Fluid-controlled grain boundary migration and switch in slip systems in a high strain, high temperature contact aureole, California, USA

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Abstract

Within the highly strained aureole surrounding the Eureka Valley – Joshua Flat – Beer Creek composite (EJB) pluton of eastern California, an inversion in microstructures and crystallographic preferred orientations (CPO) exists with distance from the contact. An inner aureole (<250m from the contact) consists of quartzites that are interbedded with marbles and calc-silicates. These quartzites are incompletely recrystallized. Most grain boundaries have migrated, although it is clear that grain boundary migration (GBM) is not extensive. Multiple data sets indicate that temperatures of deformation were above 650° C. CPOs are indicative of <a> slip in quartz. Within the outer aureole (250m to 1500m from the contact), quartzites are interbedded with pelitic schist and are completely recrystallized and microstructures are indicative of extensive GBM. CPOs are indicative of prism [c] slip. Oxygen isotope ratios in the inner aureole are only slightly shifted from their original values. Oxygen isotopes from the outer aureole are shifted more, which is consistent with equilibration with locally derived fluids. We suggest that recrystallization in the outer aureole was aided by pore water, water derived from fluid inclusions, and water generated by prograde reactions in the schists. The pore fluids in the inner aureole were also probably initially water-rich. However, during prograde reactions in the intervening calc-silicate rocks, and perhaps more importantly, between calcite cement and quartz in the quartzites, the pore fluid composition in the inner aureole changed to become dominated by CO₂, which acted as a non-wetting phase and decreased the fugacity of water slowing grain boundary mobility. Low water fugacity also suppressed the activity of prism [c] slip.

Therefore, we propose that dry conditions or a grain boundary fluid with a significant non-wetting component (CO₂) can result in apparent temperatures of deformation that are more than 100 °C lower than the real temperatures of deformation.

1. Introduction

It is known that the amount and type of fluid has a significant control on the creep strength of the crust which is often based on the characteristics of the deformation of quartz (Fitz Gerald et al., 2006; Menegon et al., 2011). The rheology of the crust is often viewed as strength profiles of dry and wet quartz based on experimental results (Kohlstedt et al., 1995; Mackwell et al 1998; Jackson 2002, Burov and Watts; 2006). The addition of a few tenths of weight percent water can drastically decrease the creep stress of deformed quartzites (Mainprice and Paterson, 1984; Kronenberg and Tullis, 1984; Post et al., 1996; Post and Tullis, 1998) and can also drop the transition temperature between dislocation creep regimes by approximately 100° C (Hirth and Tullis, 1992). Decrease of creep strength with depth is thought to be a result of the increase in water fugacity with increasing confining pressure (Kohlstedt et al., 1995; Post et al., 1996; Selverstone, 2005; Chernak et al., 2009; Holyoke and Kronenberg, 2013). The actual process of how water causes weakening is still unclear, although it is thought to occur when water-associated crystal defects increase in number. TEM studies indicate that vacuum dried Heavitree quartzite shows very little evidence for dislocation climb and dislocation glide appears to be difficult, whereas water-enriched Heavitree quartzite shows ample evidence for climb and glide (Mainprice and Jaoul, 2009).

Most data on recrystallization mechanisms come from experiments because it is difficult to document the fluid environment that existed during natural deformation. The experiments of Tullis and Yund (1982) documented how grain boundary migration and resulting grain growth was enhanced when the grain boundaries were wet. In natural systems it is much harder to document the role fluids played because fluids can move through rocks and ultimately become trapped (or not) at various stages during deformation or exhumation. There are few studies that directly tie the fluid composition that existed during deformation to the type of deformation. Mancktelow and Pennacchioni (2004) observed different recrystallization mechanisms and drastically different grain sizes in Alpine quartzites that were deformed in water-deficient conditions versus water-rich conditions. Wawrzyniec et al. (1999) used fluid inclusions to document a correlation between CO₂-rich zones and brittle deformation along the Simplon fault in the Alps. In contrast, they observed that the water-rich zones were deformed by ductile deformation mechanisms.

Increasing the X_{CO_2} of the fluid not only decreases the $X_{\text{H}_2\text{O}}$, it may also decrease the ability of the water to penetrate the crystal. First, by decreasing the $X_{\text{H}_2\text{O}}$, the $f_{\text{H}_2\text{O}}$ is decreased, which may inhibit dislocation climb (Selverstone, 2005; Mainprice and Jaol, 2009). Secondly, CO₂ is a non-wetting fluid that will tend to isolate any intergranular water on grain boundaries and therefore decrease the effective permeability for water. Even very low amounts of CO₂ can promote non-wetting behavior (Watson and Brenan, 1987; Holness and Graham, 1995). Therefore, the creep strength of quartz apparently does not depend solely on the

amount of *intragranular* water, but also on the amount of *intergranular* water (Post et al., 1996; Post and Tullis, 1998).

In this study we document differences in the microstructures and in the slip systems between the inner and outer aureole quartzites of the Eureka Valley-Joshua Flat-Beer Creek (EJB) composite pluton and attribute these differences to the fluid composition that existed during deformation. Both parts of the aureole were intensely deformed at temperatures above 600° C. An initial study of five samples from part of this aureole (Morgan and Law, 2004) indicated that variable amounts of available water were the cause of the differences in slip-systems and amounts of recrystallization but more data was needed to validate the change between the inner and outer aureole and also to support the fluid-controlled hypothesis. Here we present the data from twenty-two samples collected across the aureole (we include the data from the 5 original samples) and use stable isotopes and metamorphic mineral assemblages to make the case that the inner aureole quartzites deformed under dry conditions, probably due to the increase of X_{CO_2} within the fluids due to in-situ prograde calc-silicate reactions. For a more detailed discussion of the metamorphism and partial melting within the aureole see Nabelek and Morgan (2012) and for a discussion on the emplacement of the EJB pluton see Morgan et al., (2013).

