C-SLIP IN QUARTZ FROM SUBSOLIDUS DEFORMED GRANITE

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ABSTRACT


The slip behaviour of quartz from a granite and granitic veins during subsolidus high temperature conditions (700°–800°C) of deformation is studied by lattice preferred orientation (LPO) and microstructural techniques. LPOs with a point maximum of [C] axes parallel to the shear direction, which is close to the stretching lineation, are presented and interpreted with the aid of light and electron microscopy in terms of [C] as the dominant slip direction during plastic deformation. Microscopy of such samples reveals the presence of (0001) tilt boundaries and free [0001] Burgers vector dislocations in edge orientations. A change in LPO in a decreasing temperature environment from [C] axes parallel to the shear direction to [C] axes normal to the shear direction documents a fabric transition from the rare [C] slip to the more common [a] slip fabric. It is concluded that conditions of high temperature and partial pressure of water are required to initiate dominant [C] slip during deformation at geological strain rates.

INTRODUCTION

As emphasized by Bouchez et al. (1984) reviewing the quartz lattice preferred orientations (LPO) in natural tectonites, [C] axes [0001] in quartz usually dispose at a high angle to the stretching lineation (X direction of the finite strain), suggesting intracrystalline slip in quartz along a lattice direction contained in the basal plane. In fact, experimental and TEM evidence show that the \( \langle a \rangle = \langle 1120 \rangle \) axes are easy directions for intracrystalline slip in quartz (Christie et al., 1964; Griggs, 1967; McLaren and Retchford, 1969; Baeta and Ashbee, 1969). LPO diagrams also argue for \( \langle a \rangle \) being the dominant slip direction in natural tectonites (Bouchez, 1978; Schmid et al., 1981) the associated slip planes being in a zone around \( \langle a \rangle \), with dominant (0001) at low temperature \( (T) \), and \( \{10\overline{1}0\} \) and probably \( \{10\overline{1}1\} \) at higher \( T \) (Wilson, 1975; Bouchez and Pêcher, 1981).

Kirby and McCormick (1979) have demonstrated the dominant activity of the \{11\overline{1}20\} [C] slip system during creep experiments of wet synthetic quartz single
crystals. If also dominant in quartz from natural tectonites, this system with a \([C]\) slip direction should reveal itself by LPO diagrams having \([C]\) mainly disposed close to the stretching direction \((X)\). Such a peculiar pattern has been described by Schmid et al. (1981) in the case of a quartzite from Galicia (Spain) deformed under upper amphibolite facies \(T\)-conditions; however, the latter \([C]\)-axis pattern transpires to be a common \(Y\)-maximum fabric if the finite strain frame is redefined using, as strain markers, the reoriented and stretched needles included in the quartz grains (Bouchez et al., 1984).

Composite \([C]\)-axis patterns made both of the classical single girdle at a high angle to \(X\) and of a maximum close to \(X\), have been found in the quartz-rich migmatites overthrust by the Ronda peridotites (Spain; Bouchez et al., 1985; Tubia and Cuevas, 1986). In these migmatites it has been shown that the quartz grains having their \([C]\) axes close to \(X\) frequently show optically visible subgrain boundaries (SGBs) parallel, or nearly so, to the \((0001)\) basal plane; in contrast, grains having \([C]\) along the girdle do not show any basal SGB. This strongly suggests the activity of \([C]\)-slip, at least for one category of grains, and TEM evidence was found in the form of \((0001)\) tilt walls.

The present work bears on the solid state deformation of quartz within migmatites and granites deformed at temperatures close to the solidus. Typical patterns with \([C]\) close to \(X\) finite are obtained. It is argued that this fabric is the result of dominant \([C]\) slip. The transition to the classical \([C]\)-axis patterns due to \(\langle a\rangle\) slip is observed towards lower \(T\) during progressive deformation.

**GEOLOGICAL SETTING**

In the central part of the Hercynian Vosges massif near Gérardmer (France), an Axial Unit (Hameurt, 1967) mainly made of granitic rocks (not studied here), is separated from a Western Unit by a moderately dipping \((0^\circ-30^\circ)\) granitic thrust sole which is hectometric in thickness (Fig. 1).

