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The deformation of plagioclase investigated using electron backscatter diffraction crystallographic preferred orientation data

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Abstract

Extensive plagioclase grain orientation datasets have been collected using the rapid and accurate electron backscatter diffraction SEM technique. The crystallographic preferred orientation (CPO) of a porphyroclastic-rich plagioclase layer and an equigranular plagioclase layer in an amphibolite facies shear zone from Harris, Scotland, were investigated in order to study their CPOs and deformation mechanisms. The layers have random CPOs. Together with other microstructures and misorientation distributions, these suggest dynamic recrystallisation followed by dominant grain boundary diffusion creep. This is surprising given the coarse (100–250 μm) recrystallised grain-size of the layers. Plagioclase within tectonites of these grain-sizes in the amphibolite facies are generally inferred to deform by dislocation creep from microstructures and observed strong CPOs. Deformation mechanism maps show that this style of deformation is possible at very slow strain rates. © 2001 Elsevier Science Ltd. All rights reserved.

Keywords: Plagioclase deformation; Grain boundary diffusion creep; Crystallographic preferred orientations; Electron backscatter diffraction

1. Introduction

Crystallographic preferred orientations (CPO) are powerful for investigating rock deformation mechanisms (Schmid and Casey, 1986; Wenk and Christie, 1991). CPOs may be produced by dislocation creep (Wenk and Christie, 1991) and the rotation of inequant grains (Shelley, 1979). Grain-size sensitive creep (GSS) may weaken pre-existing CPOs, producing random CPOs (Zhang et al., 1994) or may maintain an existing CPO (described below). Many models for GSS processes have been proposed from theory and experiments (Coble, 1963; Rutter, 1983; Wheeler, 1992; Paterson, 1995) and include grain boundary sliding, Coble creep and Nabarro Herring creep. Flow laws indicate these processes are strongly sensitive to grain-size and are enhanced at finer grain-sizes. For all deformation mechanisms the symmetry of the CPO reflects that of the process that generated it (Paterson and Weiss, 1961).

The deformation behaviour of plagioclase is of interest, especially in respect to the lower crust where plagioclase-rich rocks are common. Detailed studies on shear zones where the conditions of deformation have been constrained can give further understanding to mineral and rock deformation. In this study the CPOs and deformation mechanisms of plagioclase were investigated in two nearly monophase plagioclase layers of a gneiss in a shear zone. In addition, microstructures and grain misorientation relations gave support for suggested deformation mechanisms.

2. Geological setting

The shear zone examined here is 8 m wide and is developed in a banded meta-anorthosite body in Harris, Scotland (Fig. 1a). The shear zone occurs at the margin of the body at GR NG08 0259 8712. The meta-anorthosite is one of several meta-igneous bodies in South Harris (mainly metadiorites and metabasites) that occur within NW–SE trending belts of metasedimentary rocks. Radiometric dating (Cliff et al., 1983) gives an early Laxfordian age (2250 Ma) for the igneous assemblage. The igneous bodies were affected by regional metamorphism. A metabasite body near Rodel (Fig. 1a) contains green amphibole, orthopyroxene and clinopyroxene in textural equilibrium suggesting this metamorphism was of upper amphibolite/lower granulite facies. The meta-anorthosite comprises plagioclase with a coarse polygonal texture and mafic bands containing mainly clinopyroxene, almandine garnet and green amphibole. Retrogressive amphibolite facies shearing followed granulite facies metamorphism. Large-scale shear deforms the metasediments but impose only

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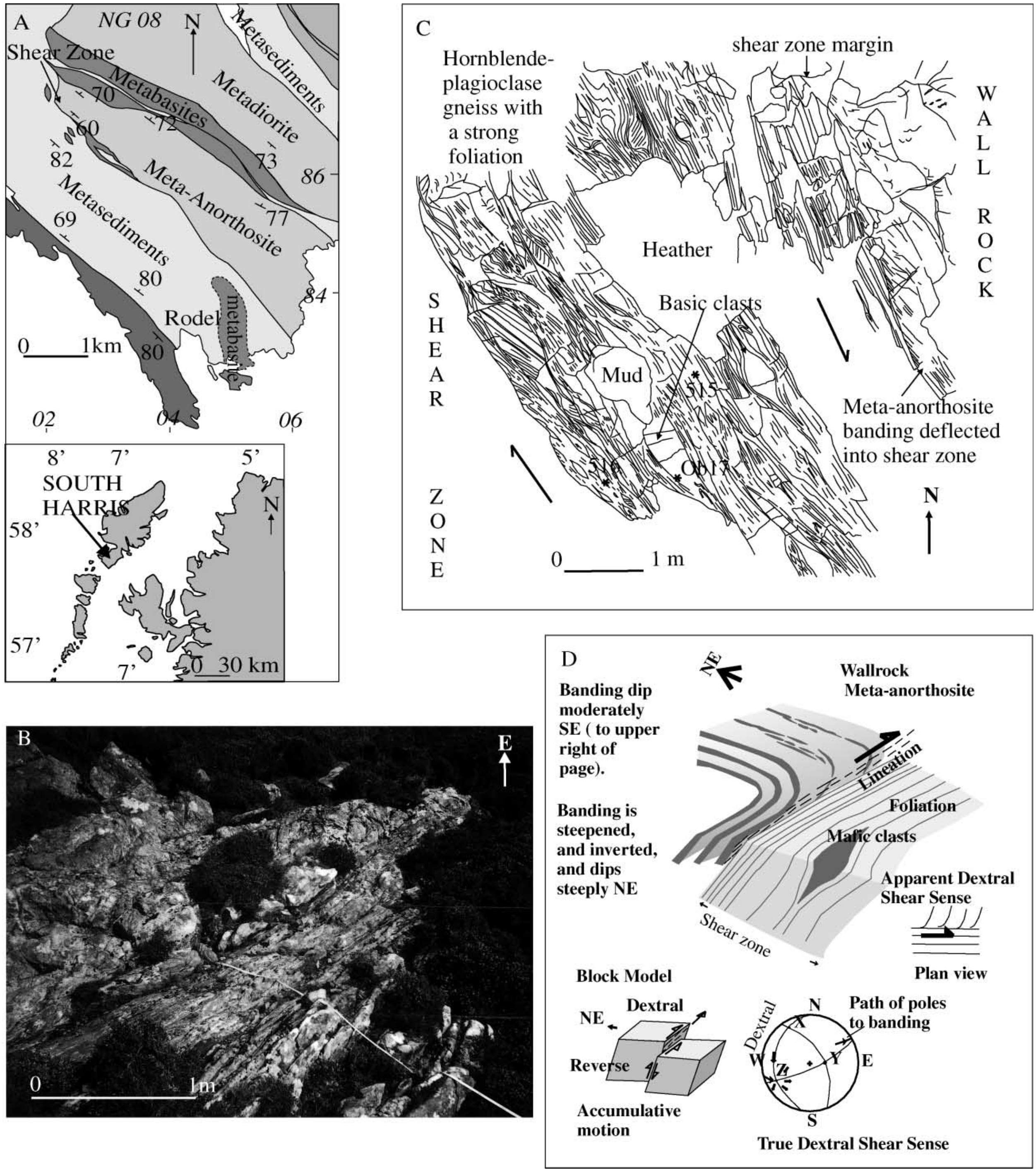


