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Title: Fabrics produced mimetically during static metamorphism in retrogressed eclogites from the Zermatt-Saas zone, Western Italian Alps

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Abstract

Lattice preferred orientations (LPOs) are commonly interpreted to form by dislocation creep. Consequently they are used to infer deformation at the metamorphic grade at which the minerals were stable, especially if those minerals show a shape fabric. Here we show that LPOs can occur through mimicry of a pre-existing LPO, so they formed statically, not during deformation. Omphacite and glaucophane LPOs occur in eclogite facies rocks from the Zermatt-Saas Unit of the Northwest Italian Alps. Barroisite grew during greenschist facies retrogression and has an LPO controlled significantly by the eclogite facies omphacite and glaucophane LPOs, rather than directly by deformation. Using spatially resolved lattice orientation data from the three key minerals, collected using electron backscatter diffraction, we deploy a new technique of interphase misorientation distribution analysis to prove this. Barroisite LPO develops by mimicry of omphacite (via a particular lattice orientation relationship) and by direct topotactic and epitactic replacement of glaucophane. LPO in turn
influenced anisotropic grain growth, resulting in a barroisite grain shape fabric. Thus regional retrogression during exhumation of the Zermatt-Saas high-pressure rocks was, in large part, static, rather than dynamic as previously interpreted. In general the possibility of mimetic fabrics forming during metamorphic reactions must be borne in mind when interpreting direct structural observations and seismic anisotropy data in terms of deformation, in both crust and mantle.

**Keywords:** interphase misorientation distribution, mimetic, lattice preferred orientation, eclogite, omphacite, Alps

1 **Introduction**

Lattice preferred orientations (LPO) of minerals in metamorphic rocks are commonly interpreted to result from dislocation creep at conditions where those minerals are stable (Bascou et al., 2001; Brenker, 1998; Brenker et al., 2002; Buatier et al., 1991; Philippot and Van Roermund, 1992). Consequently they can be used to deduce the conditions of deformation. When grains are also elongate the evidence for crystal plasticity is persuasive. However, when minerals grow at different stages in a rock’s evolution, the new minerals may nucleate and grow on existing ones such that the crystallographic orientation of the new mineral is related to that of the one it is replacing – mimetic growth. This is particularly evident when the two minerals have some common structure, e.g. chlorite replacing biotite, though that is not essential: for example orthorhombic sillimanite may have orientation controlled by monoclinic (pseudohexagonal) biotite (Yardley, 1977) and orthorhombic andalusite may be controlled by triclinic kyanite (Wheeler et al., 2004). Similar ‘special orientation relationships’ are noted in material sciences. The relation between bcc-Fe and fcc-Fe, the Kurdjumov-Sachs relationship (Kurdjumov and Sachs, 1930), is key to understanding microstructural evolution during steel manufacture. Studies of materials such as Nb solid
solution precipitates, Nb$_5$Si$_3$ intermetallics, materials in the Zr-ZrB$_2$ system and WC-Co alloys (Bounhoure et al., 2008; Champion and Hagege, 1992; Cheng et al., 2009) show that local interfacial configurations between materials organise to minimise interfacial energies, resulting in special orientation relationships forming. It has even been shown that orientation can be predicted in epitactical growth in fcc-bcc interfaces (Gotoh and Arai, 1986).

Two styles of mimetic relationships can exist: expitaxy and topotaxy. Epitaxy refers to an overgrowth, oriented with respect to the reactant, whether it is simply growing around or as a result of a reaction with the substrate. Topotaxy requires the conversion of a reactant into a new oriented product (Shannon and Rossi, 1964) and involves any chemical, solid state reaction that leads to the formation of a new material with a crystal orientation that correlates to that of the initial material (Lotgering, 1959) i.e. replacive to some extent. Topotaxy refers to all oriented conversions of a reactant to a product with the exception of those that are epitactic. Within this definition there exists a continuum of degrees of topotaxy depending on the amount of replacement.

Once the possibility of mimetic growth is accepted, it follows that a pre-existing LPO can be inherited (in modified form) by a developing phase. For example, suppose that the c-axis of an early mineral controls the c-axis of a new mineral, it is likely the new c-axis fabric will resemble the old one. For example, Seward et al. (2004) show that a polycrystal of hexagonal Ti, with strong LPO, gives rise to a polycrystal of cubic Ti with a very strong fabric that can be explained by the “Burgers orientation relationship” between the two phases. This aspect of fabric formation remains relatively unexplored in the geological literature. This study is motivated by the importance of understanding whether successive metamorphic events (at different grades) are static or dynamic, and by our own experience of strong mimetic controls on LPO in metals such as Ti.
We targeted partially retrogressed eclogites from the Gressoney Valley area of the Zermatt-
Saas unit, Western Alps (Figure 1). These were chosen because petrography shows that
omphacite, glaucophane and barroisite all have strong shape fabrics, and the degree of
retrogressive overprinting varies, so a spectrum of microstructural states can be explored.