2. Geologic Setting

The central White-Inyo Range in eastern California (Fig. 1) is dominantly composed of a moderately deformed Neoproterozoic to upper Cambrian sedimentary package that is over 7,000 m thick and formed part of the Cordilleran

passive margin sequence (Stewart et al., 1966). Fossils are still preserved throughout the range (Nelson et al., 1991) although there are zones of intense deformation adjacent to plutons and thrust faults. A well-developed slaty cleavage is observed in the pelitic units and a spaced fracture cleavage is often found in the carbonates. The sedimentary sequence was subjected to greenschist facies metamorphism (chlorite zone) prior to pluton emplacement (Ernst, 1996). The structure of the range is dominated by an anticline-syncline pair trending NNE with wavelengths on a scale of 15-20 km (Fig. 1). Many smaller-scale associated folds can be found throughout the range. In a structural window in the topographically lowest part of the center of the range, the Neoproterozoic to upper Cambrian sequence is on top of and in thrust contact with a Devonian and Mississippian sedimentary sequence (Stewart et al., 1966; Dunne et al., 1978; Corbett et al., 1988). Thrusting was to the east and is constrained to have formed between the early Permian and earliest Triassic (Snow 1992; Stevens et al., 1997). The large-scale folding in the range is probably related to ramps in the underlying thrusts (Corbett et al., 1988). An older, NE-trending set of folds with wavelengths of 4-5 km is also associated with a slaty cleavage and the resulting interference of NE and NW folds has produced a series of structural domes and basins (Nelson et al. 1991; Morgan and Law, 1998). Morgan and Law (1998) and Morgan et al., (2013) document how both periods of deformation predate the emplacement of Jurassic and Cretaceous plutons.

3. Description of undeformed sedimentary formations

Nelson (1962) was the first to map in detail and describe the sedimentary sequence within the central White-Inyo Range. Within the EJB aureole traverse studied here, the oldest sedimentary unit is the lower Cambrian Harkless Formation, which dominates most of the study area. The following description is mostly based on the work of Nelson (1962). The Harkless Formation (~650 m thick) is dominantly composed of green to gray shale with subordinate interbeds of fine-grained sandstone and locally thick sequences of quartzite. The quartzite is not always present but when it is, it is notable for its abundant skolithos trace fossils. The top of the Harkless is defined by thin limestone beds. Above the Harkless Formation is the Saline Valley Formation (~275 m thick), which is divided into a lower and an upper half. The lower half is dominated by sandstone and shale and capped by limestone. The upper half is a mix of sandstone, shale, and limestone. Above the Saline Valley Formation is the Mule Spring Formation (~300 m thick, Nelson, 1962), which is a blue-grey, locally dolomitic, limestone. The Mule Spring Formation caps the lower Cambrian sequence. The Middle Cambrian begins with the Monola Formation (~400 m), which is defined by thinly bedded siltstone and shale. The Monola Formation abuts against and is intruded concordantly by the EJB pluton in the area studied.

This study examines an 1800 m traverse on the east side of the EJB pluton (Fig. 2) where all lithological layering and the foliation is subvertical to steeply dipping. The Harkless Formation takes up more than 1500 m of the traverse. The reason for this is that the traverse does not follow a straight line and the sedimentary units do not maintain their north-south strike away from the pluton.

The traverse is initially oriented east-west (Fig. 2) and perpendicular to the concordant formations at the pluton contact. At approximately 300 m from the pluton contact, the sedimentary formations rotate approximately 45° clockwise to strike NE. Exposure is limited at this distance from the contact and in order to continue sampling quartzite, the traverse rotates from its E-W trend to become NE trending to follow an exposed ridge of the Harkless Formation.

Within the traverse (Fig. 3), the main difference between the Saline Valley quartzites, found closer to the pluton, and the Harkless quartzites is that the Saline Valley Formation is interlayered with calc-silicate and marble layers while the Harkless Formation is interlayered with andalusite-sillimanite schists.

4. Strain and kinematic framework

A detailed analysis of the strain and kinematics of the aureole associated with the emplacement of the EJB pluton has been addressed in a separate paper (Morgan et al., 2013). Here we summarize the salient points; 1) the deformation is a result of the forceful emplacement of the EJB pluton and therefore the timespan of the strain and the thermal pulse is limited to the magmatic life of the pluton. 2) The percent shortening in the inner aureole, based on the change in the thickness of the Saline Valley Formation (based on comparison with published regional values to the thickness in the aureole) is between 43 and 47%. We were unable to measure the finite strain in the outer aureole. 3) The strain is dominated by flattening; chocolate-tablet boudinage is common throughout the entire aureole. The greatest separation between boudins is always in a vertical direction and where lineations

are observable, they are subvertical (Fig. 16 in Morgan et al., 2013). Therefore, all thin sections have been cut perpendicular to the foliation and are in a vertical plane.

5. Methods

Percent recrystallization was accomplished by taking digital images of four random areas of each sample/thin section. Individual grains were counted and grains with smooth and rounded boundaries covering a minimum of $\sim\frac{1}{4}$ of their circumference and having a shape consistent with an original sedimentary grain were recorded as not being recrystallized. Grains with straight, cusped, or irregular boundaries covering more than $\frac{3}{4}$ of their circumference were recorded as recrystallized. The distance of $\frac{1}{4}$ of the grain circumference was chosen to increase the confidence that the round nature of the boundary was sedimentary in origin and not just a coincidence of recrystallization/migration. A minimum of 300 grains were counted per sample with a maximum of 700 grains. CL images were produced at the Rensselaer Polytechnic Institute using a Cameca SX-100 electron microprobe and at Central Michigan University using a Hitachi 3400N-II SEM. All quartzite samples collected along the traverse were analyzed for their crystallographic preferred orientations (CPOs) with a universal stage attached to a petrographic microscope. All consist of almost pure quartz (>95% quartz) except for 7-50-2, which is $\sim$ 85% quartz. Given the large grain size ($\sim$ 0.5 mm), only 200-300 grains per thin section were measured optically given the constraints of the stage and size of thin section. Two marble samples were also measured for CPO. Two quartzites were also chosen for CPO using Electron Backscatter Diffraction (EBSD); one from

the inner aureole (W-71) and from the outer aureole (W-68). EBSD analyses were carried out at the University of Wisconsin-Madison using a Hitachi S3400 SEM and using the HKL (now Oxford) EBSD system. The electron beam was moved each step by the average grain size of the sample. Less than 10% of EBSD patterns could not be indexed. For analysis of oxygen isotope ratios ($^{18}\text{O}/^{16}\text{O}$), quartzite samples were prepared by crushing, acid washing, and were hand-picked to remove accessory minerals. Oxygen was liberated by laser fluorination and analyzed by a dual-inlet gas-source mass spectrometer at the University of Wisconsin-Madison (Valley et al. 1995).