Fig. 1. (a) Geological map of SE Harris showing geographical and geological location of the shear zone, and insert map showing the location of South Harris. (b) Photograph of the shear zone. Tape measure extends across the shear zone; wallrock is to the upper left. (c) A base map (to scale) of the shear zone showing the location of samples 508, 515 and Ob17 described in the text and highlighting the general orientation of the banding, foliation and other features seen in the photograph. Note that to relate to the photograph, the photograph view is facing east. (d) A diagram showing the structural features and kinematics of the shear zone.

localised marginal strain on the meta-anorthosite. Some smaller-scale shear zones, such as the 8-m-wide one studied here, cut and deform the meta-anorthosite.

The shear zone contains a mafic gneiss comprising plagioclase- and hornblende-rich layers, which are 1 mm–1 cm wide. The shear zone dips steeply NE, and contains a strong foliation parallel to the shear zone margins (Fig. 1b and d) defined by gneissose layering and a strong shape fabric in the hornblende-rich layers. The lineation plunges gently NW. The apparent rotation of the wallrock banding into the shear zone (Wheeler, 1987) indicates a dextral reverse shear sense (Fig. 1d). This is also suggested from the asymmetry of mafic clasts. Clinopyroxene and garnet in the wallrock are rare in the shear zone. The composition of the gneiss, containing hornblende, plagioclase (andesine, An_{34-44}) and minor quartz, biotite and opaque, suggests amphibolite facies deformation. This paper concentrates on plagioclase deformation and the deformation of other minerals is not described in detail. The plagioclase layers are nearly monophase but contain minor quartz. Some of these layers are porphyroclastic and others are equigranular. The two layers studied in detail are a porphyroclastic layer within sample 515 and an equigranular layer within sample 516. The locations of these samples and of a comparison sample of parent material (sample Ob17) are shown in Fig. 1c.

3. Methods

3.1. CPO measurement

Grain orientation data were collected using the electron backscatter diffraction (EBSD) method (see Prior et al. (1999) for details of this method). This is an easy method of measuring plagioclase grain orientations and large datasets can be collected quickly (e.g. in this study 450 grains were measured in 2 days). The data were collected using a live atomic number contrast image as grain boundaries were visible due to iron-rich impurities and voids and the beam could be directed at different grains in turn. Diffraction patterns collected from each grain were indexed using the programme Channel + (Schmidt and Olesen, 1989). A reflector file for albite (Prior and Wheeler, 1999) was used to index the diffraction patterns of the intermediate composition (labradorite and andesine) plagioclase in this study. Only indexing solutions that gave a good fit of the simulated EBSD patterns were accepted. This was often possible. Indexing of this plagioclase was thought accurate as practical experiments (Prior, personal communication) show that plagioclase diffraction patterns do not change from albite up to at least An_{65}. In addition, measuring orientations in single grains either side of twin boundaries gave the expected twin orientation relationships.

The data were collected using an accelerating voltage of 20 kV, a beam current of ~3 nA with a working distance of ~24 mm. The specimens were chemically polished and left uncoated, with a layer of carbon paint around the specimen margins.

In triclinic crystals all crystal directions may be distinguished, i.e. it is possible to distinguish between [uvw] and $[\bar{u}\bar{v}\bar{w}]$. The CPO data are therefore displayed as upper and lower hemisphere pole figures.

3.2. Misorientation distribution data

Misorientation distribution data can give more information about grain orientation relations than the CPO, giving further insight into deformation processes. The concepts of misorientation and misorientation distributions has been established by experimental and numerical work (including that of Grimmer (1980) and Moraweic (1997)). Application of misorientation data to the interpretation of grainscale processes in rocks has only recently developed and is proving invaluable in the interpretation of deformation processes (Lloyd et al., 1997; Fliervoet et al., 1999; Wheeler et al., 2001). A brief description of this relatively new type of data is included here.