2 Geological Context

The Zermatt-Saas Unit is a layer of ophiolitic material obducted onto the European margin
during the Alpine Orogeny. The protolith is Late Jurassic, oceanic lithosphere which formed
the crust of the Piemontese (Tethyan) Sea that originally separated the older continental crust
of the European continent and the African margin (Amato et al., 1999). The area of the
Zermatt-Saas Unit studied here has experienced high-pressure metamorphism, at least up to
eclogite facies. Pressure and temperature estimates are around 1.75-2.0 GPa and 550-600 °C
(Barnicoat and Fry, 1986) but higher pressures ~2.5-3.0 GPa are recorded from the Pfulwe
area, 15 km north of the study area (Bucher et al., 2005). Timing of peak pressure is dated
using garnet at 48 - 41Ma (Lapen et al., 2003; Amato et al., 1999). The Zermatt-Saas unit lies
structurally above the continental Monte Rosa unit, also metamorphosed to eclogite facies
dated at 42 Ma from rutile, indicating a common metamorphic history for both units (Lapen et
al., 2007). The Combin Zone, a pervasively deformed greenschist facies unit, lies structurally
above the Zermatt-Saas. This geological contact represents a significant metamorphic break,
between eclogite facies rocks (with lower grade overprints) below, and greenschist facies
rocks above, which have never experienced eclogite facies. The pervasive foliation present in
the Combin Zone is diagnosed as an extensional shear zone (the Gressoney Shear Zone -
GSZ) by Reddy et al., (1999) which unroofed the eclogite facies rocks beneath from 60-100
km depth to 30 km.
In the area of the Zermatt-Saas studied here, Punta Telcio, lithologies comprise metabasic and metagabbroic rocks, serpentinites and small amounts of metasediments. Metabasic rocks contain high-pressure minerals omphacite and garnet, though in many areas this assemblage is retrogressively overprinted. Two main retrogressive events are noted in the Zermatt-Saas Unit. The first is a blueschist facies overprint (1.2-1.5 GPa and ≤ 500 °C) noted by growth of glaucophane and the second event is an amphibolite/greenschist event (<1.0 GPa and < 550-500 °C) (Fry and Barnicoat, 1987; Reinecke, 1998; van der Klauw et al., 1997). The growth of blue-green, slightly sodic amphibole, barroisite, is likely associated with the early stages of the second stage of retrogression given the stability range of barroisite (400 – 500 °C and 4 – 10 kbar) (Otsuki and Banno, 1990). Retrogression probably occurred during movement on the GSZ, the base of which is less than 1 km above the study area (Fig. 1). While the Combin Zone (GSZ) shows pervasive high strain, understanding the extent of deformation in the eclogites during exhumation is dependent upon correct interpretation of the fabrics present in the retrogressive phases.

3 Petrography

The Zermatt-Saas Unit in the Gressoney Valley is well exposed on Punta Telcio. High-pressure assemblages exist in patches of fresh eclogite and partially retrogressed eclogite, surrounded by amphibolites. Samples collected for this study represent various transitional stages of the conversion from eclogite facies, through a blueschist event, into a widespread amphibolite/greenschist facies event. The following is a description of the main petrographical features of these rocks important to the aim of this paper.

Eclogite facies assemblages in rocks from Punta Telcio are composed mainly of omphacite and garnet. Omphacite typically has a small grain size (100 - 600μm), and rims of fine-grained symplectite. Omphacite forms a shape fabric, which defines the foliation of the rock.
Figure 2). This omphacite shape fabric wraps around garnet grains in the rock that have variable morphologies; porphyroblastic, poikiloblastic and atoll forms and euhedral to irregular shapes with variable grain sizes (300 µm to 2 mm).

While prograde glaucophane is documented in the Zermatt-Saas (Barnicoat and Fry, 1986) it is not observed in the Punta Telcio assemblage. Glaucophane that occurs in Punta Telcio samples is representative of the first retrogressive phase of the Zermatt-Saas. It contains common garnet, omphacite and zoisite grains inclusions showing it postdates the prograde assemblage. Glaucophane displays two morphologies, the first being large (0.5-2.5 mm), elongate grains that define a shape fabric parallel to the omphacite fabric (Figure 2). These grains may be isolated or form a connected grain network. The second morphology consists of large (1-3 mm), subhedral nematoblasts of glaucophane that cut the omphacite fabric obliquely.

Barroisite appears in a variety of places in the partially retrogressed eclogite and is representative of the second stage of retrogression experienced by the unit. Barroisite forms reaction rims around garnets and also occurs as inclusions within garnet. It forms reaction rims around glaucophane grains and in many places epitaxial and topotactical overgrowth of barroisite on glaucophane is observed (Figure 2). The distinction between pale purple glaucophane and darker blue-green barroisite is clear, whilst the similar crystallographic orientations are obvious optically from the near-parallelism of extinction positions. Barroisite grains growing around garnet and replacing glaucophane vary widely in size (50 µm – 1 mm). A second stage of finer-grained barroisite growth is distinguished replacing the first stage.