6. Microstructures and petrography

Microstructural analysis of the samples collected along the traverse reveal a transition from incompletely recrystallized quartzites collected within the first 250 m from the pluton contact (Saline Valley Formation and the first 10 m of the Harkless Formation) to completely recrystallized quartzites between 250 m and 1800 m (Harkless Formation). The crystallographic preferred orientations (CPOs), the accessory minerals within the quartzites, and the mineralogy of the interbedded units in these two zones are all different. Therefore we have divided the aureole into inner and outer zones based on these differences.

We first describe the microstructures of the (relatively) undeformed Harkless quartzite collected 22 km to the SW of the EJB pluton. These regionally deformed quartzites have only been metamorphosed at lower greenschist-facies

conditions (Ernst, 1996) and were unaffected by the plutons. After this baseline description, we describe the microstructures of samples collected along the traverse.

6.1 Harkless quartzite far from pluton (starting material)

Several samples were collected 22 km southwest of the EJB pluton. These quartzites contain abundant skolithos tubes and bedding is obvious. A slaty cleavage is well developed in the pelitic units. A variably developed spaced fracture cleavage is the only macroscopic evidence of the regional deformation within the quartzites. Microscopically the deformation is consistent with greenschist-facies metamorphism. Bulging recrystallization (cf. Stipp et al., 2010) at grain boundaries along with dissolution creep between ill-fitting grains has resulted in finely sutured grain boundaries (Fig. 4) with some very small, recrystallized new grains (5-10 micrometers) and a tightly fitting network of grains with little to no pore space. Dust rings (fluid inclusions and other impurities) indicate the original elliptical shape of many grains. Subvertically orientated fluid-inclusion trails and cracks, which cross grain boundaries, are ubiquitous and are also associated with the regional deformation. Undulose extinction sweeps across most grains and extinction in deformation bands is common.

6.2 Aureole traverse samples

6.2.1 Inner aureole quartzites: Samples collected between 0 and 250 m from pluton contact

235 The first 40 m of the traverse is occupied by the Monola Formation, which is
236 a fine-grained quartz-biotite schist. The Mule Spring marble begins at 40 m from
237 the pluton-contact and occupies the next 70 m. The Mule Spring marble contains
238 fosterite and diopside and the grain size varies drastically. The Saline Valley
239 Formation begins at 110 m from the pluton-contact and ends at the contact with the
240 Harkless Formation 240 m from the pluton. The Saline Valley Formation consists of
241 interbedded marble, calc-silicate, quartzite, and minor andalusite schist. The first
242 10 m of the top of the Harkless Formation also contain marble layers and quartzites
243 from this interval have similar attributes to the Saline Valley Formation samples.
244 Therefore, the separation between the inner and outer aureole occurs at this
245 distance (250 m) and not at the contact between the two formations.

246 Microstructures of quartzites within parts of the aureole have previously
247 been described by Morgan and Law (2004); samples W-71, W-72, W-73 from the
248 inner aureole and samples W-75 and W-68 from the outer aureole. Some of those
249 sample descriptions and grain size analyses need to be revised here. Small (10-20
250 μm dia.) grains on the boundaries of relict quartz grains in sample W-71 were
251 originally identified optically as quartz subgrains produced through subgrain
252 rotation recrystallization (SGR). This interpretation was incorrect. In the present
253 study, these small grains were identified using cathodoluminescence (CL) and SEM-
254 EDS as wollastonite and minor K-feldspar. Under CL, wollastonite has a lime-yellow
255 color and K-feldspar is blue. This means that the grain size piezometer for the
256 dynamic recrystallization of quartz used by Morgan and Law (2004) for this sample
257 was incorrect. All samples from the entire traverse exhibit grain boundaries

indicative of recrystallization by Grain Boundary Migration (GBM), which is expected for temperatures of deformation above 500 °C (Stipp et al., 2010).

A more precise definition of recrystallization is needed here to completely describe the differences in microstructure between the inner and outer aureole quartzites. We interpret the grain boundary outlines to indicate that although almost all of the original sedimentary grains have undergone migration in the inner aureole (except for sample 7-50-2 in which wollastonite constitutes ~15%), they have not migrated far. Most grains still have an elongate shape that is defined by a slightly cusped/lobate margin, so that the original boundaries seems to have only migrated slightly (Fig. 5). We define these grains as migrated by GBM although it is apparent that these quartzites have not undergone extensive, sweeping, GBM. In most inner aureole samples there are also a few grains that have parts of their boundaries that are smoothly rounded and we interpret those parts of the grain boundaries as unmigrated by GBM, even though other parts of their boundaries have undergone migration. Some of these rounded grain boundaries have fine wollastonite and or K-feldspar at their margins and were probably pinned. Although some of these rounded boundaries do not have pinning secondary phases at their margins. Our interpretation of the inner aureole microstructures is that pinning occurred at some grain boundaries but it is not the primary variable preventing extensive GBM. In addition, in all samples from the inner aureole, the accessory phases are not aligned or are not strongly aligned and the foliation is not well developed. These microstructures are very different than the outer aureole microstructures where the quartzites are interpreted to have undergone dynamic

recrystallization where the grain boundaries have swept across the sample removing any record of its sedimentary origin.

In the following paragraphs, we document the details of all the quartzite microstructures, beginning with samples in the inner aureole and then moving into the outer aureole.

At 135-137m from the contact where quartzites first appear, the quartzites contain interstitial wollastonite as an accessory phase and almost all grains seem to be slightly modified sedimentary grains. In the first two quartzite samples of the traverse (7-49-3, W-71; Fig. 3a), rare rounded parts of grains are observed where there is a fringe of fine-grained wollastonite on the grain boundary (Fig. 6a) and also where there is no wollastonite on the grain boundary (Fig. 6b). Most grains have an irregular but elliptical shape and boundaries are lobate where one grain has grown into another. In these samples, 99% and 95% of the grains have had their boundaries migrated by GBM (see Table 1).

