Title: The interplay and effects of deformation and crystallized melt on the rheology of the lower continental crust, Fiordland, New Zealand

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Abstract

Microstructural, electron backscatter diffraction (EBSD), and misorientation analyses of a migmatitic granulite-facies orthogneiss from the exhumed lower crust of a Cretaceous continental arc in Fiordland, New Zealand show how deformation was accommodated during and after episodes of melt infiltration and high-grade metamorphism. Microstructures in garnet, omphacite, plagioclase, and K-feldspar suggest that an early stage of deformation was achieved by dislocation creep of omphacite and plagioclase, with subsequent deformation becoming partitioned into plagioclase. Continued deformation after melt infiltration resulted in strain localization in the leucosome of the migmatite, where a change of plagioclase deformation mechanism promoted the onset of grain boundary sliding, most likely accommodated by diffusion creep, in fine recrystallized plagioclase grains. Our results suggest three distinctive transitions in the rheology of the lower crust of this continental arc, where initial weakening was primarily achieved by deformation of both omphacite and plagioclase. Subsequent strain localization in plagioclase of the leucosome indicates that the zones of former melt are weaker than the restite, and that changes in deformation mechanisms within plagioclase, and an evolution of its strength, primarily control the rheology of the lower crust during and after episodes of melting and magma addition.

1. Introduction

Determining the mechanisms by which upper crustal extension and the extensional collapse of thick (≥ 65 km) orogenic crust are accommodated in the deep crust is an important aspect of the study of continental deformation. In particular, the lower crust plays a key role in this process because viscosity and density anomalies within this layer greatly influence how the deformation

initiates and evolves (Wijns et al., 2005; Tírel et al., 2008; Rey et al., 2009; 2011). For example, the presence of hot, partially molten material at depth has been linked to the lateral flow of the deep crust over large distances, and has helped explain variations in crustal thickness, orogen width, topography, and crustal structure (e.g., Xie et al., 2004; Gerbault et al., 2005; Jamieson et al., 2011). Studies of gneiss domes and metamorphic core complexes further indicate that vertical flow, and the exhumation of deep crustal material, is closely linked to both these lateral displacements and a reduction of lower crustal viscosity by magmatism and partial melting (Kruckenberg et al., 2011; Betka and Klepeis, 2013; Whitney et al., 2013; Platt et al., 2015).

Although the presence of partially molten material commonly is cited as an important mechanism of crustal weakening (Vanderhaeghe, 2009; Kruckenberg et al., 2011; Jamieson et al., 2011; Tikoff et al., 2013), the close spatial and temporal association of melt production, melt migration and crystallization, and deformation can make it difficult to interpret how these processes influence one another in many natural settings (e.g., Brown and Solar, 1998; Rosenberg and Handy, 2005; Vanderhaeghe, 2009; Whitney et al., 2013; Sawyer, 2014). Nevertheless, recent work using mineral microstructures from rocks that were once partially molten have provided important new information on how melt achieves the weakening of lower crustal materials, and how deformation and melt migration are interrelated (e.g., Závada et al., 2007; Sawyer, 2014; Hasalová et al., 2015).

For this study, we investigated deformation mechanisms and microstructures in plagioclase-rich leucosome and garnet-omphacite restite from a migmatitic high-strain shear zone in an exhumed section of lower crust at Breaksea Entrance in Fiordland, New Zealand. This section records the

localization of strain into shear zones following a period of Early Cretaceous (123-115 Ma) magmatism, granulite-facies metamorphism, and the formation of migmatitic gneiss domes (De Paoli et al., 2009; Betka and Klepeis, 2013; Klepeis et al., 2016). The abundance of plagioclase-rich leucosome within the high-strain zone suggests that the behavior of plagioclase is crucial for understanding how strain localized within the lower crust. Elsewhere, previous work suggests that this process could occur by the transition from a grain size insensitive deformation mechanism, such as dislocation creep, to a grain size sensitive mechanism, such as grain boundary sliding accommodated by diffusion creep (e.g., Gomez Barreiro et al., 2007; Precigout and Hirth, 2014; Czaplinska et al., 2015). We tested this possibility using both field and microstructural observations in the Fiordland example. Our goals included: (1) to compare microstructures of the leucosome and restite to interpret how strain localizes within the leucosome of migmatitic granulite gneiss, (2) to evaluate the evolution of plagioclase microstructures and deformation mechanisms within the leucosome, and (3) to determine how this evolution influenced strain localization processes and lower crustal rheology.

2. Geologic Setting

Our study site is located in a part of the lower crustal section exposed in western Fiordland that records Early Cretaceous magmatism at the root of a long-lived arc near the coast of Gondwana (Figure 1a). Over 250 Ma of subduction and arc magmatism collectively formed the Median Batholith (Mortimer et al., 1999), and culminated in the emplacement of the Western Fiordland Orthogneiss (WFO) (Bradshaw, 1990; Allibone et al., 2009a). This latter batholith is composed of several large intrusions of mostly dioritic to monzodioritic composition (Allibone et al., 2009b). Its emplacement within a contractional regime (Daczko et al., 2001; Klepeis et al.,

2004; Klepeis et al., 2016) over a relatively short (125-115 Ma) period of time (Mattinson et al., 1986; Tulloch and Kimbrough, 2003; Hollis et al., 2003; 2004; Sadowski, 2015) helped build a >65 km-thick crust (Figure 1b) (De Paoli et al., 2009; Clarke et al., 2013).

By ~106 Ma, arc magmatism and crustal thickening gave way to crustal thinning and horizontal extension (Bradshaw, 1989; Mortimer, 2008; Tulloch et al., 2009; Klepeis et al., 2016; Schwartz et al., 2016), marking the onset of deformation that led to orogenic collapse and Gondwana break-up. The horizontal extension was accommodated by a series of upper amphibolite-facies shear zones in the lower crust (Gibson et al., 1988; Klepeis et al., 2007; 2016; Schwartz et al., 2016) and by metamorphic core complexes and other extensional features within the middle and upper crust (Tulloch and Kimbrough 1989; Spell et al., 2000; Scott and Cooper, 2006; Kula et al., 2007). Our study area is located within a portion of the WFO exposed at Breaksea Entrance (Figure 2) that lies below the Resolution Island Shear Zone (Figure 2). This shear zone records extension during an early stage of rifting that led to Gondwana breakup (Betka and Klepeis, 2013).