Barroisite also occurs as completely separate, single grains or as patches of smaller grains within the matrix of the rock. Single grains can have sizes between 1 mm and 200 µm. Patches of smaller grains have sizes <100 µm. Matrix grains have a shape fabric parallel to
omphacite and glaucophane fabrics. Rather than forming an interconnected network as omphacite and glaucophane do, barroisite exists as isolated grains or patches that sometimes form discontinuous, interconnected networks.

The petrography described here holds the possibility that any LPO found in omphacite may have a mimetic control on any LPO developed in retrogressive glaucophane and barroisite. Likewise any LPO developed in the glaucophane may have a mimetic control on the later appearance of barroisite.

4 Methods

4.1 EBSD Data Collection

All crystallographic orientation data is collected by EBSD at the Liverpool University Microstructure Research Laboratory using a CamScan X 500 crystal probe scanning electron microprobe (SEM) equipped with a thermionic field emission gun and a FASTRACK stage. An accelerating voltage of 20 kV with a typical beam current of ~45-50 nA is used. The angular resolution of this technique is typically better than 1° and spatial resolution is ~0.1µm. All of the data acquisition is carried out automatically using either a stitched matrix of maps with a typical step size of 1.5-3 µm (mapped by moving the electron beam) or by using the FASTRACK stage to collect data on rectangular grids with 2.5-10 µm spacing (mapped by moving the stage) (Prior et al., 2002). Also orientation data across an entire thin section is collected using the FASTRACK stage with spacing of 300-350 µm.

4.2 EBSD Data Processing

Electron backscatter patterns (EBSP) are indexed using the software package CHANNEL+ v5 from Oxford Instruments Ltd. Initial processing on beam mapped grains and stage mapped grains (with spacing between points ≤10 µm) involves the removal of isolated points that have
been incorrectly indexed (wild spikes). What then follows is the processing procedure laid out in (Prior et al., 2009) which involves manipulating the band contrast (BC) of the scanned area. For this study most of the EBSD maps have noise reduction carried out for non indexed points that have at least five neighbouring indexed points.

The similar crystallography (Table 1) of barroisite and glaucophane creates leads to indistinguishable EBSPs between them causing software to often index the correct orientation but the wrong phase (Figure 3). This problem has no effect on crystal orientation measurements as confirmed by comparison of the mean angular deviation (MAD) of both phases, which are approximately the same for each solution, and the near identical Euler angles that each solution provides within a 1.5° error bar.

This misindexing problem is resolved by performing EBSD scans alongside energy dispersive X-ray (EDX) scans. This provides each point with orientation data, an EBSP phase diagnosis and also element abundances. Chemical maps show a strong contrast in calcium content between barroisite and glaucophane forming a reliable way of distinguishing them. Data is run through a MATLAB program that changes the phase identification based on upper and lower limits imposed on the calcium count. The resulting output has the barroisite and glaucophane data points correctly distinguished.

4.3 Intraphase and Interphase Misorientation

To search for special orientation relationships, we must examine the relative orientations of phase A and phase B where they touch. Before describing “interphase” misorientation analysis, we must review the more basic idea of “intraphase” misorientation analysis.

EBSPs provide the full crystallographic orientation of a point allowing calculation of the rotation required to transform the orientation of that point onto the orientation at another point.
As the crystal lattice parameters are the same at both points, a rotation exists that maps all axes and planes from one point onto the other. The rotation which encompasses an axis and angle is termed the misorientation (Wheeler et al., 2001). The number of misorientation solutions that can be calculated depends on mineral symmetry. Lower symmetry minerals have fewer misorientation solutions, i.e. triclinic minerals have one solution whereas cubic minerals have 24. Convention uses the rotation with the smallest angle. This practise allows examination of misorientations across, for example, individual grain or subgrain boundaries. When concerned with one phase and the misorientations within that one phase we refer to this idea as intraphase misorientation.

Interphase misorientation differs in that it is the transform of the crystal orientation at a point of a certain mineral onto the crystal orientation at a point of a different mineral. As the lattice parameters of both phases may be different, there is no longer a unique rotation that can be defined. In general this is true even if the two phases have the same symmetry, e.g. if both phases are monoclinic, but the \( \beta \) angle is different, a rotation resulting in all crystal axes in phase A and phase B being parallel does not exist. Consequently there are choices to be made about which axes and/or planes are to be compared between phases. These choices may be governed by prior knowledge of possible mimetic relationships, hypotheses based on the atomic structures or they may be made arbitrarily to explore possibilities. As all the minerals considered for this study (omphacite, glaucophane and barroisite) are monoclinic we choose to compare corresponding axes. While the unit cell lengths and \( \beta \) angles of all three minerals are similar there are small differences, which may influence comparison between them. Table 1 shows the unit cell lengths and \( \beta \) angles used to index the three minerals. It is important to note, from a geometric view, a unit cell description of a monoclinic lattice is still an equally valid description if we rename a as c and c as a (with a flip of b-axis orientation to maintain handedness). It should not be assumed that a axes will “prefer” to be parallel: this sort of
10 assumption should be justified in terms of atomic structure, and/or tested against other possibilities.