In the sample collected at 180 m from the pluton (sample 7-50-2; Fig. 3a), wollastonite constitutes approximately 15% of the rock and only a few grains have undergone any amount of GBM. Most grains retain their original round shape (Fig. 7). The quartz grains are dominantly matrix-supported and the matrix is almost completely composed of wollastonite. Very minor amounts of calcite and potassium feldspar have also been identified using CL. There are small areas in this thin section where GBM has occurred and these are the only areas lacking wollastonite between grains. Most of the grains in this sample have slightly embayed margins against smaller wollastonite crystals, suggesting that the reaction that formed

wollastonite, calcite + quartz $\rightarrow$ wollastonite + CO₂, consumed small quantities of quartz at grain boundaries. Locally, wollastonite crystals appear intergrown with quartz grains and extend outwards in a radial pattern that fills the pore space between adjacent quartz grains (Fig. 7). These textural relationships between wollastonite and quartz grains suggest that there has been little to no grain boundary sliding since the wollastonite-forming reaction has occurred.

At 190 m from the pluton contact, the quartzite consists almost wholly of quartz (7-51; Fig. 3a), with only a few small grains of muscovite observed. This sample has been completely recrystallized by GBM. This is the closest quartzite sample to the pluton that exhibits complete recrystallization.

Further from the pluton (between 195-220m from the contact), quartzites contain calcite + diopside and relict quartz grains exist, but all grains have had parts of their boundaries modified by migration. 94% of grains were determined to have undergone GBM. Some original ellipsoidal sedimentary grain shapes are detectable even though their lobate boundaries indicate that some GBM has occurred (Fig. 5). Chessboard extinction patterns of subgrains are also observed. Diopside occurs mostly in clumps of randomly oriented grains.

In the last sample from the inner aureole, collected 225 m from the pluton, 97% of the grains were determined to have undergone GBM. The accessory minerals are indicative of the outer aureole (sillimanite plus muscovite), yet the CPO is indicative of the inner aureole.

Cathodoluminescence (CL) images of quartz grains from the inner aureole indicate that some grains have been intensely fractured, which is not apparent in

plain or polarized light. CL images from samples W-71 (Fig. 8) and 7-50-2 exhibit bright luminescing quartz grains with a network of dark intersecting straight and curved linear features that we interpret as healed fractures. The new quartz in these fractures luminesces much less than the original quartz. Ellipsoidal and irregular (recrystallized ?) grains both possess these fracture patterns. In cross-polarized light images, these microstructures may have been interpreted as indicating that bulging recrystallization or GBM occurred, but under CL, these fractures are apparent and suggest that fracturing and rotation of grain pieces accommodated at least part of the strain.

In summary, the grain boundary textures from all of the inner aureole quartzites indicate that recrystallization occurred by GBM (e.g., Mancktelow and Pennacchioni, 2004). GBM was not extensive – even though most grains have experienced some migration, some original sedimentary boundaries are almost always observable, and most grains are still elongate but now have cusped/lobate boundaries. Some of the boundaries were pinned by secondary minerals but in most cases the lack of extensive and sweeping GBM cannot be explained by pinning. Foliation is subvertical and not well developed and is defined by ellipsoidal grains that have experienced moderate GBM and by a weak alignment of small minor grains. In the one sample with wollastonite on almost all grain boundaries, very little GBM has occurred. CL images indicate that relict quartz grains deformed plastically and cataclastically. GBM microstructures are documented by the interlobate grain boundaries, which occur in all directions, and by the presence of

small accessory minerals that are aligned parallel to the foliation within quartz grains.

6.2.2 Outer aureole quartzites: Samples collected between 250 and 1800 m from pluton contact

The quartzites beginning at 250 m from the pluton contact become interlayered with andalusite or sillimanite schist (although some samples contain both aluminosilicates). Many of the quartzites have abundant clumps of fibrolite. In these samples the microstructure is very different from those in the inner aureole samples as all evidence for original sedimentary grain boundaries is gone. This extensive GBM of samples continues until a distance of 500 m from the pluton contact. The foliation within this part of the outer aureole is partly defined by small, aligned grains of muscovite and biotite that mark the borders of many grains but are also found in the interior of grains. These inclusions indicate extensive and sweeping GBM (Fig. 9). The foliation is more strongly developed than in the inner aureole and grain boundaries that are parallel to the foliation are dominantly straight and long whereas boundaries perpendicular to the foliation are often step-like and shorter. It appears that many foliation-parallel quartz grain boundaries were initially pinned by biotite and muscovite during migration, although with continued migration, grain boundaries were able to grow past these micas.

Other evidence for extensive GBM is that large areas of thin sections have almost the same exact crystallographic orientation (discussed below). Deformation bands are consistently between 30 & 40° to the foliation, although in large areas of

similar crystallographic orientation, chessboard patterns of subgrain formation are observed.

Beginning at 500 m from the pluton, there is evidence once again for relict grains/incomplete recrystallization, although it is not clear whether rounded grain boundaries reflect original sedimentary grains. Samples collected at 660 m again seem to have experienced extensive GBM. Conversely, from 830 m to the end of the traverse at 1540 m, there is evidence for less extensive GBM. Samples collected at 1540 m exhibit well-defined relict grains and these outcrops also preserve skolithos trace fossils, although these samples have fibrolite as an accessory phase and still exhibit CPOs indicative of prism [c] slip.

CL images of quartz grains in the outer aureole exhibit very different microstructures from the quartzites in the inner aureole. These grains exhibit some curved linear features resembling fractures, but they extend across the entire grain and are much less abundant and narrower. In contrast to the fracture patterns exhibited by the grains closer to the pluton, some grains from the outer aureole exhibit a pattern of narrow repetitive curved bands of light and dark luminescence similar to growth rings (Fig. 10). These kinds of light and dark bands have been observed by Piazzolo et al., (2005) in quartzites in the aureole of the Ballachulish igneous complex and have been interpreted as markers of GBM.