In the study area (Figure 2), the WFO is composed mainly of two distinctive rock units. The oldest is the ~123 Ma Breaksea Orthogneiss (Klepeis et al., 2016), which consists of light-colored monzodioritic and monzogabbroic rock interlayered with dark layers (S_1) of garnet pyroxenite, harzburgite, and hornblende peridotite (De Paoli et al., 2009; Clarke et al., 2013). This unit was emplaced at depths of >65 km (1.8–2.0 GPa; De Paoli et al., 2009) and later juxtaposed against the Paleozoic Deep Cove Gneiss by the Resolution Island shear zone (Betka and Klepeis, 2013). The ~115 Ma Resolution Orthogneiss intruded the Breaksea Orthogneiss at

depths ranging from 40 to 50 km (~ 1.8 GPa) and is composed mostly of hornblende diorite (De Paoli et al., 2009; Allibone et al., 2009b; Hout et al., 2012). Both units record pulses of penetrative deformation, high-grade metamorphism, and crustal melting that accompanied and outlasted emplacement of the Resolution Orthogneiss. The deformation culminated in the formation of a series of kilometer-scale gneiss domes that record the vertical (diapiric) flow of hot, partially molten lower crust at the deepest levels of the arc (Betka and Klepeis, 2013; Klepeis et al. 2016). We focused our work on one well-exposed gneiss dome (Figure 2) where microstructures in a migmatitic high-strain zone permit interpretation of the mechanisms that aided strain localization following magmatism during the last stages of contraction affecting the lower crust.

The gneiss domes at Breaksea Entrance are defined by concentric orientations of high-strain zones and penetrative, migmatitic gneissic foliations (S_2) (Betka and Klepeis, 2013). Subdomes that lie within larger gneiss domes are common (Figure 2). In outcrops of Breaksea Orthogneiss, the high-strain zones deform and transpose an old igneous banding (S_1) defined by variations in the modes of plagioclase, garnet, and omphacite pyroxene (De Paoli et al., 2009; Clarke et al., 2013). High strains are indicated by tight folds of S_1 ; large (1-2 cm) strain shadows composed of plagioclase, K-feldspar and quartz (leucosome) around mafic clusters (Figures 3a and 3b); and asymmetric shear bands filled with leucosome. The collection of leucosome in strain shadows around mafic aggregates and in shear bands suggests subsolidus deformation immediately followed melt crystallization within the lower crustal section. Geochronologic data from both zircon and garnet within the high-strain zones are consistent with simultaneous melt infiltration and deformation (see Klepeis et al., 2016 and Schwartz et al., 2016 for discussion). Outside the

strain shadows, S_2 foliation wraps disaggregated garnet-pyroxenite blocks, forming high-strain zones where S_2 is characterized by partially to completely recrystallized plagioclase ribbons and stretched garnet-pyroxene clusters (Figures 3c). High strains around these blocks are indicated by increases in the lengths of strain shadows and in the aspect ratios of garnet-pyroxene clusters, and decreases in the grain size of recrystallized plagioclase. High-strain zones in the Resolution Orthogneiss show dismembered dikes and layers (S_1) of hornblende-rich metagabbro within a locally migmatitic metadioritic host (Figure 3d).

Low strain areas of the Breaksea Orthogneiss commonly preserve delicate migmatitic structures that are absent in high-strain areas. These include layers of disaggregating garnet-pyroxenite surrounded by plagioclase-rich leucosome (Figure 3e). Fractures that penetrate the mafic layers are filled with plagioclase-rich material derived from the matrix, indicating the mafic layers broke apart within a partially molten host (arrows, Figure 3e). Leucosome in low-strain areas adjacent to garnet-pyroxene blocks (Figure 3c) also forms angular shapes with many branching apophyses that are absent in the high-strain zones. Plagioclase is coarse and shows less grain-size reduction by crystal-plastic deformation mechanisms compared to the high strain areas, suggesting that these low-strain areas preserve an early part of the history as melts moved through and crystallized within the lower crust.

For our microstructural and EBSD study, we selected a sample (0656) of migmatitic omphacite granulite (Figures 3a and 3b) from a high-strain zone that surrounds a subdome within a larger (~2 km diameter) gneiss dome on Resolution Island (Figure 2) where thermobarometric calculations of P-T conditions during granulite metamorphism range from 1.96 $\pm$ 0.23 GPa and 895 $\pm$ 100°C to 1.80 $\pm$ 0.22 GPa and 897 $\pm$ 105°C (DePaoli et al., 2009). The high-strain zone

cuts through exposures of both the Breaksea and Resolution orthogneisses, indicating that it formed after the emplacement of these two units. The sample is an L-S tectonite from a fragment of the former unit that was intruded by the latter and subsequently became entrained within the high-strain zone. A thin section was cut from a surface oriented perpendicular to foliation (S_2) (162 45 SW) and parallel to a plagioclase and pyroxene mineral lineation (L_2) (295 36). In outcrop, foliation planes are defined by stretched and flattened garnet-omphacite pyroxene clusters that are enveloped by leucosome composed of plagioclase and K-feldspar with minor quartz (Figures 3a, 3b).

3. Methods

3.1 Light microscope petrography

The petrographic microscope was used to identify the most abundant mineral phases, characterize their microstructures, and select areas for EBSD analyses.

3.2 EBSD

The sample was ultrapolished in preparation for EBSD analyses using a Buehler Minimet®1000 grinder-polisher. The thin section was polished for an hour in each of four polishing fluids, including 6- μm , 3- μm , and 1- μm diamond suspension solutions, and colloidal silica solution.

EBSD analyses were used to 1) confirm mineral identifications made by petrographic observation, 2) reveal detailed microstructures, 3) measure grain sizes and grain diameters, 4) measure the crystallographic orientation of mineral grains, and 5) generate misorientation data. Energy Dispersive Spectroscopy (EDS) data were collected simultaneously with EBSD data to

aid in phase identification and data processing. EBSD and EDS data were collected at California State University Northridge using an FEI Quanta 600 SEM equipped with an Oxford Instruments Nordlys EBSD detector, an Oxford Instruments INCA x-sight EDS detector, and the Oxford Instruments AZtec EBSD and EDS acquisition software. Data were collected from an uncoated sample under low vacuum conditions (20 Pa H₂O) to minimize charging.

A total of four EBSD maps were acquired using both Large Area Mapping (LAM) and conventional beam mapping techniques. We used LAM to create three EBSD maps: a 15,000 µm x 4000 µm map targeting coarse-grained garnet, omphacite, and plagioclase microstructures, a 1200 x 1200 µm map targeting a plagioclase porphyroclast and the rim of a omphacite porphyroclast, and a 1200 µm x 1600 µm map targeting coarse- and fine-grained plagioclase microstructures. We used conventional beam mapping to make an inset map of an inclusion within a garnet grain featured in the first large area map. All SEM working conditions and EBSD acquisition parameters used in the collection of the maps are summarized in Table 1.

3.2.1. EBSD Data processing

EBSD map data were processed using the Oxford Instruments Channel 5 software suite. Noise reduction was completed with the Tango program to remove erroneous data points in all of the maps. Noise reduction included removal of wild spikes and iterative extrapolation of zero solutions that were surrounded by 8 and 7 neighbor-pixels of the same phase. Additional and more specialized post-processing was completed for two large area maps. Following noise reduction, systematic misindexing was removed within plagioclase, biotite, and quartz in the 15,000 µm x 4000 µm large area map. In the 1200 µm x 1600 µm large area map, EDS K Kα

map data were used to identify K-feldspar grains that were misindexed as plagioclase, and subsetting was used to change the erroneous plagioclase grains to K-feldspar. Systematic misindexing was removed within orthoclase, plagioclase, biotite, and quartz.