In what follows, it is essential to bear in mind that some crystal directions are “polar”: for example in a monoclinic mineral the b axis [010] is not symmetrically equivalent to [0-10], whilst the a axis [100] is symmetrically equivalent to [-100]. Consequently, in comparing two grains the b axes can be up to 180° apart whilst the a axes cannot be more than 90° apart.

Here we propose two different methods to explore interphase misorientation. In the first, a single direction or plane in phase A and a single direction or plane in phase B are chosen. The angle between the directions (abbreviated D-D) or planes (P-P) is calculated. This study investigates D-D relationships for <100>, <010> and <001>, and P-P relationships for (100) and (001). Note that the P-P angle for (010) will be the same as the D-D angle of <010> and so is omitted. The second type of calculation chooses a plane and a direction in that plane for phase A, and similarly for phase B (abbreviated PD-PD). A rotation is imposed so that the planes and directions are made parallel. One can envisage this as two rotations, one to align the planes and a second to align the directions in those planes. For each of the three planes, there are two directions that lie within it, therefore allowing construction of six PD-PD relationships. However, given the symmetry of monoclinic phases, not all of six relationships are independent. For example, rotating <001> in (100) in a monoclinic mineral so that it is aligned with <001> in (100) of another monoclinic mineral, <010> automatically becomes aligned as well since the angle between <001> and <010> is always 90°. This means that (100)<010> and (100)<001> PD-PD maps show the same information. There are 4 independent combinations; (100)<001>, (100)<010>, (001)<010> and (001)<100>. With (010)<100> and (010)<001> PD-PD maps, slight differences exist between the two as when one direction is lined up between two monoclinic phases, the other direction will not due to
slight variations in the $\beta$ angles. As this difference is quite small it produces negligible differences between the two maps and as such both can be represented by one map.

### 4.4 Intraphase and Interphase Misorientation Angle Distribution Analysis

Intraphase misorientation angle distribution analysis takes neighbour-pair misorientation angle distributions and random-pair misorientation angle distributions and tests for statistically significant differences between them. The differences between the two distributions hold important implications for interpreting microstructures (Wheeler et al., 2001). If a statistically significant difference exists then neighbouring grains have undergone a physical interaction and/or neighbouring pairs have been derived in a significant number of cases from a common parent microstructure: inheritance.

Here we propose that *interphase* misorientation angle distribution analysis can be conducted in the same fashion and aid in the investigation of mimetic fabrics, by investigating whether a statistically significant difference between neighbour pair and random pair misorientation distributions exists across a phase boundary. Interphase neighbour pair misorientation angle distributions are produced through a purpose-made MATLAB program. This program systematically searches the EBSD map for a pixel of a desired phase that has a neighbouring pixel of a desired second phase. Searching continues until all such touching pixel pairs have been found. For this study phase pairs are omphacite/barroisite and glaucophane/barroisite.

The program then takes the orientation data, in the form of Euler angles (Euler 1, Euler 2 and Euler 3), of the touching pixels and converts them to the $a$, $b$, and $c$ axes (as unit vectors) of a monoclinic mineral (the crystal system of all phases concerned with this study) with a given $\beta$ angle (Table 1). The calculation is done using the direction cosine matrix (DCM) (Ulrich and Mainprice, 2005) which rotates the direction from the sample to the crystal orientation. The
final stage of the program takes this data and calculates the D-D angles between the <100>, <010> and <001> directions of each phase pair.

To acquire an interphase random pair misorientation angle distribution a second MATLAB program is used. The program creates a random interphase pair by selecting a random pixel of a desired phase and then selects a random pixel of a second desired phase. It continues to do this for a preset number of random pixel pairs. The program then takes the orientation data from both randomly selected pixels (one of each phase) and performs the same processes carried out in the interphase neighbour pair misorientation angle distribution program to determine the misorientation between them.

For each sample an interphase neighbour and interphase random misorientation angle distribution is obtained. These are then displayed as cumulative frequency curves on the same graph for each sample.

5 Results

5.1 Pole Figures

Omphacite pole figures (Figure 4) show an LPO where \{010\} poles form point maxima normal to foliation and the <001> directions are dispersed in a girdle within the foliation plane. (100) poles have a random distribution.