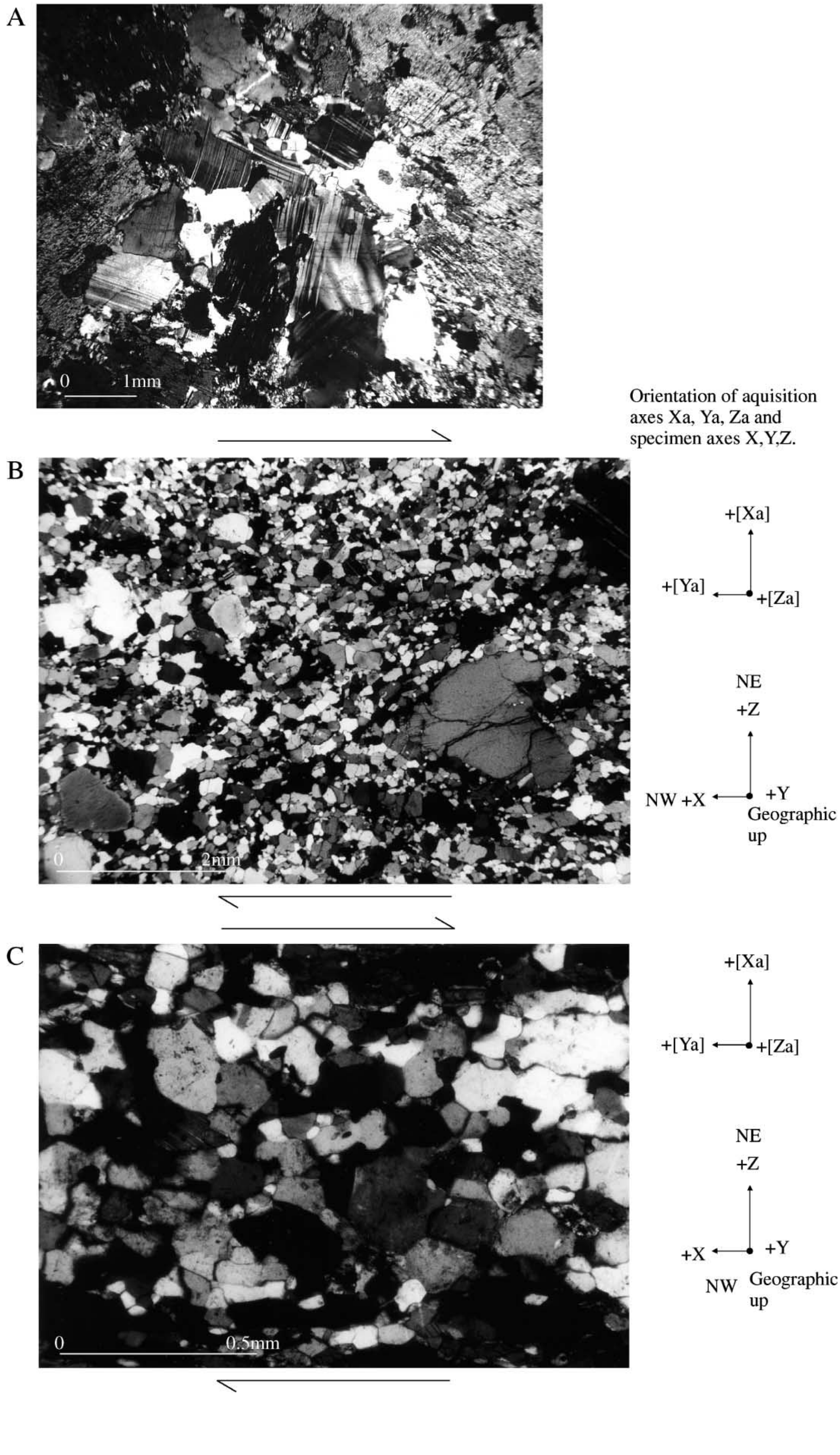
The misorientation between two grains is the difference in lattice orientation between them. The misorientation is described by a rotation angle about an axis of rotation, termed misorientation angle and misorientation axis. For minerals with high crystal symmetry, many different misorientations can describe the misorientation between two grains. The misorientation with the smallest misorientation angle is used. For minerals with lower symmetry there are fewer possible misorientation solutions between two grains. Plagioclase has the lowest crystal symmetry and for this symmetry there is only one misorientation solution for any grain pair. The misorientation distribution is the range of different misorientations between grain pairs in a sample. Two types of misorientation distribution can be considered (Wheeler et al., 2001). A random-pair distribution is the distribution produced by choosing grain pairs at random, i.e. pairs of grains that are not necessarily neighbouring grains. Ideally, all grain pairs would be selected, but as this would generate a very large dataset a smaller but statistically representative dataset is collected. A neighbour-pair misorientation distribution is the range of misorientations between neighbouring grains and the size of the dataset depends on the number of adjacent grains measured.

The angles and axes of misorientation can be plotted separately on 2D diagrams, making distributions easy to visualise. Angles of misorientation are plotted on histograms to show the frequencies of different misorientation angles. Categories of: very low 0–10°, low 10–40°, intermediate 40–90°, and high 90–180° misorientation angles are chosen in this study for description of data sets. Misorientation axes are related to a crystallographic reference frame and are plotted on inverse pole figures. The misorientation axes orientations associated with different angles of misorientation can be shown on different inverse pole figures.

4. Description: microstructures, CPOs and misorientation distribution data

4.1. Wallrock

The meta-anorthosite wallrock varies from unbanded to finely banded, containing clusters of thin (~1–4 mm), widely spaced (10 cm–1 m) mafic strands and bands. The


A
0 1mm
Orientation of aquisition axes Xa, Ya, Za and specimen axes X,Y,Z.
B
0 2mm
+[Xa]
+[Ya]
+[Za]
NE
+Z
NW +X
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C
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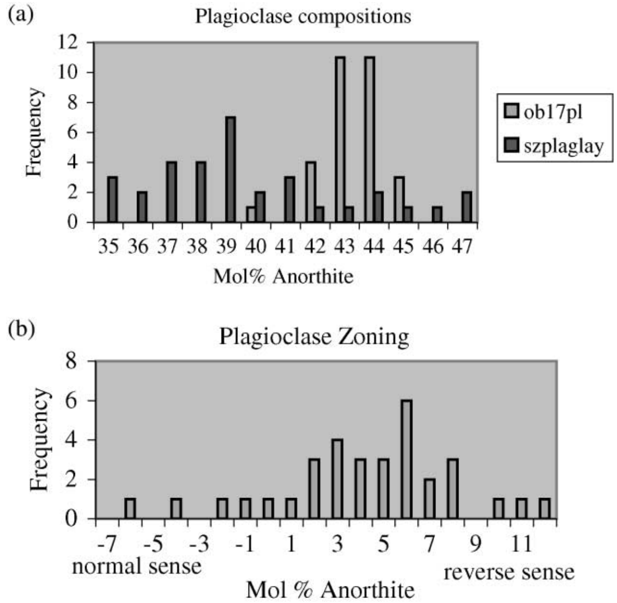

Fig. 3. Histograms showing the composition of the plagioclase in the mafic clast compared with that in the equigranular layers of the shear zone (a), and (b) the zoning of plagiclase in equigranular layers of the shear zone. The values plotted in (b) are core compositions subtracted from rim compositions.

shear zone gneiss is more mafic and was probably derived from more thickly banded meta-anorthosite, which is observed elsewhere in the body, containing 0.5-m-wide closely spaced bands. Large (10 cm–4 m long) mafic clasts occur in the shear zone, which are wrapped by the gneiss. These are oblong or rectangular and have tails of drawn out clast material. The clasts contain a coarse grained mafic aggregate, which encloses aggregates of coarse plagioclase. These clasts are thought to be parts of thick meta-anorthosite bands, which were deformed, forming the gneiss. Sample Ob17 was taken from the margin of one of these clasts. The plagioclase aggregates in the clast margin comprise coarse anhedral grains with undulose extinction and deformation bands (Fig. 2a). Subgrain boundaries are well developed. Patches of finer grains and subgrains with similar extinction positions are found. Tapering twins are developed in parts of the coarse grains. The coarse plagioclase is andesine, An_{40-45} in composition (Fig. 3a).