3.2.2. Display of EBSD data

EBSD map data are displayed using the Oxford Instruments Channel 5 Tango program. We generate band contrast maps to evaluate the quality of the indexing, phase maps to see the distribution of minerals within a map, and EDS element maps to see the distribution of a given element within the map area. In the maps, we classify grain boundaries as having $\geq 10^\circ$ of misorientation between adjacent grains, and we designate low-angle subgrain boundaries as having between 2 and 10° of misorientation between adjacent parts of a grain.

Crystallographic orientation data are displayed in contoured pole figures created with the MTEX toolbox for MATLAB (Bachmann et al., 2010). In the pole figures, data are displayed in the kinematic reference frame in lower hemisphere, equal area projections. Contours of the pole figures are measured in multiples of uniform distribution (m.u.d.). MTEX-generated contoured pole figures were calculated with an Orientation Distribution Function (ODF) using a de la Vallee Poussin kernel and a halfwidth of 10° . The strength of the fabric in the pole figures was measured with MTEX using the M-index (Skemer et al., 2005) and the J-index (Bunge, 1982). We use the convention (*hkl*) to indicate a specific crystallographic plane and [*uvw*] to indicate a specific crystallographic direction.

3.2.3 Misorientation data generated from EBSD data

We display the orientation of plagioclase misorientation axes to aid in the identification of deformation mechanisms. We show the orientation of misorientation axes relative to the crystal reference frame (i.e., in inverse pole figures), and we also display the orientation of misorientation axes in the kinematic reference frame (i.e., in pole figures) to see how they are oriented relative to foliation and lineation. We display misorientation angles in a misorientation angle distribution histogram where the relative frequency of occurrence is plotted against the angle of misorientation as measured in degrees. In the histograms, misorientation angles are binned into 5° increments, and the minimum misorientation angle is 3° to minimize errors associated with measurement at very low angles (e.g., Prior et al., 1999).

4. Results

4.1 Microstructures from light petrography

Petrographic and microstructural observations reveal that the major mineral phases in the orthogneiss are garnet, omphacite, and plagioclase, and that the latter two minerals display evidence of intragrain crystal plasticity. Garnet grains are moderately coarse-grained (~500-1000 µm diameter), and have irregular and embayed shapes (Figure 4a), but there are also some very coarse-grained (~2000 µm diameter) idiomorphic grains. Due to the isotropic nature of garnet, no intragrain microstructures can be observed optically. Omphacite forms large asymmetric porphyroclasts (~4000 µm diameter) that are partially recrystallized into coarse grains (~100-1000 µm diameter), showing rounded to sinuous grain boundaries (Figure 4b), and some subgrain development and sweeping extinction. The omphacite porphyroclasts are rimmed by fine-grained material in which it is difficult to identify specific minerals and their microstructures (Figure 4c). Plagioclase forms rare porphyroclasts, and where present, the

porphyroclasts have well-developed subgrains at their rims (Figure 4c). The porphyroclasts are mantled by small, recrystallized grains with sinuous grain boundaries. Aside from the rare porphyroclasts, most plagioclase forms a fine-grained (~10-40 μm diameter) matrix around the coarse-grained garnet and omphacite grains (Figure 4d). The fine-grained plagioclase has irregular and sinuous grain boundaries indicative of dynamic recrystallization.

4.2 Microstructures from EBSD

The large EBSD map reveals the presence of a strong gneissic foliation (S_2) that is marked by alternating aggregates of plagioclase and omphacite, with garnet distributed within both plagioclase- and omphacite-rich aggregates, and K-feldspar concentrated within the plagioclase-rich aggregates (Figure 5a). We concentrate on the microstructures of these four minerals because 1) they are the four most abundant phases, and 2) the four phases enable characterization of the relationships between the most abundant minerals in the leucosome (plagioclase and K-feldspar) and the more refractory components (garnet and omphacite) of our migmatite sample.

The irregular and equant shaped garnet grains are similar in that they show little evidence for recrystallization, but are different in terms of inclusion content. Both the irregular and equant grains have very rare low-angle boundaries, and are coarse, single grains without smaller grains developed along their rims or in grain interiors, indicating a lack of appreciable recrystallization and crystal-plastic deformation. These garnet microstructures are also apparent from an additional orientation map, backscattered electron (BSE) images, and EDS element maps shown in Appendix 1. The equant grains tend to have quartz inclusions (bottom right hand corner of Figure 5a), whereas inclusions in the irregular-shaped grains are rare, but where present (white

arrows in Figure 5a), are composed of quartz, K-feldspar and/or plagioclase (Figure 5b). In the inclusion, the K-feldspar grain is untwinned, irregularly shaped with some cusped edges, and it cuts across plagioclase twinning (Figure 5b). EDS element maps of all garnet grains (Appendix 1) show minor and gradational zoning where the outer ~50 μm of the rims are enriched in Fe and depleted in Mg regardless of grain shape; there are no zoning patterns that correlate with the very rare low-angle boundaries, nor are there sharp zoning boundaries within other parts of grains.

The omphacite grains show evidence of moderate crystal-plastic deformation. The porphyroclasts are partially recrystallized and form elongate, asymmetric shapes. Most of the recrystallized grains have not been incorporated and mixed into the matrix of recrystallized plagioclase grains, although there is evidence for minor mixing based on the presence of interstitial phases (see omphacite grain in center of Figure 5a). Many recrystallized grains within the porphyroclasts have irregular and sinuous grain boundaries, and others have substantial subgrain development. Within the porphyroclasts, there are garnet grains that are interstitial to the recrystallized grains (bottom left corner of Figure 5a), and in some omphacite porphyroclasts, larger (~1000 μm) elongate garnet grains are found along porphyroclast rims. The fine-grained material along the omphacite porphyroclast rims that is difficult to optically identify is shown to be an intergrowth of secondary sodic pyroxene, plagioclase, hornblende, and biotite that is a retrograde symplectite texture (Figure 5c).

Plagioclase grains show evidence of extensive crystal-plastic deformation associated with dynamic recrystallization. The plagioclase is mostly recrystallized, forming a fine-grained matrix with few relict porphyroclasts that are in various stages of recrystallization. The plagioclase

grains of the matrix have a pronounced reduction in grain size relative to the relict porphyroclasts and relative to the recrystallized omphacite grains. The plagioclase matrix grains contain interstitial grains of K-feldspar, quartz, biotite, kyanite and various accessory minerals, some of which are elongate in shape. Porphyroclasts are rare (top left and center right parts of Figure 5a), and the largest porphyroclast (Figure 5c) is elongate and has substantial subgrain development along its rims. The subgrains are similar in size to some adjacent recrystallized grains.