The Western Unit is composed of migmatites related to a low-pressure metamorphism, locally intruded by cogenetic granitic bodies of various sizes (10 m to 1 km); at their base the migmatites become truly granitic in composition and structure, with various grain-sizes (fine to porphyritic), constituting the above defined granitic thrust sole. The whole Western Unit appears as a thrust pile with a ubiquitous flat foliation and an associated NE–SW trending stretching lineation; it grades downwards into the granitic sole, highly deformed according to the same kinematic axes, showing in particular numerous microshear planes or \(C\)-planes of Berthé et al. (1979). The various microstructural criteria applied to the rocks of the pile, including those which are valid for magmatic structures (Blumenfeld, 1983), constantly indicate a southwestern shear direction.

Numerous late magmatic to early post-magmatic veins made of fine-grained granite, centimetric to decimetric in thickness, intrude the porphyritic granites of the
granitic sole. These veins often display a clearly defined internal structure (foliation and lineation) attributed to post-injection relative movements of the vein-walls.

The deformed samples which are studied here for their quartz LPOs were collected from the two following environments: (1) in the granites and locally migmatites of the thrust sole, and (2) in the sheared granitic veins, mainly in the Tronces Quarries (star in Fig. 1a).

PROGRESSIVE DEFORMATION FROM MAGMATIC TO SOLID STATE

The migmatites and granites which we are concerned with here were deforming during the transition from the magmatic state to the solid state. This is demonstrated in the granitic sole (Fig. 2) as well as in the veins by the progressive transition from microstructures due to magmatic flow (Class A) to microstructures due to solid state deformation (Class C) having similarly oriented stain-induced planar and linear shape fabrics.

The microstructures

In the magmatic state corresponding to the Class A microstructure, the foliation (XY plane) and lineation (X direction) are marked by the shape preferred orienta-
Fig. 2. Microstructural zonation (A, B, C) and mineral stretching lineation pattern of the studied area.  
Class A: Magmatic structure; undeformed quartz aggregates.  
Class B: Incipient solid state deformation of quartz in ellipsoidal quartz-aggregates.  
Class C: Typical elongate to ribbon-shaped quartz aggregates with developing micro-shear planes  
(incipient orthogneissification).  
Arrows: subhorizontal lineations; tips pointing in the direction of thrusting.

tion (SPO) of the micas, plagioclases and K-feldspars (see Bouchez et al., 1981); large quartz grains, assembled in subspherical (undeformed) aggregates, do not show any clearly defined lattice preferred orientation (LPO). At this stage the granitic material is assumed to have been entirely deformed as a crystal-melt system.

The onset of solid state deformation helps to define the Class B microstructure, marked by quartz-aggregates becoming ellipsoidal in shape; quartz LPO may be well-defined, but no micro-shear planes are observed yet. At this stage the material may have experienced some deformation still as a crystal-melt system, but with a minor amount of intergranular melt present. The crystals forming a rigid frame (Van der Molen and Paterson, 1979) have to deform plastically. Quartz, being the most ductile mineral of the paragenesis, readily deforms.

The appearance of micro-shear planes helps to define the Class C microstructure, indicating without ambiguity that the granite experienced a certain amount of strain entirely in the solid state. The micro-shear planes, which may become closely spaced, and their associated mineral aggregate lineation are used to define respectively the bulk shear plane ($P_3$) and the shear direction ($I_3$; Fig. 3a). The latter kinematic framework is slightly oblique to the $XYZ$ finite strain frame by an amount $\theta$ around the $Y$ axis which depends on the strain magnitude. The quartz-aggregates flattened parallel to $XY$ and elongate parallel to $X$ usually have a pronounced
Fig. 3. Finite strain (X, Y, Z) and kinematic (Ps, LS) frames.

a. Micro-shear planes Ps (Class C microstructure) in a vein and associated aggregate lineation LS at a dihedral angle θ to the XY plane of the finite strain defined by the shape-preferred orientation of the crystals (specimen FlO; see Fig. 5d). Note that Ps always appears parallel to the vein walls (WALL); hence when micro-shear planes are absent (Class A and B microstructures) θ is taken as the dihedral angle between XY and WALL.

b. Lower hemisphere stereogram of the orientation of needle shaped inclusion long axes within quartz grains. Maximum density (6.5%) is close to X, confirming the orientation of the finite strain X-axis; 60 measurements (N) from specimen F5.

this stage the amount of plastic strain in quartz is sufficient to reorient elongate inclusions, mainly rutile needles, parallel to X (Fig. 3b), thus confirming X to be the long axis of the finite strain ellipsoid. With increasing strain, quartz may become ribbon-shaped.