Glaucophane pole figures (Figure 4) also display an LPO. This commonly takes the form of a point maximum in the <001> directions parallel to the lineation. Variation in the <001> directions occurs as weak girdles within the foliation plane (sample S5.2). The (100) poles form a girdle distribution perpendicular to the foliation plane and the (010) poles tend to form point maxima also normal to the foliation. Random distributions are also observed in the (010) poles in sample S5.2 and S6.13.
Barroisite pole figures (figure 4) show alignment of the \{100\} poles normal to the foliation with minor girdles normal to the lineation. \<001\> directions are dominantly aligned with the lineation with subsidiary girdles in the plane of the foliation. \{010\} poles show weak alignment (NB although it is difficult to compare to the other axes of the barroisite because the contours are different) that varies between a foliation-normal point maximum (S6.8) to weak, lineation normal girdles (S5.2).

5.2 P-P, D-D and PD-PD Maps and Interphase Misorientation Angle Distribution Analysis

5.2.1 Omphacite / Barroisite

Histograms of D-D angles for omphacite/barroisite are shown in Figure 5(a-c). Cumulative frequency curves for \<100\> show a striking peak in omphacite/barroisite neighbour pair misorientation at \~30^\circ\. For \<010\> a peak in the neighbour pairs occurs at 180° and for \<001\> a peak occurs at misorientations <5°. Cumulative frequency curves for all samples show the same 30° neighbour pair misorientation peak for \<100\>, an anti-correlation pattern for the neighbour pair misorientations of \<010\> with a sharp increase at 180° and a high frequency of low angle misorientation (<5°) for \<001\>. Figure 5(d) shows a histogram for PD-PD angles, using a \(100\)<010> combination. Note the high frequency of 170-180° misorientations.

5.2.2 Glaucophane / Barroisite

D-D, P-P, PD-PD maps and interphase misorientation angle distribution histograms for all samples, show misorientation across the majority of glaucophane/barroisite phase boundaries is <10°. Details are not illustrated here since the pattern is simple – barroisite grows epitaxially on glaucophane, as their unit cells are of similar shape. In a few areas where a glaucophane grain is bordered by more than one grain of barroisite the interphase misorientation between the glaucophane grain and some of the barroisite grains is >10°.
6 Discussion

The crystallography of omphacite, glaucophane and barroisite already provide us with an intuitive idea of whether they may be mimetic on each other (Deer et al., 1982). All three minerals are monoclinic with very similar crystal lattice parameters. An important difference is that omphacite has a shorter 010 length than glaucophane and barroisite. Also, while all three are monoclinic, omphacite is a clinopyroxene (single chain) and glaucophane and barroisite are amphiboles (double chain). This would suggest that an amphibole would find it difficult to directly mimic the crystal lattice of a clinopyroxene. Between amphiboles the idea of mimicry seems much more plausible.

In the eclogites of Punta Telcio omphacite has a strong LPO and when they are retrogressed barroisite also displays a strong LPO. The two minerals are texturally not in equilibrium with each other in the Zermatt-Saas Unit and thus developed their LPOs during separate events in its evolution. Omphacite LPO will have formed during the prograde or early retrograde portion of the metamorphic history of the unit. Barroisite, appearing in the mineral assemblage during later exhumation of the unit, will have formed its LPO then, through a variety of means which we attempt to identify here.

6.1 Mimicry of Omphacite

6.1.1 Omphacite / Barroisite Orientation Relationships

If barroisite LPOs were not influenced in any way by omphacite then the neighbour-pair and random-pair graphs would coincide, but they are however, statistically different. The high frequency misorientation signals can all be explained by a ‘special relationship’ between the crystallography of omphacite and adjacent barroisite (Figure 6). The relationship can be expressed with notation used in metallurgy, as:
Where barroisite has formed around omphacite due to retrogression it has preferentially
(though possibly not exclusively) grown in that orientation relationship. We will refer to
boundaries with the particular orientation relationship as “special” boundaries.

Large barroisite grains are often surrounded by smaller omphacite grains in the partially
retrogressed eclogites of Punta Telcio, appearing to have grown to engulf several omphacite
grains. This is, to an extent, confirmed by the lack of special boundaries, except along
particular segments of the barroisite grain (Figure 7a). Once barroisite was nucleated with an
orientation relationship relative to a particular omphacite, it could then grow at the expense of
other omphacite grains regardless of their orientation, engulfing and destroying them.

Notably, though, separate segments of the barroisite boundary display the special
relationship. As the omphacite has a strong LPO, occasionally other omphacite grains with a
similar orientation to the parent omphacite grain the barroisite nucleated on are encountered
so that a ‘special relationship’ misorientation boundary is achieved again (Figure 7b).