K-feldspar is closely associated with plagioclase and shows very little evidence for crystal-plastic deformation. K-feldspar is restricted to the interstitial spaces between both coarse- and fine-grained plagioclase and, in places, appears to form inclusions within coarse-grained plagioclase (Figure 5a). The K-feldspar grains have variable grain size, ranging from about 50 μm to 250 μm in diameter. The coarser grains have some subgrain development, but there is no core and mantle structure. The K-feldspar grains are irregularly shaped and sometimes elongate, with some curvature to the grain boundaries between adjacent K-feldspar grains. These microstructures are also evident from BSE images of K-feldspar that are in Appendix 2.

4.3 Pole Figures for garnet, omphacite, plagioclase, and K-feldspar

The crystallographic orientations of garnet grains are shown as scatter plots in pole figure space (Figure 6a) because there are too few grains to contour, which also precludes measurement of LPO strength using the J-index and M-index. However, observation of the garnet pole figures does not reveal any apparent clustering of grain orientations, which indicates the absence of LPO.

The omphacite pole figures show a non-random and moderate LPO pattern (Figure 6b). A strong maximum of poles to (100) is oriented perpendicular to foliation. A girdle of poles to (010) is oriented highly oblique to the foliation, and a broad maximum of [001] is offset $\sim 20^\circ$ counterclockwise from the lineation. The J- and M-index values of 4.0 and 0.05, respectively, indicate a moderate strength LPO.

The plagioclase pole figures show a non-random and weak LPO pattern (Figure 6c). There are weak maxima of poles to (010) oriented sub-perpendicular to foliation, with a slight sinistral offset. The *c*-axes show two sets of maxima; one set is offset $\sim 20^\circ$ counterclockwise from the lineation, and the other is offset $\sim 20^\circ$ clockwise from the foliation normal. A weak LPO is indicated by the J- and M-index values of 2.4 and 0.02, respectively.

The plagioclase data collected from the map shown in Figure 5a were subsetted based on grain size, and the crystallographic orientation of coarse and fine grain size fractions were plotted separately. We define coarse- and fine-grained plagioclase as having grain areas of $> \sim 10,500 \mu\text{m}^2$ (greater than 36 pixels) and $< \sim 10,500 \mu\text{m}^2$, respectively. This threshold value is the minimum area of a subset of grains showing porphyroclast microstructure. The coarse-grained plagioclase (Figure 6d) has a LPO pattern that is almost identical to the bulk plagioclase LPO, but with maxima that are far more clustered relative to the bulk plagioclase LPO (Figure 6c). A strong LPO is indicated by the J- and M-index values of 12.0 and 0.11, respectively. In contrast, the fine-grained plagioclase (Figure 6e) has a weak LPO that bears slight resemblance to the bulk plagioclase LPO (Figure 6c). The fine-grained plagioclase displays a weak maximum of poles to

(010) oriented perpendicular to foliation at the bottom of the pole figure, and the [001] axes form a hint of a girdle sub-parallel to the primitive circle in the bottom right portion of the pole figure (Figure 6e). The near-absence of LPO is indicated by the J- and M-index values of 1.4 and 0.02, respectively.

The K-feldspar pole figures show a non-random and weak LPO pattern that is not aligned with the kinematic reference frame. There are two maxima of [100] that do not align with lineation, a maximum of poles to (010) that does not align with the foliation normal, and several small clusters of [001] that do not align with the lineation. A weak LPO is indicated by the J- and M-index values of 2.5 and 0.14, respectively.

4.4 Inset EBSD Map of coarse- and fine-grained plagioclase in leucosome

We made a high resolution inset EBSD map (location shown in Figure 5a) to investigate the microstructures of coarse- and fine-grained plagioclase in the leucosome, and to examine the microstructural relationship between plagioclase and K-feldspar within the leucosome (Figure 7a). We made a corresponding set of EDS element maps to confirm phase indexing (Appendix 3). An excerpt of the EBSD phase map, with a corresponding excerpt of the EDS K K α map, demonstrates that the two phases are compositionally distinct and that K-feldspar has been properly indexed (Figure 7b). BSE images of selected K-feldspar-plagioclase microstructures in Figure 7 are shown in Appendix 2.

The inset map reveals that K-feldspar is present in the interstitial spaces of the coarser-grained plagioclase grains and in the interiors of those grains, but it is entirely absent from the fine-

grained plagioclase regions (Figure 7a). The central corridor of the map where K-feldspar is present is also distinctive in that it has coarser-grained quartz, more abundant kyanite, and an apparent paucity of biotite, omphacite, and hornblende. Where K-feldspar is present, the coarser K-feldspar grains have cusped grain boundaries with very low dihedral angles, and the finer K-feldspar grains appear as irregular-shaped inclusions within plagioclase grains. Both coarse- and fine-grained K-feldspar grains are untwinned and cut across tapered and curved twins in plagioclase (white arrows in Figure 7a). Low-angle boundaries are present in K-feldspar, but not abundant; when present, they are commonly in coarser grains.

4.5 Pole figures of plagioclase in inset EBSD map

We subset the map into two regions, one in which plagioclase is coarser-grained and the other in which plagioclase is finer-grained, and we display the crystallographic orientation of the plagioclase in each region in pole figures. In the coarser region (avg. grain diameter of 20.7 μm) (Figure 7c), the plagioclase grains have a non-random and moderately strong LPO indicated by J-index and M-index values of 8.3 and 0.07, respectively (Figure 7d). A broad cluster of [100] is oriented close to the center of the pole figure. Poles to (010) form two sets of maxima; one set is highly oblique to the foliation, and is oriented $\sim 20^\circ$ counterclockwise from the foliation normal, and the second set is nearly perpendicular to the first, with maxima along the primitive circle. The [001] axes form broad maxima along the primitive circle, with the densest clustering nearly parallel to lineation.

In the finer-grained region of plagioclase (avg. grain diameter of 13.9 μm) (Figure 7e), the plagioclase grains have a non-random and moderate LPO where the J-index and M-index values

are 3.4 and 0.04, respectively (Figure 7f). The [100] axes form maxima that are nearly parallel to the foliation pole. Poles to (010) form a cluster, but it is not oriented parallel to the foliation pole. The [001] axes form a hint of a girdle that contains a broad maximum along the primitive circle, where the maximum is within 20° counterclockwise of the lineation.

4.6 Misorientation axes of plagioclase in coarse- and fine-grained regions of inset map

In the region of coarse-grained plagioclase, there is both clustering of plagioclase misorientation axes in the crystal reference frame and in the kinematic reference frame. The misorientation axes cluster about [-100] and [001] in the crystal reference frame, both of which lie within the trace of the (010) plane (Figure 8a). In the kinematic reference frame, there is a maximum that is offset down and to the left of the Y strain axis (Figure 8b), which is most consistent with low-angle misorientations being tilt-type (e.g., Figure 3 of Kruse et al., 2001). The neighbor-pair misorientation angle distribution deviates from random for angles between 3 and 30°, with the highest values of relative frequency for angles <10° (Figure 8c). The high frequency of angles <10° indicates that there is a greater number of neighbor pairs with similar orientations relative to a randomly oriented population of grains, which is typical for subgrain formation.