In the thrust pile the progressive transition from A to C relates directly to the distance from the basal contact of the pile and corresponds to a unique kinematic episode: whatever the microstructural class, the foliation–lineation framework (XYZ) stays constant in orientation and the sense of shear is always directed towards the southwest (Fig. 2). In the veins the transition from A to C is not observed within a single vein but within a collection of neighbouring veins. In this case, a polyphased deformation cannot be ruled out but a sub-solidus temperature can still be deduced during the activity of the micro-shear planes (Class C), as will be shown now.
Fig. 4. Pressure–temperature diagram indicating the stability field (shaded) of the various phases present in the micro-shear planes of the granitic veins (Class C microstructures). Water pressure is equated with total pressure. This field is centered on 675°C, 3.5 kbar (350 MPa). Note that the field is cross-cut by the H₂O saturated granite solidus (from Wyllie et al., 1976). K—from Kerrich (1972); H—from Holdaway (1971); HL—from Holdaway and Lee (1977).

**T,P estimate during strain in the veins (Class C)**

The stable mineral association which is present in the micro-shear planes of the veins is: K-feldspar, frequently myrmekitic plagioclase, magnetite, sheaf-like sillimanite and recrystallized biotite. Microprobe analyses of the biotites (Camebax; University of Nancy I) give (Fe/Fe + Mg) ratios between 0.53 and 0.67. Taking the biotite equilibrium data in the latter paragenesis (Wones and Eugster, 1965) a minimum temperature of 700°C can be deduced. However, the aluminous character of the biotite may slightly decrease the equilibrium temperature (Rutherford, 1973).

Presence of stable and unaltered cordierite in some veins, having (Fe/Fe + Mg) ratios around 0.48, associated with biotite, sillimanite and K-feldspars, helps to define the univariant equilibrium curve in the $P,T$ space shown in Fig. 4 (Holdaway and Lee, 1977).

The $P,T$ stability field which best accounts for the latter data is centered around $T = 650°-700°C$ and $P = 350$ MPa (shaded in Fig. 4), clearly located near the hydrated granite solidus.

**QUARTZ LATTICE PREFERRED ORIENTATION**

The LPOs in quartz have been studied (1) with the conventional U-stage for the [C]-axes, and (2) with the X-Ray texture goniometer of F. Lhote and J.P. Uriot (Lhote et al., 1969) for the $\langle m \rangle = \langle 10 \overline{1} 0 \rangle$ and $\langle a \rangle = \langle 11 2 0 \rangle$ axes.
LPOs in the veins (Fig. 5)

In the veins displaying microstructures of Class A (undeformed quartz grains) or Class B (incipient quartz deformation), the [C]-axes diagrams are ill-defined, having

Fig. 5. LPO of [C] and ⟨m⟩ axes from the granitic veins. Class A, B or C signifies the microstructural type (see text). X, Y and Z are the finite strain axes. LS is the shear direction. For the [C]-axis fabrics (a–g) the labels are, from the top: specimen number, number of measurements, maximum density, θ (see Fig. 3a), and shear strain γ (γ = 2/tan 2θ); contours: 0.5, 1.5, 2.5, 3.5 and 5.0%. For the ⟨m⟩-axes fabrics (f–h) the contour values are in multiples of uniform density with maximum and minimum recorded densities.
numerous and dispersed submaxima (Fig. 5a, b and c). Within the Class C microstructures (presence of micro-shear planes and moderate to large finite strain in quartz) the [C] axes clearly concentrate into a maximum located at a high angle to both Y and Z finite axes. With increasing strain (γ), as determined by the decreasing θ values (see Ramsay, 1980), the [C] maxima become more pronounced and locate preferentially around the Ls shear direction (Fig. 5d, e and g).