6.1.2 Omphacite and Barroisite LPOs

An obvious explanation for the omphacite LPO is crystal plasticity, with some dominant slip
systems causing lattice alignment (Bascou et al., 2001; Brenker, 1998; Brenker et al., 2002;
Buatier et al., 1991; Philippot and Van Roermund, 1992) but there have also been suggestions
that omphacite LPO can form during diffusion creep (Godard, 1988; Godard and Van
Roermund, 1995; Helmstaedt et al., 1972; Mauler et al., 2001). Detailed discussion of the
deformation mechanism(s) is outside the scope of this contribution and does not influence our
conclusions here: all that matters is that the omphacite LPO formed during eclogite facies
deformation, and provides an initial microstructural “template”.

In the special relationship, the <001> axes and (001) planes of barroisite are inherited from
those of omphacite, and this is sufficient to explain the broad similarity between those
components for the omphacite and barroisite LPOs. Our data show that the barroisite LPO is
mimetic on that of omphacite and deformation is not required to produce the barroisite LPO.
This explains an otherwise puzzling observation: there is commonly not enough barroisite in
the partially retrogressed rocks to form a connected network, indeed the barroisite can appear
as isolated grains as discussed above. These disconnected grains could have deformed
plastically only if the omphacite matrix were deforming – but that would have caused the
complete breakdown of the omphacite since it was outside its stability field. Static mimetic
growth resolves this paradox.

There are two remaining problems with this model. First, barroisite grains in these rocks
contain low angle (<10°) subgrain walls as shown by maps and intraphase misorientation
angle distribution analyses from all samples. This would suggest that some dislocation creep
and recovery has affected the barroisite, which could point towards plasticity as a cause of
LPO. However, very little strain is required to induce subgrain walls – they may postdate an
already established LPO, and not be of major significance. Alternatively, or in addition, if the
barroisite mimics the omphacite, it could even inherit existing dislocation substructures
known to be present (McNamara, 2009) in that phase. Piazolo et al., (2006) have shown that
subgrain walls can be inherited during recrystallisation in rock salt, so there is a precedent for
this proposal.
The second problem with the hypothesis of inherited LPO is that pole figures for omphacite and barroisite show some important differences. Rather than the \( <001 > \) girdle and \( (010) \) point maxima shown by omphacite, the barroisite \( (010) \) poles can often develop more of a girdle distribution normal to the foliation (S6.13) and a stronger point maxima signature parallel to the lineation in the \( <001 > \) poles. In addition all samples show that omphacite has a random pattern in the \( (100) \) poles whereas barroisite \( (100) \) poles tend to form a girdle distribution perpendicular to the foliation. So while some samples show barroisite and omphacite to have similar pole figures other samples do not, meaning that omphacite may be having some mimetic control but this is not the complete story: the next section provides further insight.

6.2 Mimicry of Glaucophane

6.2.1 Glaucophane / Barroisite Orientation Relationship

Where seen to be touching, the relationship is commonly simple and unambiguous: all lattice vectors are parallel.

6.2.2 Glaucophane and Barroisite LPOs

Unlike barroisite, glaucophane can be stable at eclogite facies. From field and thin section observations it is known that glaucophane growth occurred either during the eclogite facies conditions or in a separate retrogressive blueschist facies event, before the greenschist / amphibolite facies event that resulted in barroisite formation. Large grains of glaucophane grew and formed an LPO here. The glaucophane in some places forms an interconnected network but there are isolated grains still aligned as part of the shape fabric. It is possible that the glaucophane was deformed at the same time as the omphacite and formed its LPO via crystal plasticity: the lack of a connected network is not a problem. Evidence for a later period of static glaucophane growth comes from the large glaucophane nematoblasts that cut the
fabric of the rock obliquely. Either these grains nucleated and grew after those that did so
during deformation or they represent preferential growth of glaucophane in a certain
orientation, or they nucleated and then grew in a preferential direction.

Optical petrography shows barroisite around the edges of glaucophane crystals with very
straight glaucophane/barroisite contacts. This suggests epitaxy on a faceted glaucophane,
with the barroisite as overgrowths. However other textures show that glaucophane is being
topotactically replaced as well as being overgrown by barroisite. There is the possibility then,
that barroisite LPO is mimetic on that of glaucophane, as well as on that of omphacite.
Topotaxy might lead to the complete disappearance of glaucophane, with only a cryptic
signature in the barroisite LPO as its legacy. When glaucophane has been indexed, its LPO is
quite similar to that of barroisite. Glaucophane and barroisite both display point maxima or
weak girdles normal to the foliation in the (100) poles, point maxima or weak girdles normal
to the foliation in the (010) poles and point maxima parallel to the lineation or weak girdles
within the foliation plane in the <001> poles. Table 2 shows the modal abundances of the key
minerals in the studied samples – glaucophane is not always present, but S6.18 (for example)
may have been retrogressed so much that all glaucophane has been replaced.