In the region of fine-grained plagioclase, there is clustering of plagioclase misorientation axes in the crystal reference frame, but a random orientation of plagioclase misorientation axes in the kinematic reference frame. Similar to the coarse-grained region, the misorientation axes cluster about [-100] and, to a lesser degree, [001], both of which are contained within the trace of the (010) plane (Figure 8d). In the kinematic reference frame, there is a diffuse and irregular-shaped region of higher multiples of uniform distribution (m.u.d.) values, but no clear maximum (Figure

8e). The maximum m.u.d. value in this plot is lower relative to the m.u.d. for the coarse-grained plagioclase, indicating a much weaker fabric. Similar to the coarse-grained region, the neighbor-pair misorientation angle distribution deviates from random for angles between 3 and 30° and there is a peak at angles <10°, but there are much higher values of relative frequency associated with the 10-30° range of angles relative to the coarse-grained region.

5. Discussion

5.1 Microstructural evidence of melt crystallization in leucosome

Our microstructural observations confirm that the plagioclase and K-feldspar minerals that comprise the leucosome of our sample crystallized from melt. The K-feldspar inclusions within the coarse-grained plagioclase grains (Figures 7a, 7c) represent an antiperthitic igneous texture where a high-temperature, disordered ternary feldspar crystallizes from melt and then exsolves untwinned K-feldspar into irregular-shaped blebs upon cooling, resulting in an albitic plagioclase host (e.g., Hubbard, 1965; Menegon et al., 2013). The antiperthite textures in the sample are consistent with those that have been identified in plagioclase-rich migmatitic components of the Breaksea Orthogneiss (DePaoli et al., 2009; Clarke et al., 2013; Chapman et al., 2015). The coarse K-feldspar grains along the grain boundaries of coarse-grained plagioclase (Figure 7a) could either represent an aggregation of antiperthite resulting from strain in the host albitic plagioclase (Hubbard, 1965), or are remnants of the crystallized melt. Both K-feldspar and quartz have very low dihedral angles and cusped grain shapes in the vicinity of the coarse-grained plagioclase (Figure 7a), which are traits that are commonly used to interpret the former presence of melt in pore spaces and along grain boundaries (Holness and Sawyer, 2008).

The microstructures also demonstrate that some of the garnet in the leucosome is peritectic and grew in the presence of melt. Garnet growth in the presence of melt is demonstrated by the presence of plagioclase-K-feldspar-quartz inclusions within garnet (Figure 5b), as these “nanogranites” are evidence of anatectic melt preservation (Cesare et al., 2009). The nanogranite inclusion is inside an elongate, irregularly shaped garnet grain within the plagioclase-rich leucosome, which may suggest that the other irregularly shaped and embayed garnet grains in the leucosome may be peritectic in origin, too. Similarly shaped garnets are observed in granulites with limited amounts of trapped hydrous melts, and are interpreted as having interacted with melt (Holness et al., 2011).

5.2 Deformation in the presence of melt

The microstructures are consistent with field observations of the high strain migmatite that indicate deformation in the presence of melt. Deformation of the restite is implied by the high aspect ratio of garnet-omphacite clusters, and deformation is inferred in the leucosome based on the grain size reduction of plagioclase. We interpret the microstructural evidence for how this deformation is accommodated in the paragraphs below.

5.2.1. Deformation of refractory garnet-omphacite clusters

The elongation of garnet-omphacite clusters that comprise the refractory part of the migmatite is achieved by the partial recrystallization of omphacite, but not of garnet. Recrystallized omphacite grains with irregular and sinuous grain boundaries and others with substantial subgrain development indicate a moderate amount of crystal-plastic deformation. The high proportion of coarse porphyroblast grains relative to small, recrystallized grains suggests that

recovery was limited, but the moderate LPO pattern (Figure 6b) is consistent with the activity of at least one slip system. In contrast, garnet does not show any evidence for recrystallization, despite the prevalence of irregular, embayed grain shapes that could suggest grain boundary mobility. The paucity of low-angle boundary development (Figure 5a) and lattice distortion (Appendix 1) indicates that garnet grains are relatively unstrained. Further, the presence of a random LPO pattern (Figure 6a) is consistent with a lack of crystal-plastic deformation and slip system activity in garnet. EDS map data (Appendix 1) do not show sharp compositional zoning patterns indicating that dissolution-precipitation creep processes contributed to the elongation of grains, either (Alvarez-Valero et al., 2005). Given the minimal deformation of garnet, we focus on omphacite as the rheology-controlling mineral of the restite.

The partially recrystallized omphacite grains have a LPO pattern that is consistent with two known slip systems in clinopyroxene. The LPO pattern is consistent with either the (100) [001] or (010)[001] slip systems (Figure 6b). The (100)[001] slip system has been recognized in high temperature (800-900°C), single crystal experiments on diopside, and rarely in naturally-deformed clinopyroxene (Bascou et al., 2002; Brenker et al., 2002). However, where the '(100) fabric' is found in naturally-deformed clinopyroxene, it is associated with omphacite in eclogites (Philippot and Van Roermund, 1992). The most common omphacite slip system is (010)[001], where variations in the geometry of the LPO have been interpreted as an indication of strain asymmetry; a girdle of (010)-poles normal to the foliation and a strong concentration of [001] axes along the lineation are attributed to a constrictional fabric (Ábalos, 1997; Bascou et al., 2002), which is consistent with observations of L>S tectonites in the field. We prefer the latter

interpretation of slip system owing to its more frequent recognition in naturally deformed, high-grade, granulite- and eclogite-facies metamorphic rocks.

5.2.2. Deformation of plagioclase-rich leucosome

The deformation of the leucosome is primarily achieved by extensive crystal-plastic deformation and dynamic recrystallization of plagioclase. Plagioclase is the most abundant phase in the leucosome and displays microstructures indicative of crystal-plastic deformation, whereas the far less abundant and dispersed K-feldspar exhibits microstructures that are inconsistent with significant crystal-plastic deformation. Rare plagioclase porphyroclasts with well-developed subgrains and core-and-mantle structure indicate both crystal-plastic deformation of porphyroclasts and recovery through a subgrain rotation recrystallization mechanism of dynamic recrystallization (Figure 5c). The prevalence of the fine-grained recrystallized matrix and the paucity of porphyroclasts suggest that a significant amount of grain size reduction was achieved through subgrain rotation recrystallization. Dynamic recrystallization promoted the development of a LPO pattern (Figure 6c) that is most consistent with the (010)[001] slip system, which is the 'easy slip system' found in plagioclase deformed at high temperatures ($>700^{\circ}\text{C}$) (Olsen and Kohlstedt, 1984; Ji and Mainprice, 1987; Montardi and Mainprice, 1987; Kruse et al., 2001; Stünitz et al., 2003; Mehl and Hirth, 2008; Svahnberg and Piazzolo, 2010).