4.2. Shear zone: porphyroclastic layer

The porphyroclastic plagioclase-rich layer (sample 515) of the mafic gneiss contains coarse (1–3 mm) irregular shaped plagioclase porphyroclasts in a finer plagioclase–quartz matrix (Fig. 2b). The porphyroclasts have strong undulose extinction. Internal variation in lattice orientation is indicated in orientation contrast images and by clustering on a stereonet of orientation measurements from within single porphyroclasts (not shown here). The porphyroclasts have serrated grain boundaries. Parts of some porphryoclasts comprise small patches of finer grains/subgrains. The matrix comprises equant or slightly inequant, 100–250 μms sub-rectangular plagioclase grains. Many of the matrix plagioclase grains show optically visible zoning. The zones form single continuous rims around a core. The rims vary in thickness around the grain cores; the rims are sub-rectangular and more inequant than their rectangular cores. Some of the inequant matrix grains show a shape preferred orientation (SPO) parallel to the lineation, whilst others are oriented at a high angle to the lineation. Less frequently, double or single crescent shaped zones are found. The cores of grains are not always exactly in the centre of grains. Plagioclase grain boundaries are mostly straight or slightly curved. The grains have slight undulose extinction and tapering twins are found in 5–10% of grains.

The crystallographic preferred orientations of all 10 porphyroclasts in the specimen were measured (data not presented) and it was found that these porphyroclasts are randomly oriented, however, as this dataset is so small this is not a reliable deduction. The data presented in Fig. 4a and b are the orientations of the matrix grains in sample 515. The matrix grains are randomly oriented. Stereonets for grains at different distances from a porphyroclast (not shown here) show their orientation is no less random closer to the porphyroclast.

Fig. 5a shows neighbour- and random-pair misorientation distributions for matrix grains near to a porphyroclast. These histograms are very similar and show high frequencies of high angles of misorientation between grains. The frequency of angles increases with misorientation angle. The small peaks and irregularities within these trends are not thought to be statistically significant. The misorientation axes for both distributions are randomly oriented. Fig. 5b shows the misorientation distributions for neighbouring grains away from porphyroclasts. These distributions are very similar to that measured close to the porphyroclast.

Lens shaped domains of quartz, slightly coarser than the plagioclase (250–500 μm) are found in this layer, which often wrap around the porhyroclasts. The quartz is granoblastic with strongly undulose extinction and deformation bands. Close to these lenses, quartz of similar grain-size is found amongst the plagioclase. Elsewhere, grains of finer (50–80 μm) quartz occur along plagioclase grain boundaries.

Fig. 2. (a) An optical image (XPL) of a plagioclase aggregate within the mafic clast margin (sample Ob17). The aggregate comprises coarse anhedral grains with undulose extinction, subgrains and patches of finer grains. (b) Optical image (XPL) of the porphyroclastic plagioclase layer (sample 515) comprising coarser porphyroclasts in a plagioclast-rich matrix. (c) Optical image (XPL) of the equigranular plagioclase layer comprising polygonal plagioclase with minor quartz (sample 516). Note plagioclase zoning in the shear zone samples. See text for further microstructural description. Large arrows indicate the shear sense. The orientations of acquisition and specimen reference frame axes are indicated.

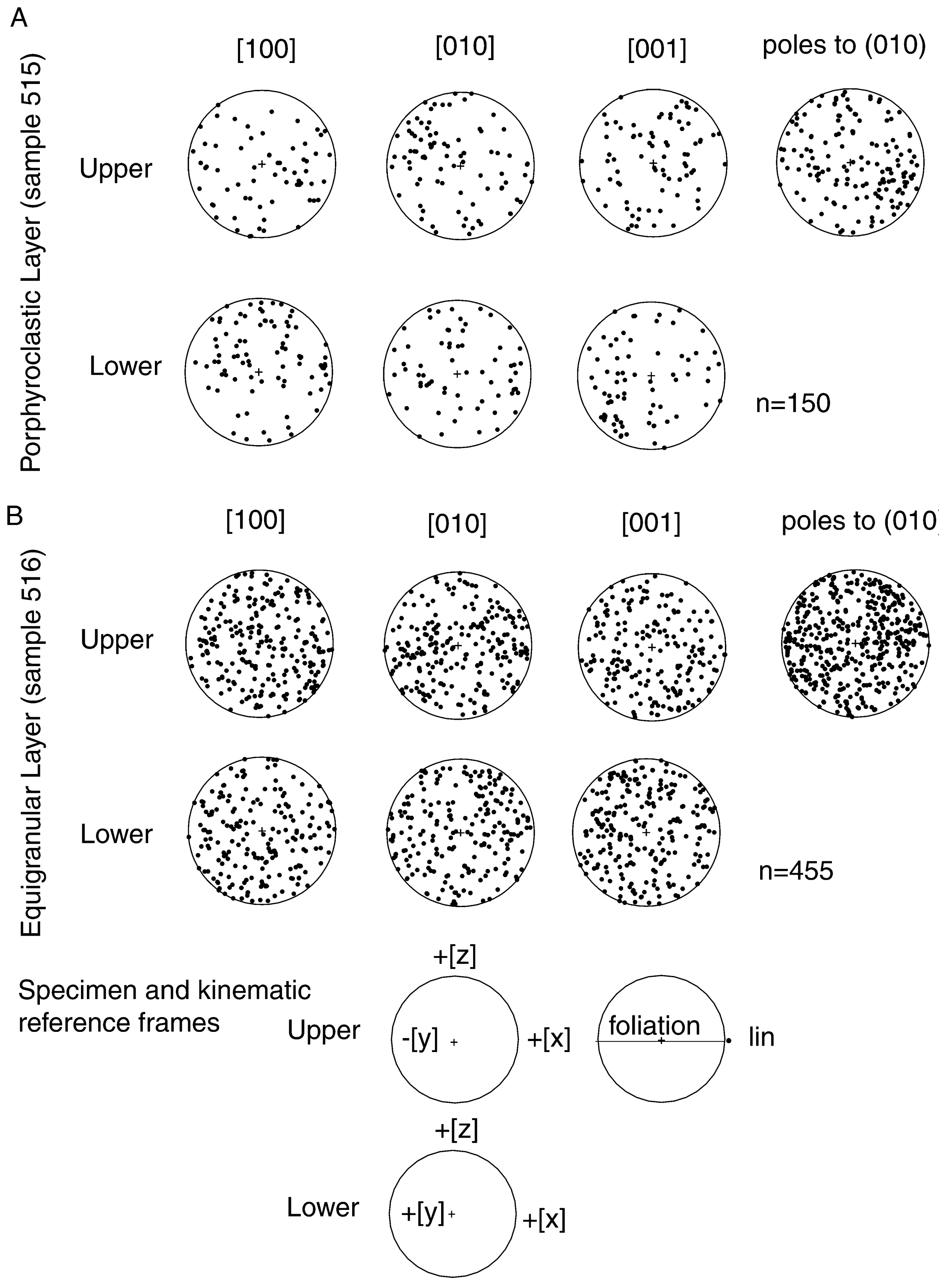

Fig. 4. Stereonets showing the crystallographic orientation of plagiclase grains in (a) the matrix grains of the porphyroclastic layer (sample 515), and (b) the equigranular layer (sample 516). Specimen and kinematic reference frames are shown.