The ⟨m⟩ axes determined for γ = 2.0 (Fig. 5f) tend to concentrate perpendicularly to Ls with a high concentration around Y; for γ = 3.5 (Fig. 5h) the ⟨m⟩ axes give a clear girdle normal to Ls with a distinct submaximum close to Z.

LPOs in the thrust pile (Fig. 6)

The [C]-axis distributions corresponding to Class A microstructures in the migmatites and granites are random or nearly so and not presented here. In contrast, pronounced LPOs are observed as soon as ellipsoidal quartz aggregates appear (Class B microstructures). Two very different types of LPO patterns can be defined:

1) The first type (Fig. 6a and b), similar to that of the deformed veins (Fig. 5e–h), is exclusively encountered among the specimens with a Class B microstructure, usually slightly deformed, but specimens with low θ values (highly strained) are also encountered. This type is illustrated by a coarse-grained porphyritic granite with [C] axes clustering close, but at an angle with X-finite (Fig. 6a) and ⟨m⟩ axes displaying a girdle distribution at a high angle to X (Fig. 6b).

2) The second type of LPO patterns (Fig. 6e and f) comes from granites belonging to the sole of the pile, highly deformed in the solid state. They usually display micro-shear planes (Class C microstructures). This type is illustrated by a granite with a calculated γ of 3.5, with [C] axes strongly clustering around the Y-finite axis and a tendency for a girdle distribution normal to Ls; the ⟨a⟩ axes mainly distribute along the XZ plane, one of the maxima being clearly parallel to Ls.

Somewhat intermediate LPO patterns are found in “moderately” deformed specimens (Fig. 6c and d; γ = 2.0): the [C] axes tend to distribute at a high angle to Ls, as in the latter type, but a “relict” submaximum remains close to Ls; the marked tendency of the ⟨m⟩ and ⟨a⟩ axes to dispose at a high angle to Y and rather close to X or Ls correlates well with the Y-maximum tendency of the [C] axes.

[C] ‖ Ls and ⟨a⟩ ‖ Ls fabrics

Two radically different types of fabric result from the latter LPO data: (1) the [C] ‖ Ls fabric, with the [C] axes mainly clustering around the Ls shear direction (Fig. 4d, e and g; Fig 5a), and (2) the ⟨a⟩ ‖ Ls fabric with the ⟨a⟩ axes clustering
around Ls (Fig. 6d and f). With increasing solid state deformation in quartz, the example from the deformed veins illustrates the transition between the nearly random to the \([C] \parallel L_s\) fabric; the thrust-pile example illustrates the transition between the \([C] \parallel L_s\) to \(\langle a \rangle \parallel L_s\) fabrics, through an "intermediate" fabric.
Fig. 7. Quartz [C]-axis fabric evolution with increasing density of micro-shear planes (Nsp) towards the base of the granitic thrust pile.

Plot of the percentage ($P_{LS}$) of $[C]$ axes closer than 45° to the shear direction, versus the number of micro-shear planes per cm in $XZ$ sections. High values of $P_{LS}$ signify dominant $[C]$ slip whereas low values reflect $\langle a \rangle$ slip. The value of $N_{SP}$ is considered to be a measure of total strain undergone since the generation of the micro-shear planes, i.e. mainly in the solid state, at decreasing temperatures. A clear transition from $[C]$ ($P_{LS} > 25$) to $\langle a \rangle$ slip is recorded with increasing $N_{SP}$.

In order to better track the evolution between $[C] \parallel L_S$ and $\langle a \rangle \parallel L_S$ fabrics, 14 $[C]$-axis orientation patterns have been determined with the U-stage in a set of variously deformed granites and migmatites of the thrust pile. The resulting data (Fig. 7) are plotted in terms of $P_{LS}$ (percentage of $[C]$ axes closer than 45° to $L_S$) versus $N_{SP}$ (number of micro-shear planes/cm when observed in $XZ$ sections); $N_{SP}$ is considered to be a measure of the solid state strain undergone by the rock, hence by the quartz grains, with decreasing temperatures (sampling towards the base of the pile). In such a plot (Fig. 7) the Class B microstructures, devoid of micro-shear planes, naturally group along the $N_{SP} = 0$ line. The negative trend of the plot argues for the continuous transition from $[C] \parallel L_S$ to $\langle a \rangle \parallel L_S$ fabrics during persistent deformation at decreasing temperatures.