In contrast, the K-feldspar microstructures are inconsistent with dynamic recrystallization because the irregular grain boundaries are neither lobate nor bulging, which indicates limited grain boundary mobility, and no core-and-mantle structure is present (Figure 5a, Figure 7a). The lack of dynamic recrystallization microstructures coupled with a LPO that is not aligned with

kinematic reference frame (Figure 6f) together preclude deformation by a slip system. Instead, the small isolated grains are likely due to exsolution from a ternary feldspar, with the moderate LPO developing due to multiple exsolved K-feldspar grains taking on the crystallographic orientation of their host plagioclase (e.g., Hubbard, 1965). Given the near-lack of deformation microstructures in K-feldspar, we necessarily focus on plagioclase as the rheology-controlling mineral of the leucosome.

5.2.3. Rheology-controlling minerals of the restite and leucosome

Omphacite and plagioclase accommodate deformation of the restite and leucosome, respectively, but strain is partitioned differently between the minerals. Omphacite is partially recrystallized and displays a LPO consistent with known slip systems, indicating that dislocation creep was active during deformation of the restite. However, the lack of dynamic recrystallization microstructures and the presence of symplectites along grain boundaries may indicate that recrystallization was induced by differences in chemical potential between grains, which has been interpreted as strong driving force in addition to strain energy (e.g., Stünitz, 1998). In contrast, plagioclase is nearly completely recrystallized and displays both distinctive dynamic recrystallization microstructures and a LPO indicative of a high-temperature slip system, all of which are consistent with deformation by dislocation creep. The partial recrystallization of omphacite due to a combination of both strain and chemical driving potentials suggests that less deformation is accommodated in it relative to plagioclase where recrystallization appears to be nearly complete and primarily strain induced because no major compositional differences are suggested by the EDS data (Appendix 3). This partitioning could be explained by the greater strength of water-depleted restite relative to hydrolytically weakened leucosome, as has been

demonstrated by other studies (Menegon et al., 2011; Diener and Fagereng, 2014). On the basis of these relationships, we interpret plagioclase as the weaker of the two minerals. We focus on its deformation mechanisms because the abundance of plagioclase in the leucosome exerts a primary control on how strain localizes within the lower crust.

5.3 Deformation mechanisms of plagioclase

The plagioclase dynamic recrystallization microstructures and LPO patterns suggest initial deformation by dislocation creep where grain size reduction is achieved by subgrain rotation recrystallization. Both the coarse- and fine-grained plagioclase of the large map (Figure 5a) show a similar LPO pattern consistent with the (010)[001] slip system (Figures 6d and 6e), though the fine-grained pattern is far more dispersed relative to that of the coarse grains. The similarity in LPO patterns indicates a ‘host control’ relationship on the orientation of the fine-grained plagioclase, which can be achieved by a subgrain rotation recrystallization mechanism (e.g. Kruse et al., 2001). The same recrystallization mechanism and slip system can also be interpreted from the misorientation data. The misorientation data show high relative frequencies of low angle misorientations ($<10^\circ$) suggesting subgrain development in both the coarse- and fine-grained fractions (Figures 8c and 8f). The coarse- and fine-grained plagioclase have similar misorientation axis orientations in the crystal reference frame (compare Figures 8a and 8d), implying deformation by a common slip system in both the fine- and coarse-grained fractions. The tilt-boundary nature of the misorientations and the locations of both [-100] and [001] within the trace of the (010) plane are consistent with slip systems involving (010) as a glide plane, corroborating the (010)[001] slip system interpreted from the bulk plagioclase LPO pattern.

Though the fine-grained plagioclase matrix was largely produced by subgrain rotation recrystallization during dislocation creep, the weakness of the fine-grained plagioclase LPO patterns suggests subsequent modification of grain orientations by a second deformation mechanism. Though the coarse- and fine-grained plagioclase of the large map have similar LPO patterns (compare Figures 6d and 6e), the fine-grained plagioclase LPO pattern is very diffuse, with a weak LPO (Figure 6e). A similar weakening of the LPO in the fine-grained fraction is observed in the plagioclase of the inset map (compare Figures 7d and 7f). Randomization of LPO patterns is suggested by the faint girdle of [001] axes in both sets of fine-grained plagioclase patterns (Figure 6e and 7f). A weakening or randomization of LPO with decreasing recrystallized grain size is commonly associated with the onset of the grain boundary sliding (GBS) deformation mechanism in both naturally and experimentally deformed plagioclase (e.g., Law, 1990; Zhang et al., 1994; Raimbourg et al., 2008; Pearce et al., 2011; Getsinger and Hirth, 2013; Okudaira et al., 2015), and we attribute the more randomized LPO patterns in our sample to the onset of GBS.

Additional evidence for the onset of GBS in fine-grained plagioclase comes from the misorientation data associated with coarse- and fine-grained parts of the inset map (Figures 7 and 8). Though the misorientation data show that both coarse- (Figure 8a) and fine-grained plagioclase (Figure 8d) have similar misorientation axis orientations in the crystal reference frame, the misorientation axis orientations in the kinematic reference frame are markedly different for the coarse- (Figure 8b) and fine-grained (Figure 8e) fractions. The more diffuse pattern of misorientation axes for the fine-grained plagioclase indicates that these grains became more randomly oriented subsequent to deformation by subgrain rotation recrystallization

associated with dislocation creep. In addition, the relative frequency of neighbor-pair misorientations with angles ranging from 10-30° is greater for the fine-grained plagioclase relative to the coarse-grained plagioclase (compare Figures 8c and 8f). These observations are consistent with the characteristics of GBS, as grain switching and grain rotation can produce more randomized orientations of grains (e.g., Zhang et al., 1994), resulting in more high-angle boundaries between adjacent grains (Jiang et al., 2000; Bestmann and Prior, 2003; Svahnberg and Piazzolo, 2010). The randomization of misorientation axes in the kinematic reference frame and an increase in the number of high-angle (>15°) misorientations between recrystallized grains together distinguish deformation by GBS from deformation by subgrain rotation recrystallization (Bestmann and Prior, 2003), and we use these observations to infer that GBS modified the microstructures developed by dislocation creep.