7. Crystallographic preferred orientations

7.1 Inner aureole quartzite CPOs (0-225 m from pluton contact)

The CPOs from the inner aureole quartzites exhibit Y-(strain) axis maxima and incomplete cross-girdle patterns, which are mostly symmetrically distributed across the foliation (Fig. 3a). These patterns mostly indicate that basal $\langle a \rangle$, rhomb $\langle a \rangle$, and prism $\langle a \rangle$ slip operated. Opening angles vary between 90 and 100° (Morgan and Law, 2004). These fabrics are weakly to moderately well developed and cannot be used to indicate a sense of shear unambiguously. The one sample without a CPO (7-50-2) has grains that retain their sedimentary grain shapes and also has abundant intergranular wollastonite. The two marble samples from this part of the traverse have well-developed CPOs and exhibit similar patterns of symmetrical distribution of c-axes at 70-90° to the Y-axis (Fig. 3a).

7.2 Outer aureole quartzite CPOs (250 - 1800 m from pluton contact)

The CPOs from the outer aureole quartzites exhibit strong point concentrations of c-axes close to the lineation and symmetrically distributed across the foliation, indicative of prism $\langle c \rangle$ slip. Outer aureole samples also exhibit concentrations of c-axes near the Y-axis, indicating that prism $\langle a \rangle$ slip continued to operate under the same conditions as prism $\langle c \rangle$ slip. These Y-axis concentrations occur in minor amounts throughout the rest of the traverse, although they are less prevalent in the samples that exhibit a stronger foliation and which are more extensively recrystallized (W-68, Fig. 3b, for example).

The CPOs are roughly symmetrical with respect to the foliation, although there is a slight asymmetry to the pattern that is developed across four of the samples collected near the boundary with the inner aureole. Samples W-75, 7-54, 7-

55, and 7-56 (Fig. 3) all exhibit greater concentrations of c-axes that indicate a pluton-side up sense of shear. The remaining patterns are either symmetrical or not sufficiently developed to determine shear sense.

The CPOs determined using EBSD for samples W-68 and W-71 (Fig. 11) are in excellent agreement with the CPOs measured with the universal stage (Fig. 3). The addition of the orientation of the other crystallographic axes allows for a more complete understanding of the slip systems involved and more robust interpretation of the strain symmetry (i.e., plane strain vs. flattening). The significant difference between the two patterns (besides the orientation of the c-axes) is the point distribution of the a-axes in W-71, indicating that the a-axes were involved in the slip process, which is characteristic of samples deformed at low to medium metamorphic grades (Law, 1990). In contrast, the a-axis patterns from W-68 are evenly distributed about the c-axes, indicating that they were not directly involved in slip and that slip occurred along prism planes parallel to the [c] direction.

8. Metamorphic minerals across the aureole

The Mule Spring marbles contain forsterite and globules of euhedral dolomite crystals within a matrix of chrysotile (Nabelek and Morgan, 2012). The euhedral shape of the dolomite crystals along with the spherical shape of the dolomite + chrysotile subassemblage suggest that the globules are crystallized melt pockets within the dolomite matrix, consistent with the high temperatures and evidence for melting in the pelites in the western aureole (Kontny and Dietl, 2002;

Nabelek and Morgan, 2012). The presence of chrysotile requires high X_{H_2O} in the fluid.

In the calc-silicate layers of the Saline Valley Formation, approximately 90 m from the pluton, diopside occurs with grossular garnet $\pm$ vesuvianite (Nabelek and Morgan, 2012). These samples were collected on the eastern margin of the Saline Valley Formation, close to the contact with the Mule Spring Formation. In the quartzites of the Saline Valley Formation, beginning at 135 m from the pluton, the accessory minerals are wollastonite + diopside and wollastonite is found in the quartzites up to 180 m from the pluton. Between 195 and 220 m from the pluton, diopside + calcite are the dominant accessory minerals. Graphite is also found.

Beginning in the Harkless Formation at 225 m from the pluton, the accessory minerals change from calc-silicate to pelitic. The accessory minerals include coarse/crystalline and fibrous sillimanite, andalusite, muscovite, biotite, and K-feldspar. Andalusite is rich with inclusions and is unaltered in some samples and highly altered to sericite and zoisite in other samples. These andalusite grains are rounded or have irregular grain boundaries; none appear euhedral. Coarse-grained sillimanite is observed replacing andalusite. The abundance of fibrolite varies and is not associated with andalusite. The matrix also has biotite and fibrolite defining shear bands that cut the foliation (Morgan and Law, 2004).

Graphitic material is abundant in many quartzites from the inner and outer aureole. In the inner aureole, graphite occurs mostly between grains. In the outer aureole quartzites, graphite appears to have been remobilized and probably recrystallized to more ordered graphite that occurs in more localized patches and

fractures that cut through quartz grains. Although the crystallinity of the graphitic material was not determined, remobilization and increase in crystallinity of graphite is common in metamorphic terrains, but it is usually promoted by interstitial fluids (Dunn and Valley, 1992; Bergfeld et al., 1996; Huff and Nabelek, 2007).

9. Oxygen isotopes

Oxygen isotope ratios ($^{18}\text{O}/^{16}\text{O}$) were analyzed in specific quartzites along the traverse to obtain information regarding the degree of interaction of quartzite with fluids during the deformation and metamorphism associated with the intrusion. $\delta^{18}\text{O}(\text{Qtz})$ values, along with the progress of mineral reactions, are the two most used indicators of fluid sources and fluid flow (e.g., Lackey and Valley, 2004; Nabelek, 2002; Ferry et al., 2001). Lattice diffusion and dissolution-reprecipitation are the processes traditionally assumed to be responsible for isotopic exchange between minerals and fluids, although it is increasingly apparent that grain boundary migration can be a more efficient means of isotopic exchange (Nakamura et al., 2005).

$\delta^{18}\text{O}(\text{Qtz})$ values obtained for inner aureole quartzites (first 225 m) are between 13.4 and 16.5‰ VSMOW (Fig. 3a), and four of the five values are above 14.4‰. This includes the samples from the Saline Valley Formation as well as one sample collected from the first 10 m of the Harkless Formation (which includes a marble layer near the top of the formation). All of the samples collected from the outer aureole quartzites have values between 13.8 and 12.57‰ (Fig. 3b). Two quartzite samples from the Harkless Formation collected 16 km from the pluton

have $\delta^{18}\text{O}(\text{Qtz})$ values of 15.5 and 15.2‰ and are assumed to represent the pre-pluton emplacement values.