SUBSTRUCTURE IN QUARTZ: OPTICAL AND TEM DATA

As stressed in the introduction, the $[C] \parallel L_S$ fabrics are particularly unusual in natural tectonites whereas $\langle a \rangle \parallel L_S$ fabrics have been described by many authors. Consequently attention has been paid to the optical and TEM substructures in quartz from specimens displaying the latter unusual fabric.
Systematic observations under the optical microscope clearly show that quartz grains within the least deformed specimens of either the granitic pile or the veins display unusual amounts of basal (or near-basal) subgrain boundaries (SGBs), i.e. parallel to (0001). Basal SGBs (Fig. 8A) are rarely exactly planar, even over 100 µm distances, and often form approximately square subgrains with planar prismatic SGBs (Fig. 8B). Basal SGBs rarely transect a whole grain whereas the prismatic ones do, hence appearing as the most striking optical features (Fig. 8C). In the case of specimen F2, the granitic vein whose [C]-axis diagram is presented in Fig. 5a, considered to be almost undeformed in the solid state (Class A), a statistical analysis was undertaken: about 30% of the SGBs are in a basal to sub-basal orientation versus more than 60% in a prismatic one (Fig. 9A).

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Fig. 8. Optical microstructure of quartz grains in granitic veins.
A. Specimen F9 (LPO given in Fig. 5e, f): irregular basal subgrain boundaries (Class C).
B. Specimen F2 (LPO given in Fig. 5a): basal subgrain boundaries forming part of a square subgrain (Class A).
C. Specimen F7 (LPO given in Fig. 5c): straight prismatic subgrain boundaries (Class B). Scale bar: 100 µm.
With increasing strain (Class B) basal SGBs are still observed, forming a few percent of the whole population; in the case of Class C microstructures, basal SGBs are exceptional, the population being almost entirely represented by prismatic SGBs (Fig. 9b).

**TEM analysis**

To further our understanding of the glide systems operating during the high temperature deformation, TEM observations were made of specimen F2. Following optical studies individual grains containing optically visible basal SGBs were selected. Samples were ion-thinned to electron transparency and carbon coated to avoid electron charging during observation. All observations were made using a JEOL 120-CX operating at 120 kV. Analysis of slip systems was carried out using the standard technique of invisibility which had previously been shown to be effective in quartz (Ardell et al., 1974). The technique requires that the vector products \( g \cdot b = 0 \) for a pure screw dislocation, and additionally \( g \cdot b \times u = 0 \) for a pure edge dislocation, for a dislocation to be invisible, where \( g \) is the diffracting vector, \( b \) the Burgers vector and \( u \) the direction along the dislocation line; it is emphasized that these criteria do not apply to mixed dislocations. To simplify the analysis it was assumed that the dominant gliding dislocations had Burgers vectors of the type \([0001]\) or \(\frac{1}{3}[2\overline{1}10]\). This assumption seems reasonable since the other possible \(\frac{1}{3}[2\overline{1}3]\) glide always requires a resolved shear stress approximately three times that of \([0001]\) or \(\frac{1}{3}[2\overline{1}10]\) (Trépied and Doukhan, 1982). Observations were made on free dislocations and subgrain boundaries.

A typical image of free dislocations is shown in Fig. 10A and B, where dislocations with a Burgers vector \(b = \frac{1}{3}[\overline{1}2\overline{1}0]\) are out of contrast for the diffracting vector.
Fig. 10. Bright field electron transmission microscope images from specimen F2. Scale: 1 μm in all cases; 
\( \mathbf{g} \) is the diffracting vector; \( \mathbf{b} \) is the Burgers vector.

A. B. Free dislocations with \( \mathbf{b} = [0001] \) at c, and \( \mathbf{b} = \frac{1}{3}[12\bar{1}0] \) at a and L. Note loops at L elongated in the [0001] direction and the long edge segments of both dislocation types.