4.3. Equigranular layer

The equigranular layer comprises coarse (100–250 μm), equant or slightly inequant sub-polygonal plagioclase (Fig. 2c). Some inequant grains show a rough SPO parallel to the lineation, whilst others occur at high angles to the lineation. The grains have slight undulose extinction and occasional tapering twinning. The plagioclase measured in

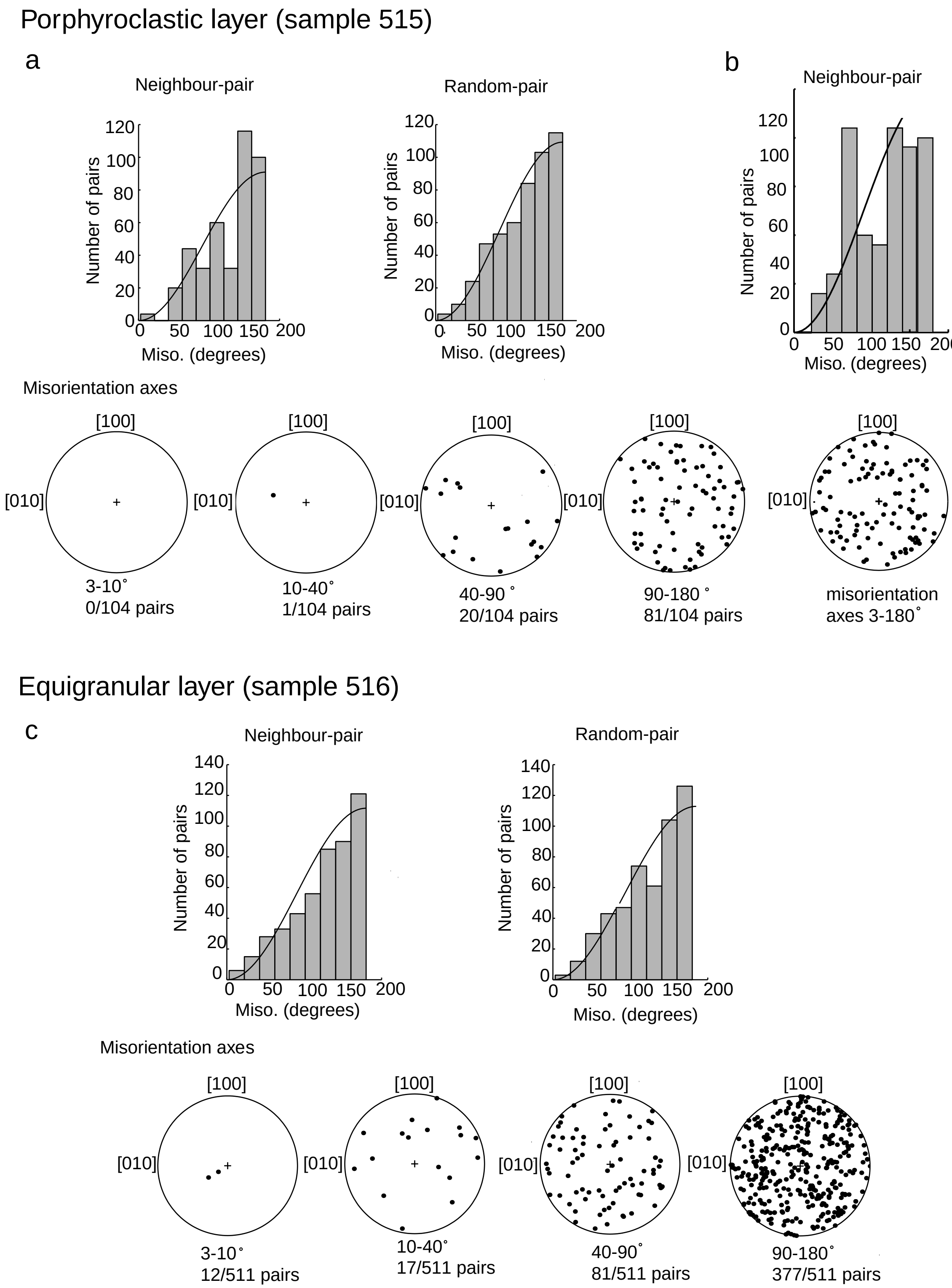

Fig. 5. Misorientation distributions for: (a, b) the matrix plagioclase of the porphyroclastic layer (sample 515), and (c) the plagioclase of the equigranular layer (sample 516) in the shear zone. For sample 515, (a) is drawn for grains close to porphyroclasts, and (b) is for grains further away from porphyroclasts. Angles of misorientation are displayed on histograms. Both random- and neighbour-pair distributions are shown. Graph lines on the histograms indicate the profile generated from a random fabric (see text). The orientations of axes of misorientation, for different misorientation angles: classes 3–10°, 10–40°, 40–90° and 90–180°, are shown on inverse pole figures. The number of pairs within each class, out of the total number of pairs taken is indicated, e.g. 28/172. Misorientation axes shown are for neighbour-pair distributions.

several equigranular plagioclase-rich layers in the gneiss was found to be An_{35-47} in composition (Fig. 3a). The plagioclase shows optically visible zoning with similar patterns to that in the matrix of the porphyroclastic layer. The zoning is mostly of reverse sense of An2–12% (Fig. 3b). Measurement of the CPO indicates that the grains are randomly oriented (Fig. 4b). Both neighbour- and random-pair misorientation histograms have profiles with high frequencies of high angles of misorientation (Fig. 5c). Misorientation axes are randomly oriented. Finer (30–50 μms) quartz grains occur along plagioclase grain boundaries and at grain boundary junctions.

