. If all of the ellipses are closed and the line width was set to 2 point width then that both of these parameters can be set to “0”. Read the instructions for a much more detailed discussion of these parameters (Mulchrone et al., 2005).

6) Run the SAPE program. All measured ellipses will appear turquoise, and space between them will appear yellow. Carefully look over the results and be sure that SAPE did not generate extra ellipses (especially if the shrink and grow values are >1). Each measured ellipse is numbered. If an ellipse has a white number the data will be exported (next step), if and ellipse has a red number this data will be omitted. Simply clicking on the number will change its status. By using the view menu it is possible to observe how SAPE fit ellipses to the tracings and view the list of data that will be exported.

7) Export Data. Setting the view menu to “list of extracted data” will allow you to observed the measured data. Values with “not to be processed” next to the data correspond with ellipses that were marked as red, and will be omitted when you export the data. Using the File -> export data option, you may choose the types of data you wish to export. For strain analysis I prefer to export the $R_f/\varnothing$ data and the option “all data.” This will give you values of $R_f$ and $\varnothing$ for each ellipse as well as values for the principal axes (a & b are the notations used by SAPE) of the ellipse.
8) Manipulate data and calculate statistics. The “all data” option exports a .sda file and the “Rₘ/phi only” option exports a .rfp file. Both of these files can be opened with Microsoft Excel™ on a PC. In Microsoft Excel™ (or a similar spreadsheet program) organize a spreadsheet to calculate the recommended statistical tests including the harmonic mean, vector mean, and symmetry test. Use the equations provided by Lisle (1985) to make columns for each of these values. It is easy to copy and paste equations in a column to quickly solve for the whole dataset. Use the harmonic mean value as an estimate of Rₑ to begin curve fitting in the following steps.

9) Plot data: Combine the data into one well-organized spreadsheet, and use the graph tool to make an Rₘ/∅ plot. The Y-axis of the plot will be a log scale, and have a 12.5 cm cycle. The x-axis will be ∅, with 1 cm for every 10 degrees (i.e. 18 cm for 180 degrees). I’ve found the best way to match these dimensions is to use the rectangle tool in the drawing toolbar of excel and draw a rectangle of the said dimension, then simply scale the plot to fit the rectangle. Check after making the plot that the dimensions are indeed 18 cm X 12.5 cm so that they will match the Rₘ/∅ curves provided by Lisle (1985).

10) Overlay plots onto theta curves using either a light table, or by scanning the curves into Adobe Illustrator™. By generating an Illustrator™ file with a
layer for each of the curves, and copying the $R_i/\varnothing$ plot for each sample into their own layers it is easy to turn on and off curve layers to estimate a best fit. Lisle (1985) suggests using the value of the harmonic mean as a starting point for selecting $R_s$ curves. Once a best-fit curve is found, print and point count to calculate the $\chi^2$ statistic for curve fitting (Lisle, 1985). Try curves with $R_s$ values slightly larger and smaller to find the lowest $\chi^2$ fit.

**OR**

11) Use the *Excel spreadsheet for finite strain analysis* to automate the statistics and curve fitting processes (Chew, 2003). This process is fast, efficient, and extremely accurate. The program requires input in the form of long and short axis magnitudes ($a$ & $b$ from a SAPE “all data” export) and $\varnothing$.

Instantaneously the program will calculate the harmonic mean, vector mean, and $I_{sym}$ statistics according the equations provided by Lisle (1985). After selecting a range of potential $R_s$ values, the *Excel*™ macro will calculate $\chi^2$ for a range of $R_i/\theta$ curves over any increment desired using the equations that relate $R_i$, $\varnothing$, $R_s$, $\theta$ and $R_S$ (Lisle, 1985). I highly recommend this program for performing a large number of strain analyses. However I also recommend going through the process manually a few times to be sure that the process is fully understood by the user.
Generating a best-fit strain ellipsoid

I used the computer software *Strain3D* (Yonkee, 1988) to calculate best-fit strain ellipsoids for each sample. *Strain3D* automates the calculations of the orientations and magnitudes of principal axes of a least-squared best-fit strain ellipsoid using the method presented by Owens (1984). Owens (1984) provides a mathematical solution for a best-fit strain ellipsoids to at least three intersecting (i.e. perpendicular) strain ellipses given the strike and dip of a plane, and the pitch of the long axis of an ellipse on that plane and the axial ratio of the ellipse. Here I present instructions for correctly preparing samples and measurements for using *Strain3D* to calculate three-dimensional strain ellipsoids from field samples, and evaluate the results. For a detailed discussion of the mathematical theory behind this solution see Owens (1984).

1) Software: Obtain a copy of *Strain3D* (Yonkee, 1988) or similar software.

2) Collect and tabulate input data, including the strike and dip of principal planes where strain ellipses were measured, the pitch of the long axis and aspect ratio ($R_s$) of strain ellipse from each of these planes. Two-dimensional strain measurements must be made on at least three, intersecting planes for each sample. Perpendicular planes work best (Owens, 1984; Yonkee, 1988). Two right-hand-rule (RHR) conventions apply for data entry into *Strain 3D* (Yonkee, 1988). While measuring strike and dip, the thumb should be aligned down dip with the index finger pointing toward the strike azimuth. For pitch, the thumb should be aligned down the pole to the plane with the index finger
measuring down from horizontal to the long axis of the strain ellipse (a range of 0°-180° clockwise from horizontal). To collect this information, “re-orient” samples to their true positions using a clay block and their original orientation markings. Mark horizontal on each plane with a pen. Measure the strike and dip of each plane (using the above RHR for strike and dip). Measure the pitch by drawing a line on each plane that is parallel to the long axis of the finite strain ellipse for that plane (recall how planes were oriented when collecting data using SAPE). Measure the pitch using a protractor and the above RHR convention.

3) Calculate ellipsoid parameters: enter the above information into Strain3D for each sample and run the program. Remember that it is critical to use the above conventions for the data or the results will be erroneous.

4) Evaluate results: Strain3D will calculate at least six best-fit solutions for every dataset. Comparing results provides a qualitative way to assess the precision of the solutions. Consistent results between each of the six outputs records the most precise solution. If results vary widely then it is likely that some of the data were entered or collected incorrectly (Yonkee, 1988). By comparing the orientations and magnitudes of the principal axes of the calculated ellipsoid it is possible to assess the accuracy of the solution. For example, if the calculated strain ellipsoid accurately records three-dimensional deformation in
the rock then the long axis should be parallel to the stretching direction, and
the short axis should be normal to the foliation plane. By plotting orientations
and magnitudes of these axes on a stereonet it is possible to compare the
results to field measurements and make an assessment of their quality.
Unfortunately, no formal treatment of errors associated with the best-fit
ellipsoid exists (Owens, 1984).