Our samples provide microstructural evidence for a transition in plagioclase deformation mechanism with decreasing recrystallized grain size. Subgrain rotation recrystallization of plagioclase during dislocation creep led to grain size reduction and the creation of a recrystallized matrix, where the grain size became small enough to promote a transition from grain size insensitive to grain size sensitive deformation, leading to the onset of GBS. Subgrain rotation recrystallization has been shown to promote grain size reduction leading to a transition in deformation mechanism (e.g., Bestmann and Prior, 2003; Raimbourg et al., 2008), which is significant because the transition from grain size insensitive dislocation creep to grain size sensitive GBS has been hypothesized as a mechanism to lower the strength of a shear zone (e.g., De Bresser et al., 2001; Raimbourg et al., 2008; Platt and Behr, 2011; Czaplinska et al., 2015). In

the case of our sample, the transition to grain size sensitive GBS may have contributed to dramatic weakening of the leucosome relative to the refractory parts of the migmatite.

Strain localization in the leucosome is achieved in part by the maintenance of a small grain size and a reduction in viscosity associated with GBS. The presence of interstitial minerals between fine-grained plagioclase grains (Figure 5a and 7a) suggests that grain size sensitive deformation may be sustained by phase mixing, which is cited as an important mechanism of impeding grain growth in rocks deforming by GBS (e.g., Herwegh et al., 2011; Kilian et al., 2011). Elsewhere in Fiordland, metamorphic reaction in granulite facies orthogneiss has produced mineral mixing at small grain sizes that deform by grain size sensitive deformation (Smith et al., 2015). Regardless of the mechanism of achieving small grain sizes, strain localization is achieved by a reduction in viscosity due to GBS being a deformation mechanism associated with Newtonian viscous behavior (i.e., the stress exponent $n=1$) when accommodated by diffusion of vacancies (e.g., Karato et al., 1986; Gomez Barreiro et al., 2007; Précigout and Hirth, 2014; Czaplinka et al., 2015). The microstructures of the plagioclase in the leucosome are consistent with GBS accommodated by diffusion creep; the small grain size of the plagioclase ($d=13.9\text{ }\mu\text{m}$) with its equant grain shapes (e.g., Svahnberg and Piazzolo, 2010), the random orientation of misorientation axes within the kinematic reference frame (e.g., Bestmann and Prior, 2003), and the shift towards higher misorientation angles between neighbor-pair grains (e.g., Svahnberg and Piazzolo, 2010) all suggest that strain localized in the leucosome due to diffusion-accommodated GBS, which may have lowered the viscosity of the shear zone such that the rheology became controlled by plagioclase deforming by GBS.

5.4 Implications for lower crustal rheology

The microstructures imply three main changes in the rheology of the granulite-facies migmatite sample through time. Both omphacite and plagioclase initially deformed by dislocation creep, indicating that a two-phase flow law involving dislocation creep of plagioclase and pyroxene (e.g. Dimanov et al., 2003; Dimanov and Dresen, 2005) may best characterize the rheology of the granulite sample at the onset of dynamic recrystallization in plagioclase, when the strength of the two minerals is more comparable. However, as dynamic recrystallization progressed, plagioclase became weaker relative to omphacite, initiating strain partitioning into the leucosome, and a transition to a rheology controlled by plagioclase alone. As dehydration melting in the lower crust concentrates water in the melt (Vernon and Clarke, 2008), the leucosome plagioclase that crystallized from this hydrous melt subsequently deformed in the presence of water. Our results suggest that flow laws describing wet dislocation creep of plagioclase (Rybacki and Dresen, 2000) may best characterize the flow of deep migmatitic crust at granulite facies conditions once plagioclase is substantially dynamically recrystallized and becomes the rheology-controlling phase. A final transition in rheology is suggested by the onset of GBS in plagioclase, where the change in plagioclase deformation mechanism was likely accompanied by a weakening of the zone deforming by GBS (e.g., Raimbourg et al., 2008; Platt and Behr, 2011), resulting in effective strain localization in the leucosome. Flow laws describing diffusion creep of plagioclase (e.g., Rybacki and Dresen, 2000) may best characterize the flow of deep crust once GBS has been activated in plagioclase. The changing rheologies implied by the evolving microstructure show that macroscopic flow of leucosome-bearing deep crust is likely to be strongly influenced by the degree of grain size reduction associated with crystal-plastic deformation, and the mineral phase(s) in which it is developed.

Assumptions about lower crust rheology have important implications for the collapse of the Cordilleran-type orogen that resulted in the rifting of Zealandia away from Gondwana. Numerical models assume power law creep of diabase for the lower crust, and the presence of up to 10% partial melt to show how deformation of the lower crust contributes to continental breakup (Rey and Muller, 2008; Rey and Muller, 2010). The results of our work suggest that the strength of the lower crust would be best modeled with flow laws for monomineralic plagioclase given the leucosome proportion and the tendency for strain to localize within leucosome. Large melt fractions in the lower crust and mantle contribute buoyancy forces that promote runaway extension in the upper crust when subduction velocity is lowered (Rey and Muller, 2010), and our work shows that this may be primarily accommodated in plagioclase-rich leucosome. The transition from dislocation creep to GBS in the plagioclase of the leucosome provides a physical explanation of how strain can quickly localize within shear zones that coalesce and accommodate the extension leading to rifting, perhaps lending insight into how deformation within the Resolution Island Shear Zone led to the eventual break-up of Gondwana.

6. Conclusions

Our microstructural and EBSD analyses of the leucosome and restite of a high-strain, granulite facies migmatite reveal details of how melt-rich rocks accommodate deformation, and they provide insight into lower crustal rheology. Plagioclase and omphacite are the rheology-controlling minerals of the leucosome and restite, respectively, but plagioclase is the weaker of the two and therefore exerts primary control on the strength of melt-rich lower crust. Plagioclase in the leucosome deforms by subgrain rotation recrystallization during dislocation creep,

resulting in grain size reduction that promotes a transition to grain size sensitive GBS. A small grain size was maintained by phase mixing, which sustained diffusion creep-accommodated GBS and promoted strain localization within the fine-grained leucosome. We interpret three main changes in rheology during the microstructural evolution of the sheared migmatite that can be modeled by dislocation creep of anorthite-diopside aggregates at low strain, followed by dislocation creep of wet plagioclase aggregates and then diffusion creep of wet plagioclase aggregates as strain became localized within the leucosome.

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Figure Captions

Figure 1. a) Regional map of Gondwana during the Early Cretaceous, showing the location of Fiordland relative to the orogenic belt and magmatic arc produced by subduction beneath Gondwana (after Mortimer, 2008). Thick black lines indicate the location of Late Cretaceous (<85 Ma) rifts. A-A' denotes schematic cross-section in Figure 1b. PNG - Papua New Guinea, HP - Hikurangi Plateau. b) Schematic cross section of the convergent margin showing location of the Western Fiordland Orthogneiss (WFO), which represents the lower crust of the Early Cretaceous magmatic arc.

Figure 2. Geologic map and cross-section showing location of study area and sample locality within a subdome of a larger gneiss dome. The study area is on Resolution Island, south of Breaksea Entrance and near the southern extent of the WFO in central Fiordland.