A problem with the hypothesis of complete glaucophane replacement in some areas of Punta
Telcio is the size discrepancy between large glaucophane and small barroisite grains.
Complete replacement of large glaucophane grains (Figure 2) would logically result in large
barroisite grains. While large barroisite grains are noted there are none with the size and
texture displayed by glaucophane. A possible explanation for this lies with retrogressive
textural relationships noted in these rocks. A second phase of green amphibole growth is
noted and texturally it can be seen to replace barroisite, which was already replacing
glaucophane. Chemical analyses show these green amphiboles to be representative of lower
grade mineralogy such as magnesio-hornblende, actinolitic hornblende and actinolite,
suggesting they formed later in the exhumation of this unit. Grain sizes of this second stage
amphibole growth are smaller than those of barroisite. Replacement of barroisite by this
second generation amphibole may explain why large barroisite grains (that have completely
replaced large glaucophane porphyroblasts) are not observed.

6.3 Microstructural Model

Our model for microstructural evolution begins with an eclogite facies tectonite, in which
omphacite has a strong LPO and shape fabric. Glaucophane may have been present and
deformed together with the omphacite, and/or it grew slightly later as crosscutting
porphyroblasts. During exhumation, barroisite formed – this is likely to have required an
influx of water, since it formed in part from anhydrous omphacite. Barroisite nucleated on
omphacite such that the b and c axes are parallel in the two phases and the a axes are ca. 30°
apart. Barroisite then grew, as isolated grains or connected networks, often with individual
barroisite grains replacing several omphacite grains. The oriented nucleation of barroisite
meant that its LPO was strongly influenced by the omphacite LPO. Barroisite grew in
elongate shapes because amphiboles have anisotropic growth rates; with fast growth parallel
to c. Barroisite also grew over or replaced glaucophane, inheriting the LPO of that mineral
too, so the final barroisite LPO is a composite inheritance. Subgrain walls in barroisite could
have been induced by minor deformation during greenschist facies but this is unlikely given
the low temperatures (Aspiroz et al., 2007). It is more likely they were inherited, c.f. subgrain
walls during recrystallisation of halite (Bestmann et al., 2005) were also inherited. The key
point is that no major deformation at greenschist facies is required to explain the barroisite
LPO and elongate grain shapes.

6.4 Regional Implications
Punta Telcio lies on the upper limb of a recumbent fold (Reddy et al., 1999) as shown in Fig. 445; this accords with a southwards increase in foliation dip in the study area (Fig. 1c). The fold closes south with a gently plunging hinge on 05°/274°. Omphacite and glaucophane mineral lineations related to late prograde/early retrograde deformation run roughly parallel to the fold hinge. Amphibole mineral lineations formed at greenschist facies also run roughly parallel to the fold hinge, and occasional axial-planar fabrics with barroisite are recorded.

Reddy et al., (1999) interpreted the tight Telcio fold as forming at greenschist facies. The amphibole lineations were formed by deformation, and the parallelism of omphacite and greenschist facies lineations were thought of as a coincidence. These barroisite lineations, though, are the expressions of grain shapes. In our revised model, the Telcio fold was established at eclogite facies. It is common in high strain deformation for fold hinges to nucleate parallel to, or rotate into, transport direction (Coward and Potts, 1983; Escher and Watterson, 1974; Sanderson, 1972, Skjernaa, 1980) so the parallelism of omphacite lineations and fold hinge is not surprising. Then, a static greenschist overprint led to the growth of barroisite. Omphacite LPO influenced barroisite LPO, which influenced barroisite grain elongation, which gave rise to the amphibole lineations visible in the field. There might have been some greenschist facies deformation, and the axial-planar fabrics recorded by Reddy et al., (1999) could relate to further flattening of the early fold structure at greenschist facies. But in our revised model, the tight fold was formed, and most deformation occurred, at eclogite facies, and it is no coincidence that later amphibole lineations parallel those in omphacite.

6.5 Implications for Seismic Anisotropy

It is common in the crust, and likely in the mantle too, that phase transformations overprint minerals with LPOs. We have shown here that an LPO can be inherited, albeit in modified
form, during metamorphism. Some inheritance is inferred in other studies e.g. where igneous
plagioclase has broken down to albite (Jiang et al., 2000), though in that case the LPO has
been modified by later deformation. LPOs are significant because they are commonly used to
infer deformation, but also because they lead to seismic anisotropy. In the mantle, seismic
anisotropy has been explained in terms of LPO so the origin of LPO is important for the
interpretation of large-scale mantle behaviour. For example, consider the subduction of
oceanic lithosphere in which the olivine rich mantle has an LPO due to deformation near the
ridge. As the olivine transforms to wadsleyite and ringwoodite during subduction, is the LPO
inherited? It appears that the high pressure phases nucleate coherently (Kerschhofer et al.,
1998) and orientation relationships are to be expected. This may mean that the new phases
grow with an LPO and hence the slab will be seismically anisotropic. It is not trivial to predict
whether mimetic growth will occur, based only on knowledge of the old and new crystal
structures. We have documented a monoclinic \(\rightarrow\) monoclinic relationship here, but sillimanite
growing on biotite is monoclinic \(\rightarrow\) orthorhombic (Yardley, 1977) and Ti phase
transformation is hexagonal \(\rightarrow\) cubic (Seward et al., 2004). Mimetic growth, with
consequent inheritance (in modified) form of LPO and seismic anisotropy, is a significant
process in the Earth.