C. D. (0001) tilt boundary composed of [0001] dislocations. The dislocation \( \mathbf{b} = \frac{1}{3}[2\bar{1}10] \) at a is in \( \mathbf{g} \cdot \mathbf{b} = 0 \) conditions for both micrographs but is only in weak contrast for \( \mathbf{g} = (\bar{1}1\bar{2}0) \), (see text).

E. (10\bar{1}0) twist boundary composed of [C] and \( \langle a_2 \rangle \) screw dislocations. \( \langle a_2 \rangle \) dislocations are out of contrast for \( \mathbf{g} = (1011) \) whereas [C] are out of contrast for \( \mathbf{g} = (1\bar{2}10) \) (inset). \( h \) denotes bubble.

F. (21\bar{1}0) tilt boundary composed of \( \frac{1}{2}[2\bar{1}10] \) dislocations parallel to [0001].

Vector \( \mathbf{g} = (10\bar{1}1) \) whereas dislocations with \( \mathbf{b} = [0001] \) are out of contrast for \( \mathbf{g} = (1\bar{2}10) \). Trace analysis of the dislocation lines reveals that both dislocations types are gliding on the (10\bar{1}0) plane with long edge and short screw segments.

The (0001) tilt boundaries are common (Fig. 10C and D) with dislocations parallel to the rotation axis [01\bar{1}0]; these boundaries are never perfectly planar:
undulations are often observed at interactions with \( \frac{1}{2}[2\overline{1}10] \) free dislocations (Fig. 10C and D, denoted by \( a \)). Note that although \( g \cdot b = 0 \) (condition for invisibility of a pure screw dislocation) the dislocation at \( a \) in Fig. 10C is in strong contrast; this is due to its mixed character with the dislocation line \( u = [12\overline{1}3] \), giving 6.8 for the magnitude of the vector product \( |g \cdot b \times u| \). In Fig. 10D the vector product is only 1.0 hence the dislocation is almost invisible.

(10\overline{1}0) twist boundaries were observed (Fig. 10E) composed of [0001] and \( \frac{1}{3}[\overline{1}2\overline{1}0] \) screw dislocations. By varying the character from screw to mixed of either Burgers vector these boundaries may have a variable boundary plane (i.e. a “flexible” character) which is either \( \perp [0001] \) or \( \perp [\overline{1}2\overline{1}0] \) (Trépied et al., 1980). (2\overline{1}10) boundaries (Fig. 10F) composed of \( \frac{1}{3}[2\overline{1}10] \) edge dislocations with a dislocation line \( u = [0001] \) are also present. Note again that \( g \cdot b = 0 \) but the edge dislocation is in strong contrast as \( g \cdot b \times u = 5.7 \); additionally we observe a symmetrical along the line contrast (Gandais et al., 1982) characteristic of \( g \cdot b = 0 \) and \( g \cdot u = 0 \).

The TEM study reveals the presence of free [0001] dislocations in an edge orientation, gliding on the (10\overline{1}0) plane. This glide system is consistent with the presence of (0001) tilt boundaries composed of [0001] edge dislocations. \( \frac{1}{3}[2\overline{1}10] \) dislocations are present in mixed or edge orientations, gliding on (10\overline{1}0) planes. Interactions between these systems can produce boundaries of variable orientation explaining the irregular nature of optically visible subgrains in this specimen. Regions dominated by \( \frac{1}{3}[2\overline{1}10] \) glide are also present as indicated by (2\overline{1}10) tilt boundaries.

DISCUSSION

The present study demonstrates the existence of specimens with LPO characterized by a maximum of \([C]\) axes located close to the \( X \) direction. The \( X \) direction is usually defined using the elongation of crystals and mineral aggregates. In the case of specimen F5, with a well-defined point-maximum of \([C]\) axes close to \( X \) (Fig. 5g), \( X \) has been ascertained to be the long axis of the finite strain ellipsoid using the elongate inclusions in the quartz grains as strain markers (Mitra, 1976; Bouchez et al., 1984). These inclusions, which may break and pull-apart, preferentially align parallel to \( X \) as shown in Fig. 3b. It further argues for the quartz grains being deformed by intracrystalline slip.