Figure 3. Photographs of field relationships in the Breaksea and Resolution orthogneisses at and near sample site 0656 (see Figure 2 for location). a) Migmatitic granulite layer in a high-strain zone of the Breaksea Orthogneiss showing deformed clusters of garnet (Grt) and omphacite (Omph) rimmed by diopside-plagioclase symplectites within a foliated (S_2) plagioclase-rich matrix. b) Close-up view of migmatitic omphacite granulite gneiss sampled for detailed study and EBSD analysis. Dashed line shows orientation of granulite facies foliation (S_2). Photograph shows 1 to 2 cm-long strain shadows of plagioclase (Pl)-rich leucosome surrounding clusters of

coarse garnet and omphacite in a high-strain zone. c) Block of garnet pyroxenite gneiss wrapped by external S_2 folia in a high-strain zone of Breaksea Orthogneiss that was once partially molten. Note the pockets of weakly deformed migmatite with strain shadows of the garnet pyroxenite block. Arrow labeled 'gradient' shows the transition from low to high strain that was accompanied by recrystallization. d) Tightly folded and stretched layers (S_1) of hornblende-rich gabbro in a high-strain zone of the Resolution Orthogneiss. e) Disaggregating layer of garnet pyroxenite in a low-strain, migmatitic part of the Breaksea Orthogneiss. Fractures filled with leucosome derived from the matrix (arrows) indicate that the mafic layer broke apart within a partially molten host.

Figure 4. Photomicrographs of granulite sample under cross polarized light. a) Isotropic garnet (grt) grains have irregular and embayed shapes. Garnet is surrounded by recrystallized plagioclase (pl). b) Omphacite porphyroclasts (omph) have rounded grain boundaries, and are surrounded by fine-grained material. Biotite (bt) is present within the fine-grained material, and garnet grains have inclusions. c) A rare plagioclase porphyroclast has subgrain development along its rims. Omphacite porphyroclasts are surrounded by fine-grained minerals that are difficult to identify by optical methods. This area is featured in Figure 5c. d) Plagioclase forms a fine-grained recrystallized matrix between coarse garnet and omphacite grains.

Figure 5. EBSD phase maps. a) EBSD phase map layered on a band contrast map acquired with a 17.1 μm step size. The map shows the distribution of the four most abundant phases: garnet, omphacite, plagioclase, and K-feldspar. The locations of inset EBSD maps are designated by labeled white boxes. b) EBSD phase map of an inclusion in a garnet grain. The step size of the

map is 1 μm . The inclusion contains plagioclase, K-feldspar, and quartz. c) High resolution (2.38 μm step size) inset EBSD phase map of the plagioclase porphyroclast featured in Figure 4c and the fine-grained material along omphacite rim. The fine-grained material is a symplectite texture between plagioclase, omphacite, and hornblende.

Figure 6. Pole figures of crystallographic orientations from the EBSD map shown in Figure 5a. Data are displayed in the kinematic reference frame in lower hemisphere, equal area projections. The contoured pole figures were constructed with a halfwidth of 10° . N designates the number of grains used to construct the pole figures, and the maximum and minimum values of multiples of uniform distribution (m.u.d.) are noted. Gray scale legends showing m.u.d. are indicated at the right side of each pole figure. The orientation of the foliation and lineation are shown relative to strain axes X and Z in the schematic pole figure at bottom left. a) Scatter plot pole figures for garnet. b) Contoured pole figures for omphacite. c) Contoured pole figures for all plagioclase grains. d) Contoured pole figures for the coarse-grained plagioclase. e) Contoured pole figures for the fine-grained plagioclase. f) Contoured pole figures for K-feldspar.

Figure 7. High-resolution inset EBSD map and accompanying pole figures for plagioclase. The location of this inset map is shown in Figure 5a. The orientation of the foliation and lineation in the pole figures are shown relative to strain axes X and Z in the schematic pole figure below subfigure f. Gray scale legends showing m.u.d. are indicated at the right side of each pole figure. a) EBSD phase map layered on a band contrast map. The map was made with a 2.48 μm step size. The locations where K-feldspar grains cut plagioclase twins are shown with white arrows. b) Excerpt of phase map and accompanying EDS K $K\alpha$ map. The location of the map is shown

with a rectangular box within Figure 7a. c) EBSD phase map showing an isolated subset of the region with coarse-grained plagioclase. The average grain diameter (d) of plagioclase in this region is 20.7 μm . d) Pole figures showing the orientation of coarse-grained plagioclase. e) EBSD phase map showing an isolated subset of the region with fine-grained plagioclase. The average grain diameter (d) of plagioclase in this region is 13.9 μm . f) Pole figures showing the orientation of fine-grained plagioclase.

Figure 8. Plagioclase misorientation data from coarse- and fine-grained regions of inset map shown in Figure 7a. The number of misorientations (n) and the maximum and minimum values of multiples of uniform distribution (m.u.d.) are noted for the misorientation axes data. Gray scale legends showing m.u.d. are indicated at the right side of each misorientation axis orientation figure. Misorientation data in the crystal reference frame are displayed in equal area, upper hemisphere plots using a contour halfwidth of 15°, and misorientation data in the kinematic reference frame are displayed in equal area, lower hemisphere plots using a contour halfwidth of 15°. Misorientation angle distribution histograms show both neighbor-pair and random-pair populations relative to a theoretically random fabric. Histograms were created with a minimum angle of 3° and bins of 5°. a) Misorientation axis orientation data for coarse-grained plagioclase shown in the crystal reference frame. b) Misorientation axis orientation data for coarse-grained plagioclase shown in the kinematic reference frame, relative to the X, Y, and Z strain axes. c) Misorientation angle distribution histogram for coarse-grained plagioclase. d) Misorientation axis orientation data for fine-grained plagioclase shown in the crystal reference frame. e) Misorientation axis orientation data for fine-grained plagioclase shown in the kinematic

reference frame, relative to the X, Y, and Z strain axes. f) Misorientation angle distribution histogram for fine-grained plagioclase.

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Table 1

Table 1: Summary of EBSD Map settings

target of EBSD map	EBSD mapping technique	size of map (length in μm x height in μm)	Figure number	step size (μm)	number of LAM fields	resolution of each field (# horizontal pixels)	accelerating voltage (kV)	spot size (unitless)	emission current (μA)	working distance (mm)	band detection	Hough space
coarse-grained omphacite, garnet and plagioclase	large area map (LAM)	15,000 x 4000	Figure 5a	17.1	56	64	25	7.5	102	14.1	refined accuracy	100
plagioclase porphyroclast and omphacite rim	large area map (LAM)	1200 x 1200	Figure 5b	2.38	4	256	25	7.5	94	14.8	refined accuracy	100
fine-grained plagioclase and potassium feldspar in shear bands	large area map (LAM)	1200 x 1600	Figure 8a	2.48	9	256	25	7.0	93	13.9	refined accuracy	100
inclusion within garnet grain	beam map	80 x 150	Figure 10	1.0	N/A	N/A	25	7.5	90	14.4	refined accuracy	100