7 Conclusions

We have shown that orientation data can be successfully used to identify the existence of
mimetic LPO formation. D-D and P-P maps, where the angle between a direction or a plane
of two different phases is calculated, and PD-PD maps, where the misorientation between two
phases is calculated for a plane and a direction within that plane, as well as interphase
misorientation angle distribution analysis are excellent tools to investigate this phenomenon.
Applying these techniques to an investigation of retrogressive barroisite in the eclogite facies rocks of the Zermatt-Saas has resulted in the discovery of a ‘special orientation relationship’ of barroisite on omphacite such that it preferentially replaces the mineral through a $180^\circ$ rotation around the $<001>$ axis of omphacite. The same techniques show that this retrogressive barroisite also shows a direct crystal orientation replacement of pre-existing glaucophane, copying the crystal lattice exactly. Therefore the resulting barroisite LPO is mainly formed in these rocks by mimicry of both a pre-existing omphacite and glaucophane LPO, combining elements of each with the possibility that some crystal plasticity was involved at some stage. Barroisite $<001>$ and $(010)$ pole figures contain elements found in omphacite and glaucophane LPO patterns and barroisite $(100)$ pole figures contain patterns common to glaucophane LPO but not omphacite LPO.

Assuming that the barroisite LPO develops via mimicry of a pre-existing omphacite and glaucophane LPO, leads to the suggestion that greenschist retrogression and exhumation of the Zermatt-Saas Unit in this area was at least, for a time, static. Possible implications for the structural interpretation of a kilometre scale fold in this area show that mimetic fabrics must be carefully considered in any microstructural interpretation of a polymetamorphic rock when attempting to unravel its metamorphic and structural evolution.

Mimetic fabrics may form during other metamorphic reactions. They must be borne in mind when interpreting direct structural observations and seismic anisotropy data in terms of deformation, in both crust and mantle.

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References


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**Figure 1.** Location maps. A) Main tectonic units of the Western Alps. B) Upper Val Gressoney map and cross section, modified from Reddy et al. (1999), location marked on A). C) Representative map of part of Punta Telcio showing relationships of eclogite and retrogressed products, location marked on B). Dot marks sample 6.8 location.
**Figure 2.** A) Plane polarised light (PPL) photomicrograph showing the texture of a partially retrogressed eclogite from Punta Telcio. Three of the main stages of the metamorphic evolution are displayed here: eclogite facies (omphacite and garnet), blueschist facies (glaucophane) and greenschist facies (barroisite). B) EBSD map of a portion of the image in A) with the main mineral phases identified.
Figure 3. Diagrams showing how the EBSD mis-indexing problem between glaucophane and barroisite is solved. A) Plane polarised light photomicrograph showing a glaucophane porphyroblast (light grey) rimmed with barroisite (darker grey). B) EDX calcium map of the same grain. Dashed white line shows the boundary between the glaucophane grain and its barroisite rim. C) Pre-corrected EBSD solution of the same grain with significant mis-indexing in the barroisite rim (dashed white circle). D) Post-corrected EBSD solution with barroisite rim and glaucophane grain points correctly indexed based on calcium chemical counts.
Figure 4. Omphacite (Om), glaucophane (Gln) and barroisite (Bar) pole figures showing {100}, {010} and <001>. a) Sample S6.8. b) Sample S6.13. c) Sample S5.2. Foliation trace horizontal, lineation E-W in orientation displayed.
Figure 5. A-C) Histograms and cumulative histograms showing interphase D-D analysis for

<100>, <010> and <001> between omphacite and barroisite. Note that because of crystal

symmetry the <100> and <001> misorientations run to 90° whilst those for <010> run to

180°. D) Histogram and cumulative histogram showing complete interphase misorientation

(PD-PD) for (100)<010> between omphacite and barroisite.
Figure 6. Diagrammatical representation of the 'special orientation relationship' that barroisite employs to replace omphacite and the 180° rotation axis about <001>. 

θ₁ = 180 - β = 73.41
θ₂ = 180 - β = 74.8
180 - 73.41 - 74.8 = 31.79
Figure 7. A) <001> D-D EBSD map from sample 5.2 showing a large barroisite grain bordered by many omphacite grains, two of which are similarly oriented (arrows) – less than five degrees of interphase misorientation. B) Schematic diagram of how such a microstructure may form.