10. Temperatures of deformation

Peak temperatures that were reached in the aureole are discerned by several ways. First, the presence of sillimanite in the pelitic units in the outer aureole indicates that the temperatures were above 600°C, based on the inversion of andalusite to sillimanite (Kerrick, 1990) at pressures of 2.5 – 3.5 GPa (Nabelek and Morgan, 2012). The pressure estimate comes from the near-coexistence of andalusite, sillimanite, and now retrograded cordierite in parts of the aureole. The operation of prism [c] slip is indicative of temperatures greater than 600°C (Lister and Dornsiepen, 1982; Bouchez et al., 1984; Mainprice et al., 1986; Stipp et al., 2002). Chessboard patterns of subgrains in the quartzites indicate that temperatures were above 650° C (Kruhl, 1996). The presence of melting in the H_2O saturated Harkless and Saline Valley schists on the western side of the pluton indicates minimum temperatures of 660°C (Nabelek and Morgan, 2012). Na-in-cordierite thermometry suggests temperatures of ~730°C in the southern part of the aureole (Kontny and Dietl, 2002). Using the opening angle thermometer of quartz CPOs, Morgan and Law (2004) obtained temperatures from three samples that vary between 680°C and 740°C $\pm$ 50°C for the inner aureole quartzites.

11. Discussion

11.1 Summary of results

Table 1 summarizes the differences between the inner and outer aureole quartzites. There is a clear distinction between the degree of GBM, type of CPO, minerals present, and $\delta^{18}\text{O}$ values from the inner aureole quartzites to the outer aureole quartzites. The inner aureole quartzites exhibit sedimentary grain shapes, lower degrees of GBM, and CPOs typical of lower amphibolite facies metamorphism even though these rocks are closer to the pluton. The outer aureole quartzites exhibit microstructures indicative of complete recrystallization and extensive GBM and prism [c] slip CPOs indicative of higher temperatures (600°C +). We suggest that the inner aureole quartzites experienced higher temperatures but that a lack of water or its dilution by CO_2 on the grain boundaries prevented these rocks from undergoing extensive GBM. The low water fugacity also prevented prism [c] slip from occurring. This apparent low water activity is supported by the metamorphic mineral assemblages and by the stable isotopes discussed below. It is also clear that in some cases, grain boundaries were pinned by accessory minerals. Although, many of the unmigrated portions of grain boundaries have no accessory minerals to pin them and the grain boundaries that have experienced migration in the inner aureole have only migrated a minor amount, so that many of the grain shapes are still identifiable. Therefore, the low water fugacity may not actually stop GBM but it seems to drastically slow it down. Below we use the metamorphic mineral assemblages and the stable isotopes to argue for a dry (i.e., low water fugacity) inner aureole and then discuss the implications for GBM and deformation.

11.2 Fluid Evolution

While the evidence for fluid transport within the EJB aureole is somewhat variable depending on rock type, the data mainly support very little influx of fluids from either magmatic or meteoric sources into either the inner or outer aureole. We argue that the only fluids that were available were the fluids present initially in pores, along grain boundaries, in fluid inclusions, and the fluids that were produced during prograde mineral reactions. Within the inner aureole, these mineral reactions may have buffered the fluid composition so that X_{CO_2} increased to a level where the grain boundaries were effectively dry and resulted in the pinned/slow GBM and apparent low temperature slip-systems.

Nabelek and Morgan (2012) argued that narrow, mostly primary ranges in the $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of the Mule Spring marble (20-23‰ and 0.5-2‰, respectively) show no to very little isotopic exchange with an external fluid even though these marbles have been completely recrystallized and are highly deformed. In contrast, the limited $\delta^{18}\text{O}$ shift in calc-silicates from 20 to 16‰ in the Saline Valley Formation, near the contact with the Mule Spring marble, is consistent with some decarbonation and interaction with aqueous fluids that equilibrated with the host silicate rocks at large rock/water ratios.

The $\delta^{18}\text{O}$ values of the quartzites of the Saline Valley Formation, located within 250 m of the contact (Table 1), are mostly very similar to only slightly lower than values of the quartzites collected 22 km away from the EJB pluton (15.2-15.5‰). Although the distal samples are of the Harkless Formation, it would be fortuitous if an exchanged Saline Valley Formation had the same values as unexchanged Harkless Formation. Thus, the $\delta^{18}\text{O}$ values suggest that the inner

aureole quartzites have undergone very little exchange with magmatic or meteoric, water-rich fluids since their regional low-grade metamorphism prior to pluton emplacement. The exception is sample 49-3 (13.4‰), which exhibits minor GBM. All samples at distances >250 m from the contact have $\delta^{18}\text{O}$ values $\leq 13.8\text{‰}$, which are consistent with GBM and recrystallization that was assisted by larger amounts of water.

The mineral assemblages found in the inner aureole are somewhat mixed but mainly document a low $f\text{H}_2\text{O}$ environment. The assemblage of garnet + vesuvianite and down-shifted $\delta^{18}\text{O}$ in the Monola Formation at the contact documents focused flow of aqueous fluids in the calc-silicates (Nabelek and Morgan, 2012). The presence of wollastonite in the quartzites in the inner-most Saline Valley Formation is probably the result of the reaction between calcite cement and host quartz of the original sandstones, which is consistent with deposition of these sandstones between and just below the limestone beds of the Monola and Mule Spring formations. Because of the increasing $X(\text{CO}_2)^{\text{fluid}}$ with T of the wollastonite-producing reaction, it is likely that the CO_2 produced by the reaction rapidly displaced or diluted the intergranular water as temperature increased (Morgan and Law, 2004). The formation of interstitial diopside probably involved the reaction of a dolomite or high-Mg calcite cement with quartz, which also would have generated CO_2 .