**Dominant \([C]\) slip direction**

(1) Fabrics. By analogy with other tectonites deforming plastically and characterized by a point-maximum of the crystal slip direction close to \( X \), our \([C]\)-axis diagrams in quartz are readily interpreted as due to the dominant activity of \([C]\) as a slip direction. The best analogy comes from the olivine tectonites, with [100] most usually concentrating around \( X \) (Nicolas and Poirier, 1976), [100] being the easiest
slip direction in a wide range of conditions (ref. cit.). "Common" quartz tectonites also furnish a good analogy as they most frequently have a strong concentration of \( \langle a \rangle \) around \( X \). \( \langle a \rangle \) slip being well documented in quartz (Christie et al., 1964; Doukhan and Trépied, 1985).

Numerical simulations of fabrics, when applying the single slip hypothesis (Bouchez et al., 1983) also demonstrate the alignment of the slip direction toward \( X \) finite (Etchecopar, 1977) provided that minor grain external rotations of the grains are allowed. This condition is easily fulfilled in polycrystalline aggregates (Bouchez and Duval, 1982), for instance by kinking and twisting, with the help of grain-boundary migration for the grain accommodations.

(2) Microstructures. They further attest to the existence of \( [C] \) slip. With the optical microscope, for both categories of specimens (from the thrust pile and from the veins), SGBs which are parallel, or nearly so, to the basal plane, are systematically observed among the least deformed specimens. These SGBs must have a tilt component because pure twist SGBs of basal orientation would be unobservable under the optical microscope, their rotation axis being parallel to \( [C] \). If due to slip, the basal SGBs thus indicate \( [C] \) slip for their formation. It is noted that (1) basal SGBs are always less numerous than prismatic SGBs, and (2) the population of basal SGBs decreases dramatically with increasing strain. This is now discussed along with the TEM data.

With the TEM, the existence of basal tilt walls is confirmed; they contain edge dislocations with \( b = [0001] \) indicative of \( [C] \) slip. It is also noted that dislocations with \( b = \frac{1}{2}[21\bar{1}0] \) are always present, either free or associated with \( (2\bar{1}\bar{1}0) \) tilt walls. This implies the complementary activity of \( \langle a \rangle \) as a slip direction, consistent with the presence of optically visible prismatic SGBs. An important contribution of \( \langle a \rangle \) slip is conceivable in the case of specimen F2, for which, because it is slightly deformed, a clear \( [C] \)-axis pattern is not yet defined (Fig. 5a). More puzzling is the ubiquity of the prismatic SGBs, even in strongly deformed specimens displaying a clear \( [C] \)-maximum around \( X \) (vein F5: Fig. 5g and 9b). This may be due to a final deformation at lower temperatures, with \( b = \frac{1}{2}[21\bar{1}0] \), and for which total strain is low (say less than 30%) hence not reflected in the LPO. An alternative explanation is suggested by the faster glide rate of \( [C] \) dislocations compared to \( \langle a \rangle \) ones (Linker et al., 1984) which would result in most \( [C] \) reaching a sink (e.g., grain boundary) or annihilating, hence not forming low-energy substructures such as tilt walls.

(3) \( [C] \| L_s \) fabrics. In the case of non-coaxial deformation (say simple shear for simplification) the single slip direction interpretation is compatible with the alignment of \( [C] \), not exactly with \( X \), but with the principal kinematic axis \( L_s \), parallel to the shear direction in case of simple shear. The difference in orientation of \( X \) with respect to \( L_s \) is the basis of kinematic analysis using fabrics in many tectonites: it helps to establish the non-coaxial character of the strain and points to the direction of shear (Bouchez et al., 1983). In our specimens, and for the microstructural classes B and C, the kinematic frame is defined using the micro-shear planes (\( P_s \)) and lines
\( L_s = \) mineral stretching lineation on \( P_s \). In case of the veins, \( P_s \) is also parallel to the walls, and \( L_s \) is also parallel to the projection of \( X \) on the wall planes. Hence the latter frame is also used for specimens without micro-shear planes (Class A microstructures, Fig. 3a). Therefore, the observed alignment of our \([C]\)-maxima with the independently defined \( L_s \) shear direction (Figs. 6a and 7d, e and f) reinforces the single slip direction interpretation and its predictive potential for the shear directions in case of dominant \([C]\) slip. In the case of the granitic thrust pile (Fig. 6a) the deduced shear direction conforms with the regional data.