5. Interpretation of GSS processes

5.1. Grain-size sensitive creep and semantics

Different authors appear to use the terms grain boundary sliding and grain boundary diffusion creep differently, so we present first our definitions. Grain boundary sliding, as a general idea, is the sliding of grains past each other. Grain boundary sliding only occurs on its own in a dilatant situation such as in unconsolidated sediments and in some fault gouges. Deformation of tectonites is not often dilatant and grain boundary sliding in these situations must occur with some accommodating mechanism, such as dislocation creep (Ree, 1994) or grain boundary diffusion (e.g. Paterson, 1995, plus references below). Grain boundary diffusion (or Coble) creep is deformation by the diffusion of material from sites of high normal stress to sites of lower normal stress in response to stress gradients (Coble, 1963). Diffusion may occur in a grain boundary fluid (pressure solution) and this process may also be driven by chemical free energy, resulting in precipitation of material of new composition (Rutter, 1983). Grain boundary diffusion creep does not occur on its own but must occur with grain boundary sliding for geometric reasons (Paterson, 1995; Mori et al., 1998). Therefore, we note that the term grain boundary diffusion creep has the same meaning as grain boundary sliding accommodated by diffusion creep. In this paper, we use the term 'grain boundary diffusion creep' and it is implicit that grain boundary sliding is involved as well.

5.2. Grain boundary diffusion creep and CPO

In the next sections we discuss the CPO and microstructures that may result from grain boundary diffusion creep in order to consider this process in this study. It can be suggested, from a review of observations made in other studies, that the grain boundary sliding component may be rotational or non-rotational, which may, respectively, weaken or preserve a pre-existing CPO. In many studies where grain boundary diffusion creep is inferred, weak fabrics are observed and may suggest that rotational grain boundary sliding occurred. This is supported by examples where coarser grains thought to have deformed by dislocation creep display a strong CPO, and finer recrystallised grains thought to have deformed by grain boundary diffusion creep display a weaker version of this CPO (e.g. as observed by Ross et al. (1996)). In other studies a pre-existing CPO is unchanged and grain boundary sliding may have been non-rotational. For example Tullis and Yund (1991) suggest diffusion creep in experimentally deformed plagioclase but find a strong CPO. They suggest that this CPO was inherited from the starting material, which had a strong CPO.

5.3. Grain boundary diffusion creep and microstructures

The redistribution of material by grain boundary diffusion creep may be obvious from crescent shaped zones of material of different composition or of fibrous habit. Truncational boundaries may occur along the margins of grains, which are at low angles to the normal stress where material has been removed. Such microstructures are observed commonly in low grade tectonites and compacting sediments (Beach, 1982; Cox and Etheridge, 1989; LaFrance and Vernon, 1993). It was identification of such microstructures that led to suggestion of this process. However, in many cases such zones are not found in tectonites thought to have deformed by this process (e.g. Fliervoet et al., 1997) and in experiments where diffusion creep is indicated from derived flow laws (e.g. Dell'Angelo and Olgaard, 1995; Goldsby and Kohlstedt, 1997). Often zones do not form as the material precipitated is of the same composition, and redistribution of material is only indicated by grain flattening. In many studies, however, equant grains are observed. Dell'Angelo and Olgaard (1995) indicate that this results from grain boundary sliding, implying rotational grain boundary sliding, which would cause the edges of grains at high and low angles to the normal stress to vary. CPO weakening is inferred in some of these studies (e.g. Ross et al., 1996) supporting rotational grain boundary sliding. We suggest that where grain elongation occurs, forming a SPO (e.g. in the experiments of Tullis and Yund (1991)) accompanying grain boundary sliding may have been non-rotational. As grain boundary diffusion creep does not involve the internal strain of grains dislocation densities should remain low and grains should appear internally undeformed. Therefore, where grain elongation does not occur, this process produces a microstructure similar to that of an undeformed aggregate (for example as observed by Dell'Angelo and Olgaard (1995), Ross et al. (1996) and Dimanov et al. (1998)). The dissolution and precipitation process may result in straight, slightly curved and truncational grain boundaries.

6. Interpretation of the plagioclase aggregates

6.1. Mafic clast margin

The coarse grain-size of the mafic clasts and also of the wallrock may reflect grain growth during granulite metamorphism. The microstructures of the plagioclase within the mafic clast margin, including the deformation bands and subgrains and patches of finer grains with similar extinction positions suggest later slight strain of the meta-anorthosite by dislocation creep and incipient dynamic recrystallisation.

6.2. Initial deformation and recrystallisation mechanisms in the shear zone

The undulose extinction and the orientation variation within the plagioclase porphyroclasts in the porphyroclastic layer suggest they deformed by dislocation glide. The microstructures suggest that this strain energy was released by dynamic recrystallisation. Subgrain rotation is suggested from patches of subgrains and the serrated and serrated grain boundaries suggest grain boundary migration. The irregular shape of the porphyroclasts may reflect varying rates of dynamic recrystallisation at their margins and suggest this process was quite important. The plagioclase-rich layers are thought to have formed by the dynamic recrystallisation and straining out of the coarse plagioclase aggregates in the mafic casts. This is suggested by the textures of incipient dynamic recrystallisation of the clast plagioclase. The recrystallisation textures of the porphyroclasts in the plagioclase layers suggest they are parts of coarse grains, which have not yet fully recrystallised. The equigranular layer does not contain porphyroclasts, at least in the part of the layer examined. The plagioclase of this layer is of similar grain-size to that in the porphyroclastic layer and it is thought equigranular plagioclase layers result from complete recrystallisation of the coarse parent plagioclase by the same mechanisms. The matrix plagioclase of the porphyroclastic layer may be of the same composition as that in the equigranular layers. The optical similarity of the zoning patterns in these layers suggests the chemistry of the zoning, measured in the equigranular layer, is similar in both layers. The recrystallised plagioclase in these layers shows a similar compositional range to the coarse plagioclase in the mafic clast (Fig. 3a). Although these ranges show peaks at slightly different compositions, this is not thought to be statistically valid given the dataset sizes. This suggests dynamic recrystallisation occurred with no chemical change.

6.3. Deformation mechanisms in the matrix

The CPOs, misorientation distributions and microstructures of the matrix of the porphyroclastic layer and equigranular layer are similar, suggesting they deformed in the same manner.