Given that temperatures in the aureole exceeded 660°C (Kontny and Dietl, 2002; Morgan and Law, 2004; Nabelek and Morgan, 2012) and that reactions should progress rapidly under water rich conditions, it is likely that the inner aureole

deformation and mineral reactions all occurred very early in the prograde event while the interstitial fluid was still H₂O-rich and then stopped once the fluid became CO₂-rich. The observation that the grain boundaries have undergone limited migration is consistent with the limited isotopic exchange in the inner aureole that is generally facilitated by mineral recrystallization or GBM (Nabelek et al., 1992; Kirschner et al., 1995; Nakamura et al., 2005). Besides the apparent low temperature CPOs in the inner aureole quartzites, our model is supported by the evidence for a lack of deformation in the wollastonite-rich quartzite post-dating wollastonite growth. Finally, there is no mineralogical evidence for a late convective aqueous/magmatic fluid influx that can develop next to shallow plutons (Cui et al., 2003, Nabelek, 2007). We suggest that the lithostatic fluid pressures that likely existed at the depth of emplacement of the EJB pluton (Chen and Nabelek, 2014) and the thick sequence of Mule Spring marbles acted to prevent a late fluid influx. Marbles are commonly interpreted as fluid blocks (aquitards), as evidenced by their unshifted C and O isotopic ratios (Nabelek et al., 1984; Nabelek and Morgan, 2012) and also based on textures (Holness and Graham, 1995).

In the outer aureole, including the Harkless Formation, the $\delta^{18}\text{O}$ isotopic shift from the regional values is greater than in the inner aureole but not by much. This slightly greater shift is consistent across the outer aureole. Samples from the outer aureole vary between 12.9 and 13.8‰. The pelitic mineral assemblage in the interbedded pelites, including muscovite, biotite, andalusite $\pm$ sillimanite, all support a fluid composition dominated by water. Therefore, the water available to

the quartzites may have been pore water and the water produced by prograde reactions in the interbedded pelites.

11.3 Slip systems

Prism [c] slip has only been documented to occur in naturally deformed quartzites when the temperatures of deformation exceeded $\sim 600^{\circ}\text{C}$ and there was water present (Lister and Dornsiepen, 1982; Blumenfeld et al., 1986; Mainprice et al., 1986; Garbutt and Teyssier, 1991; Okudaira et al, 1995). Experiments on single quartz crystals document that an increase in water promotes a switch from basal $\langle a \rangle$ to prism [c] slip systems (Griggs and Blacic, 1965). Water also diffuses preferentially along the c direction within quartz (Linker and Kirby, 1981). Therefore, in environments where the activity of water is low, prism [c] slip may be suppressed even though the temperatures are high (e.g., Hippertt et. al., 2001).

It is possible that the CPOs in the inner aureole are characterized by $\langle a \rangle$ slip simply because they formed at lower amphibolite facies temperatures early on in the deformation history and then dislocation creep was shut-off at higher temperatures by the lack of water. Or, it is possible that dislocation creep continued to operate at higher temperatures but that lack of water increased the critical resolved shear stress for [c] slip so that $\langle a \rangle$ slip was favored.

Stipp et al. (2002) also observed a correlation between a switch in recrystallization mechanisms and slip systems in the Tonale Fault zone caused by a temperature gradient across the zone due to the synchronous intrusion of the Adamello batholith. Within the zone of rocks exhibiting GBM, Stipp et al.,(2002)

noticed two types of GBM microstructures – pinned and unpinned. The transition from pinned GBM to more active GBM is also accompanied by a change in slip systems. The dominant slip system that operated during GBM is prism $\langle a \rangle$ slip. Slip systems that operated during unpinned GBM are a combination of basal, prism $\langle a \rangle$ and prism $[c]$ slips.

Mancktelow and Pennacchioni (2004) examined “wet” and “dry” Alpine quartzite mylonites and observed that fluids controlled the type of recovery/recrystallization mechanisms but noted that fluids did not control the slip systems. Lack of water also reduced the mobility of grain boundaries. The dry quartzites recrystallized by subgrain rotation and grain boundary bulging developed mean grain sizes in the tens of microns even though they were deformed in the temperature range associated with GBM (500-550° C, Stipp et al, 2002). In contrast, the wet quartzites recrystallized by GBM and grain sizes are on the order of millimeters. CPOs from both sets of quartzite mylonites are dominated by $\langle a \rangle$ slip fabrics, representative of their temperatures of deformation. Direct comparison between the results of Mancktelow and Pennacchioni (2004) and ours for the EJB aureole quartzites is not possible because the lower temperatures of the Alpine deformation did not allow for prism $[c]$ slip. Regardless, their study confirms the importance of water in controlling grain boundary processes in naturally deformed rocks.

12. Conclusions

645 The microstructural, petrologic, and isotopic data collected from the high-strain,
646 high-temperature aureole of the EJB pluton indicates that fluid composition (H_2O
647 versus CO_2) was the most important variable controlling changes to the deformation
648 mechanisms and recovery of quartzite as distance increases from the pluton contact.
649 Quartzites in the inner aureole (closer to the pluton) are incompletely recrystallized
650 and the microstructures are indicative of minor amounts of GBM or “slow” GBM and
651 CL images suggest fracturing along with crystal plastic deformation. CPOs are
652 indicative of $\langle a \rangle$ slip. Oxygen isotopes are only slightly shifted from regional values.
653 The pore fluids in the inner aureole were probably initially water-rich. However,
654 during prograde reactions in the intervening calc-silicate rocks, and perhaps more
655 importantly, between calcite cement and quartz in the quartzites, the pore fluid
656 composition in the inner aureole changed to become dominated by CO_2 , which acted
657 as a non-wetting phase and decreased the available water needed for grain
658 boundary mobility. The lack of water also suppressed the activity of prism $[c]$ slip.
659 Therefore, we propose that dry conditions or a grain boundary fluid with a
660 significant non-wetting component (CO_2) can result in apparent temperatures of
661 deformation that are more than 100 °C lower than the real temperatures of
662 deformation. The outer aureole deformation and recrystallization occurred as
663 expected for temperatures above 600 – 650 °C. Recrystallization is complete and
664 CPOs are indicative of prism $[c]$ slip systems. Oxygen isotopes in the outer aureole
665 were shifted more than in the inner aureole, but only slightly, which is consistent
666 with equilibration with locally derived fluids. We suggest that recrystallization in

the outer aureole was aided by pore water, water derived from inclusions, and water generated by prograde reactions in the surrounding pelitic schists.

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FIGURE CAPTIONS

Table 1: Note: Skolithos trace fossils are observed in sample 9-3, the sample collected furthest from the pluton.