**Conditions for \([C]\) slip**

The magmatic character of the Class A microstructures, with only incipient solid state deformation features in quartz, and the progressive transition from Class A to Class C microstructures, point to deformations undergone close to the solidus temperature of the granitic material. More precisely, micro-shearing in the veins has been shown to be in equilibrium with mineral assemblages related to the 650°–700°C temperature domain (Fig. 4). This is in agreement with the observed Dauphiné twins under TEM, suggesting deformation in the \( \beta \) field. This temperature domain is similar to that which prevailed during deformation of migmatites at the base of the Ojen peridotite nappe (Spain; Tubia and Cuevas, 1986) where \([C]\) slip has been recently documented (Bouchet et al., 1985). But in the Ojen migmatites, \([C]\) is not the dominant slip direction; it is combined with \( \langle a \rangle \) slip, giving composite \([C]\)-axis patterns composed of a single girdle due to \( \langle a \rangle \) slip and a \([C]\)-maximum close to \( X \) due to \([C]\) slip.

Therefore dominant \([C]\) slip, as it is reported here, could be related not only to the high temperature environment, but also to the possible high hydrous fluid pressure (here 350 MPa) which usually accompanies granite crystallization, and which seems to be critical for quartz plasticity (Griggs, 1967). In fact, the presence of large volumes of water-rich fluids is indicated by the frequent pegmatitic borders of the veins, stable hydrous phases (e.g. micas), fluid inclusions and numerous healed microfractures as observed in TEM.

**Transition to \( \langle a \rangle \) slip**

The \( \langle a \rangle \) direction of slip, though evidenced by TEM and by optically prismatic SGBs, is considered to make only a minor contribution to total strain in quartz at high temperature (and fluid pressure?) when clear \([C]\)|| \( L_s \) fabrics are recorded. Its contribution becomes predominant, as in most natural quartz tectonites, when \( \langle a \rangle || L_s \) fabrics are observed. Their enhancement at lower temperatures is indirectly measured by the hardening of the rock supposed to be responsible for the increasing number of micro-shear planes (Fig. 7). In the present study, dominant \( \langle a \rangle \) slip gives rise to typical patterns with \([C]\) axes concentrated around the \( Y \) axis of finite strain.
(Fig. 6e) indicative of prismatic (10\bar{1}0) (Wilson, 1975) and/or rhomb (10\bar{1}1) slip planes (Bouchez and Pécher, 1981).

CONCLUSION

We propose that the combination of a very high temperature (700°–800°C) and the presence of a high partial pressure of water are required to invoke dominant [C] slip in quartz at geological strain rates. Our TEM observations indicate that the associated slip plane is (10\bar{1}0), rather than the (11\bar{2}0) reported in experiments on oriented synthetic quartz single crystals (Kirby and McCormick, 1979). Similarly we observe [C] dislocations in the edge orientation, whereas other experiments have reported a dominant screw orientation (Linker et al., 1984). These differences may be explained by the fact that, in the experiments, single crystals were oriented such that the critical resolved shear stress was a maximum on the (2\bar{1}10) slip plane whereas, in polycrystalline samples, any prismatic slip plane can be activated. In addition, the presence of edge rather than screw dislocations suggests that the diffusive anisotropy (Dennis, 1984; Giletti and Yund, 1984), evident in experiments (Linker et al., 1984) is not so important under equilibrium conditions.

The conditions for predominant [C] slip are, according to the present observations, close to or limited to the β stability field, and speculatively require water-enriched fluids. As shown by our example (Fig. 6), a small increment in finite strain, under unfavorable conditions for [C] slip, may replace [C] slip fabrics by the better known ⟨a⟩ slip ones. Such fabric transitions with falling metamorphic temperature but continuing deformation may explain why [C] slip fabrics are uncommon.

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