6.3.1. CPOs

The random CPOs of the recrystallised plagioclase grains suggests the dislocation creep inferred in the porphyroclasts did not continue in the matrix. Dislocation creep is expected to produce a strong CPO even if the starting material had a random CPO (Lister et al., 1978), and therefore we consider alternatives, which weaken CPOs, such as grain boundary diffusion creep or grain boundary sliding accommodated by dislocation creep. Nabarro Herring creep is less likely as volume diffusion rates are thought to be slow at amphibolite facies temperatures in plagioclase (Yund and Tullis, 1991). If this is the case then recrystallisation occurred with a switch from dislocation creep to GSS creep. Such a switch has been inferred by Karato et al. (1986), Rutter and Brodie (1988) and many others. It is also possible that random CPOs could result from multiple or complex deformation. Such multiple deformation could involve two generations of dislocation creep with the operation of different slip systems due to a change of kinematics, temperature or strain rate etc, the second of which destroyed the CPO of the first. Only modest strains are required to impose a CPO by dislocation creep (Lister et al., 1978). However, it would be an unlikely coincidence for two deformation phases to sum to give a random CPO. Complex deformation may involve several dominant deformation mechanisms. It is not possible to exclude these possibilities, however, the above interpretations are simpler and further evidence for these is obtained from misorientation distributions and petrographic microstructures.

6.3.2. Interpretation of misorientation distributions

The nature of misorientation distributions depends strongly on the crystal symmetry and the CPO (Wheeler et al., 2001). Comparison of neighbour-pair and random-pair misorientation distributions gives particular insight, especially where these profiles are different (termed correlated), as any differences may be due to deformation processes and give further information than the CPO (Wheeler et al., 2001). The signatures of some processes on misorientation distributions for all symmetries are briefly considered.

A random CPO for a given crystal symmetry will always produce the same random-pair distribution. It is important to know what this distribution will be in order to understand misorientation distributions. The random-pair distribution of misorientation angles plotted on a histogram is not uniform but shows increasing frequency with misorientation angle for geometric reasons (Wheeler et al., 2001). The random-pair distributions expected from random CPOs for different symmetries have been calculated (Grimmer, 1980; Moraweic, 1997). For triclinic crystals the histogram will show increasing frequency of misorientation angle up to 180° (curves on histograms in Fig. 5) as owing to the lack of symmetry, large misorientations cannot be described by smaller ones. The misorientation axes from a random CPO are randomly oriented. Processes that cause spatial weakening of CPOs such as grain boundary diffusion creep and have produced a random CPO, will generate misorientation distributions that are random. Where a pre-existing CPO is not fully destroyed, these processes may be identified by randomly oriented misorientation axes, as shown by Jiang et al. (2000). Grain boundary diffusion creep with non-rotational grain movement will preserve original CPOs but the neighbour-pair misorientation distribution will be changed.

Dislocation creep and processes forming a strong CPO may be expected to result in a greater proportion of lower angle misorientations on both random- and neighbour-pair distributions compared with that of the random-pair distribution for a random CPO. Subgrain rotation involves progressive lattice offset to form grain boundaries and therefore may result in very low angles (0–10°) of misorientation between grains, which may be reflected by a large peak of very low angle misorientations on the neighbour-pair distribution (Trimby et al., 1998).

The neighbour- and random-pair misorientation distributions of the recrystallised plagioclase show that high angles of misorientation are common between neighbouring and non-neighbouring grains, supporting a grain mixing process involving the rotation of grains. These distributions are uncorrelated. It is noted here that although correlation may indicate a process of local physical grain interaction, lack of correlation does not imply this has not occurred. The high misorientation angles between adjacent grains indicate the random matrix CPOs are not simply due to host controlled recrystallisation of grains of different original orientation, forming domains of different orientation, (Jiang et al., 2000). A grain mixing process with grain rotation is also suggested by the random orientation of the misorientation axes. These distributions are similar to that produced from a random triclinic CPO, and reflect the measured CPOs. In sample 515, low angles of misorientation are not found between neighbouring matrix grains close to porphyroclasts or further away from porphyroclasts (Fig. 5a and b). If subgrain rotation recrystallisation were the dominant process this would generate misorientation distributions that corresponded to the distance from the porphyroclast. The process of grain mixing in the matrix may have disrupted original grain orientation relations produced from subgrain rotation.

6.3.3. Petrographic microstructures

The sub-polygonal shape of some grains in the equigranular layer may suggest some grain boundary adjustment following deformation, however, this grain shape is uncommon in the porphyroclastic layer and deformation microstructures are preserved. Non-polygonal recrystallised grains in the equigranular layer are similar to those in the porphyroclastic layer. This material thought to be unaffected by later grain boundary adjustment is described as follows.

The recrystallised plagioclase grains are equant or slightly inequant with little indication of internal deformation. These microstructures suggest grain boundary sliding, consistent with the CPO and misorientation data. The other important grain mixing process, cataclastic flow, is unlikely given these microstructures. The equant shape of many grains and SPO of only some of the grains supports rotational grain boundary sliding. Grain boundary sliding requires an accommodation mechanism. The slight undulose extinction within grains may suggest some accommodation of grain boundary sliding by dislocation creep. The variation in thickness of the zones around grain cores and occasional crescent shaped zones suggest the diffusion of zone and grain material around grains. The single crescent shaped zones observed around the cores of some grains may reflect diffusion of material around only one part of a grain, perhaps as one margin of the grain was impinged on by another grain, deflecting diffusion paths or affecting the stress gradients around the grain boundary. Therefore, dominant grain boundary diffusion creep with some accommodation by dislocation creep is inferred.

The continuous nature of the zones suggests zoning formed by grain growth by the addition of material. The zoning is thought to have resulted from chemical reactions in the mafic minerals, which released ions forming plagioclase (not discussed here), which were precipitated from solution. Precipitation may have occurred around existing grains, as it was easier than nucleation. The more Ca-rich composition of the zones is thought to reflect the ions released from these reactions rather than temperature change. These fluids (not discussed here) may have enhanced grain boundary diffusion and sliding. It is thought less likely that the zoning formed by dissolution and re-precipitation of matrix material during grain boundary diffusion, as there is no compositional difference between the porphyroclasts and matrix grains suggesting the matrix is of equilibrium composition, and as the zoning increases the compositional range of the plagioclase in the shear zone.

The microstructures of the quartz lenses in the porphyroclastic layer suggest these lenses deformed by dislocation creep. Further investigation of the quartz deformation in comparison with that of the plagioclase by measurement of quartz CPOs would be interesting.