Figure 1: Geologic map of the Eureka Valley (EV) - Joshua Flat (JF)- Beer Creek (BC) (EJB) pluton in the White-Inyo Mountains, California (after Morgan and Law 2004) and location of sampling traverse on NE margin. Sample traverse shown in detail in Figure 2. Both sets of folds (large scale N to NW-trending synclines and anticlines and smaller scale NE-trending folds) predate emplacement of plutons and produced structural domes and basins (B1-B5). (After Morgan and Law, 1998). BiC, Birch Creek pluton; SH, Sage Hen pluton.

Figure 2: Geologic map of sample locations along the traverse on the NE margin of the EJB pluton.

Figure 3a: Cross-section of the western portion of traverse with location of samples collected between 0 m and 250 m from the pluton contact and corresponding CPOs with $\delta^{18}\text{O}(\text{Qtz})$ values. Modified after Morgan and Law (2004). Note that CPOs are presented with foliation vertical and perpendicular to the page and lineation vertical, similar to how samples were collected along the traverse.

Figure 3b: Profile of entire traverse with location of samples collected between 250m and 1800 m from the pluton contact and corresponding CPOs with $\delta^{18}\text{O}(\text{Qtz})$ values. Note that the scale of the traverse is different than in Figure 3a.

Figure 4: Photomicrographs of the Harkless quartzite collected 22 km SW of the EJB pluton. B is a close-up of the grains in A. Microstructures are indicative of greenschist facies regional deformation. Note that some of the grains have rounded grain boundaries as well as abundant evidence for dissolution creep along other grain boundaries.

Figure 5: Photomicrograph of W-72 collected from the inner aureole (between 210 m and 250 m from the pluton contact). Minor phases include diopside and calcite. Grain in the center has undergone minor amounts of GBM on all margins but still retains evidence of its elongate sedimentary shape. Foliation is not well developed and grain boundaries are lobate/cusped. Note the well defined subgrain boundaries within the grain. All the quartz surrounding the central grain is under extinction and therefore dark in this image. Grain in center is 1 mm long.

Figure 6: Photomicrographs of quartzites from the first 200 m of the traverse where relict sedimentary grains are observed. A. Sample W-71. Rounded grain in the center of the photograph has undergone GBM on the lower margin while top and right margin has not. Top and right margin is decorated with wollastonite and is evidence for grain boundary pinning. B. Sample 7-49-3. Rounded grain in the center of photograph has a partly migrated boundary (GBM) and a partly unmigrated boundary (arrows) yet has no other minerals pinning the boundary.

Figure 7: Photomicrograph of quartzite sample 7-52-2. Wollastonite is found on the margins of most grains and most grains retain their round shapes although slightly embayed by the quartz consuming reaction to form wollastonite from calcite.

Figure 8: Photomicrographs and CL images taken from sample W-71, collected 140 m from the contact. A. Partial recrystallization by GBM. Box in upper left corner is the grain shown in Figure 6a. B. CL image taken from center of A (box) which exhibits evidence of fracturing. C. Same area as B but in cross-polars.

Figure 9: Photomicrograph of representative quartzite (W-75) collected from the outer aureole (between 250 m and 1800 m from the pluton contact). Minor phases include biotite. Recrystallization is complete and the foliation is well developed. Grains have overgrown inclusions and small grains of biotite and some grain growth was controlled by the foliation indicative of extensive GBM.

Figure 10: CL images of sample 9-8 collected 500 m from the pluton contact in the outer aureole. Alternating bands of light and dark are interpreted to be a result of grain boundary migration.

Figure 11: EBSD patterns of the c, a, m, r, and z poles from one sample (W-71, 140 m

951 from the pluton contact, see Fig. 8 for photomicrograph) from the partially recrystallized
952 quartzites and from one sample (W-68, 250 m from the contact) from the totally
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954 perpendicular to the page and lineation vertical, similar to how samples were collected
955 along the traverse.
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Table

	Inner Aureole						
sample	7-49-3	W-71	7-50-2	7-51	W-72	W-73	7-53-2
distance from pluton (m)	135	137	180	190	195	220	225
%GBM	99%	95%	0	100%	94%	96%	97%
relic grains?	yes	yes	yes	no	yes	yes	yes
Accessory minerals	wollastonite + diopside	wollastonite	wollastonite	musc	diopside + calcite	diopside + calcite	fibrolite + musc
Slip system from CPO	Prism Basal	Prism Basal	none	Prism	Prism Basal	Prism Basal	Prism Rhomb
Oxygen isotope ($\delta^{18}O$)	13.4	14.4	16.5	14.5			14.7
foliation	weak	weak	weak	weak	weak	weak	weak

	Outer Aureole												
sample	7-54	W-75	7-55	7-56	W-68	9-8	9-7	9-6	9-5	9-4	9-2	9-1	9-3
distance from pluton (m)	250	265	290	300	370	500	660	830	1000	1040	1200	1300	1540
%GBM	100%	100%	100%	100%	100%	99%	100%	99%?	99%?	99%?	93%	88%	90%
relic grains?	no	no	no	no	no	?	no	?	?	?	yes	yes	yes
Accessory minerals	andalusite + musc	K-feldspar + musc + biotite	andalusite + fibrolite + musc	fibrolite + musc + biotite	fibrolite + musc + biotite	musc	K-feldspar + musc	musc.	fibrolite + musc + biotite	fibrolite + musc + biotite	biotite + musc.	musc. + biotite	fibrolite + musc
Slip system from CPO	Prism Basal Prism	Prism Basal Prism	Prism Basal Prism	Prism Basal Prism	Prism	Prism Prism	Prism	Prism Basal	Prism Basal	Prism Basal	no CPO	?	Prism
Oxygen isotope ($\delta^{18}O$)	13.8			13.2	12.9	12.1	13.3	13.6	13	13.2	13.2	12.8	12.7
foliation	moderate	strong	strong	strong	strong	weak	weak	weak	weak	weak	weak	weak	weak

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W-71 N = 1229

W-68 N = 537

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Sedimentary grain affected by minimal amount of grain boundary migration (GBM) on all margins but still retains evidence of its elongate sedimentary shape. Grain boundaries are lobate/cusped and the interior exhibits well defined subgrain boundaries. Surrounding grains have all experienced more extensive GBM. Grain is 1 mm long.