7. Discussion and conclusions

The CPOs, misorientation distributions and microstructures of these nearly monophase plagioclase layers suggest dislocation creep and dynamic recrystallisation of the wallrock plagioclase. This may have been followed by a switch to GSS deformation in the finer recrystallised grains involving grain boundary diffusion creep. The porphyroclasts may or may not have been affected and rotated by the surrounding matrix to form a random CPO. The critical grain-size for GSS creep depends strongly on the material and conditions of deformation. This is illustrated by studies of calcite, such as the calcite aggregate deformation experiments of Walker et al. (1990). Walker et al. (1990) suggest a switch from dislocation creep to grain boundary creep mechanisms at different grain-sizes at different deformation conditions. Busch and van der Plujim (1995) infer important dislocation creep and grain boundary sliding in a calcite ultramylonite of 20–30 μm, from the Bancroft shear zone, Ontario, and suggest this mylonite deformed in a transitional regime. They note that this is compatible with Walker et al. (1990), who infer such a regime at comparable conditions, at similar grain-sizes. Studies inferring grain boundary diffusion creep in plagioclase of different compositions in experiments by Tullis and Yund (1991) and Dimanov et al. (1998, 1999), observe grain-sizes of 2–16 μms. The critical grain-size at which transition to GSS creep in plagioclase occurs has not been investigated and it is not possible to use values suggested for other minerals, which may deform differently or values from other plagioclase studies where deformation occurred at different conditions. Generally the critical grain-size will be much coarser at natural compared with experimental strain rates. However, the grain-size of the plagioclase in this study may be unusually coarse (100–250 μm) for GSS creep at amphibolite facies as most studies of naturally deformed monophase plagioclase layers at amphibolite facies and above of these grain-sizes observe microstructures, suggesting dominant dislocation creep (Olsen and Kohlstedt, 1985; Olsen, 1987; Ji et al., 1988; Ji and Mainprice, 1990). A study of the general form of deformation mechanism maps for many minerals (e.g. those of Fliervoet et al. (1999) for olivine) shows that they have a field of GSS creep, which decreases in size with increasing grain-size, but shows that GSS creep is possible at coarse grain-sizes at very slow strain rates. Although deformation mechanism maps have their limitations they often agree with mechanisms suggested from the microstructures and CPOs of natural tectonites at the inferred conditions of deformation. Therefore, GSS deformation may have been activated in the plagioclase in this shear zone due to unusually low strain rates. In view of the porphyroclastic layer, this illustrates the fact noted by Heilbronner and Bruhn (1998) and Wang (1994) that the rheology of a rock of variable grain-size may not be described by a single flow law, different grain-size fractions deforming by different mechanisms. Different deformation mechanisms are suggested from microstructures in different grain-size fractions in many other tectonite studies. Tectonites of variable grain-size are common in the crust and this is important in the consideration of crustal deformation. Also, it should be noted that the flow law of the suggested dominant deformation mechanism from the microstructures may not correctly describe rheology where subsidiary or several mechanisms operate.

Grain boundary sliding deformation is also suggested by Jensen and Starkey (1985) and Kruse and Stunitz (1999) in mixed phase hornblende–plagioclase layers containing plagioclase of 100–150 μm and 24–41 μm, respectively, from shear zones developed in the Jotun nappe meta-anorthosite units, Norway. Therefore, perhaps the switch to GSS creep occurs at grain-sizes below 150 μm in plagioclase. However, these grainsizes are quite coarse, and many other studies find evidence for dislocation creep in plagioclase at these grainsizes. Bruhn et al. (1999) infer that interphase diffusion may greatly enhance diffusion rates in mixed phase aggregates. Therefore, perhaps the critical grain-size for GSS mechanisms in polymineralic aggregates is very different to that for single-phase aggregates.

Plagioclase rich rocks are common in the lower crust, which comprises intermediate-mafic anorthositic granulites (Ranalli, 1995). Although plagioclase deformation is not as well studied as that of quartz or olivine there has been an increase in the study of plagioclase deformation, both natural and experimental, in the last few years. Many workers suggest that coarse plagioclase aggregates deform by dislocation creep at amphibolite facies and above (Jensen and Starkey, 1985; Olsen and Kohlstedt, 1985; Ji and Mainprice, 1990). Most of these studies, leading to this suggestion, have been from investigation of plagioclase-rich tectonites in upper amphibolite–granulite facies shear zones developed in meta-anorthosite bodies. These authors observe strong CPOs and evidence for dislocation creep, including optical microstructures and TEM observed dislocation structures and densities, and suggest the strong CPOs were produced by this process. The CPOs measured are commonly of the same type, and are suggested to indicate (010)[001] slip. Shelley (1977, 1979, 1986, 1989a,b) questions whether these CPOs can be attributed to dislocation creep given that similar plagioclase CPOs are observed in greenschists facies rocks where dislocation creep may not operate in plagioclase. He notes that it is possible that several processes produce similar CPOs (Shelley, 1986) but that a common origin for this dominantly observed CPO should be considered. Shelley suggests that the CPOs found at high metamorphic grades may not be produced by dislocation creep but by other mechanisms, which are clearly suggested at greenschist facies, such as oriented growth. Dislocation creep can be a subsidiary process during/following a different CPO forming process, for example as Shelley (1989a,b) suggests for quartz in polymineralic schists and therefore he suggests a CPO and microstructural evidence for dislocation creep is not evidence for CPO formation by this mechanism. We agree that where plagioclase forms part of a polymineralic assemblage and/or does not form aggregates, a strong CPO may have formed by a different mechanism, at any facies, as the plagioclase may deform differently in this situation due to the different rheology of the phases and their interaction, and also that these mechanisms produce similar CPOs. However, the CPOs observed in the high metamorphic grade plagioclase aggregates cited above may have been produced by dislocation creep due to the microstructural evidence suggesting well established and extensive dislocation creep. Dislocation creep may be the most important mechanism in plagioclase at high grades in nearly monophase coarse grained aggregates, which may be quite common in the lower crust. Therefore, evidence for extensive dislocation creep in amphibolite facies plagioclase-rich rocks may indicate CPO formation by dislocation creep, and the lack of evidence for dislocation creep in coarse near-monophase plagioclase layers in this study is surprising and, as stated, perhaps due to low strain rates. There is good experimental and natural evidence for grain-size sensitive creep in feldspar at fine grain-sizes (Tullis and Yund, 1991; Molli, 1994; Fliervoet et al., 1997). This study highlights the strong control of environmental conditions on deformation mechanisms. The temperature, pressure, stress, strain rate, pore fluid pressure, as well as composition, grain-size, grain-size distribution and microstructures (Handy, 1989, 1990) must all be considered in assessing the deformation style of deformed tectonites.

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