

Microfabrics of UHP metamorphic granites in the Dora Maira Massif, western Alps – no evidence of deformation at great depth

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ABSTRACT In the Dora Maira Massif, western Alps, essentially undeformed ultrahigh-pressure (UHP) metamorphic granites (Brossasco granite) are embedded in, and locally grade into, granite gneisses or augengneisses and mylonites. In this study, the quartz microfabrics of the undeformed granites are compared against the augengneisses and mylonites in a representative number of samples from several locations. In the undeformed granites, the fine-grained quartz aggregates that formed from coesite upon decompression are characterized by a foam structure and random crystallographic orientation. In the deformed granites, the quartz microstructures and the crystallographic preferred orientation (CPO) indicate deformation by dislocation creep. Most of the deformation of the granites (if not all) must have happened at a late stage during exhumation, after transformation of coesite to quartz, at greenschist facies conditions in the middle crust. The deformed granites provide no evidence of deformation during subduction, at (U)HP metamorphic conditions, and in the earlier stages of exhumation. The diameter of internally undeformed slices of continental crust subducted to and exhumed from about 100 km can exceed that of the presently exposed Brossasco granite, i.e. it can be on the kilometre scale.

Key words: coesite; deformation; metagranites; microfabrics; quartz; UHP metamorphism.

INTRODUCTION

Many ultrahigh-pressure (UHP) metamorphic rocks identified so far do not show much evidence of significant deformation at UHP conditions (e.g. Biino & Compagnoni, 1992; Michard *et al.*, 1995; Stöckhert & Renner, 1998; Oberhaensli *et al.*, 2002; Stöckhert, 2002; Zhang *et al.*, 2005). The lack of penetrative deformation indicates strain localization into very weak shear zones presumably deforming by dissolution precipitation creep and granular flow (in a broad sense) in the presence of an aqueous fluid (e.g. Stöckhert & Renner, 1998; Stöckhert, 2002). Although expected to be common features in a subduction zone environment, shear zones that unequivocally formed during subduction, UHP metamorphism or early stages of exhumation are rarely described (e.g. Zhao *et al.*, 2003, 2005; Terry & Heidelbach, 2006).

In the Dora Maira Massif, western Alps, essentially undeformed UHP metamorphic granites (Brossasco granite) are surrounded by, and locally grade into, granite gneisses or augengneisses and mylonites (e.g. Chopin *et al.*, 1991; Henry *et al.*, 1993; Compagnoni *et al.*, 1994). The mineral assemblage of the augengneisses and mylonites indicates metamorphic overprint at greenschist facies conditions (Henry *et al.*, 1993; Compagnoni *et al.*, 1994).

In the present study, the quartz microstructures and textures of the metagranites are analysed to explore whether their deformation took place during subduction, at UHP metamorphic conditions or during exhumation. The question is whether the inhomoge-

neous deformation of the granites reflects processes in the deep level of a collisional belt or in a subduction channel (e.g. Gerya & Stöckhert, 2006), as opposed to a late overprint at an intermediate crustal level.

The relic coesite bearing the so-called pyrope-quartzites and adjacent jadeite-kyanite-quartz rocks, both embedded in the granite gneisses and characterized by a coarse-grained foam microstructure of quartz, were the subject of a previous study (Lenze *et al.*, 2005). The results indicate an episode of grain-scale deformation at very high differential stresses of more than 0.3 GPa, prior to annealing during exhumation. The very high stress levels are recorded by mechanical twinning of jadeite. The unsystematic orientation distribution of jadeite crystals with and without mechanical twins, supported by that of kyanite crystals with and without kink bands, led to the conclusion that this deformation cannot be driven by far-field tectonic stresses. Instead, the highly inhomogeneous stress field is attributed to the volumetric strain inherent in the transformation of coesite to quartz. Furthermore, the high stress values indicate that this transformation must have taken place within very short time durations, precluding significant relaxation during the progress of the reaction at temperatures $\sim 700^\circ\text{C}$. Comparisons with experimental results suggest that the time scale of transformation is similar to that of the laboratory experiments. Subsequently, the microstructure of quartz formed from coesite became modified by grain growth during static annealing and cooling from temperatures as high as $\sim 700^\circ\text{C}$. The preservation of this quartz

microstructure rules out any later pervasive deformation of these rocks, and consequently indicates a very low level of differential stress during exhumation.

The lack of early and deep-seated deformation of large bodies of UHP metamorphic rocks has implications for the stress level and deformation mechanisms in subducted continental crust, as well as for the length scale of internally undeformed crustal slices subducted to and exhumed from depths of 100 km.

GEOLOGICAL SETTING

The Dora Maira Massif together with the Monte Rosa and the Gran Paradiso Massifs represent basement nappes of the inner Penninic realm (e.g. Schmid *et al.*, 2005) in the western Alps, and are characterized by high pressure (HP) and – in the case of Dora Maira – UHP metamorphic assemblages. These nappes are proposed to be part of the former European continental margin, subducted beneath the overriding Apulian plate and stacked to become nappes during continent–continent collision during the Tertiary (Debelmas & Lemoine, 1970; Dal Piaz & Lombardo, 1986; Pfiffner, 2005; Schmid *et al.*, 2005). An alternative concept was proposed by Gerya & Stöckhert (2006), where the (U)HP metamorphic continental crust is derived by subduction erosion from the front of the overriding plate, cycled through a subduction channel, and accreted to the deep forearc, prior to continent–continent collision.

In the Dora Maira Massif, slices of continental crust reached a depth of more than 100 km and underwent alpine UHP metamorphism with pressures of about 3.7 GPa at temperatures up to 800°C (e.g. Chopin *et al.*, 1991; Schertl *et al.*, 1991; Michard *et al.*, 1995). The age of the UHP metamorphism is suggested to be about 35 Ma (Gebauer *et al.*, 1997; Rubatto & Hermann, 2001). The UHP unit (e.g. Michard *et al.*, 1995) also termed ‘coesite-bearing terrane’ (Chopin & Schertl, 2000) or ‘coesite-bearing Brossasco Isasca unit’ (Biino & Compagnoni, 1992), is exposed in the southern part of the Dora Maira Massif. This unit is generally divided into two subunits based on the association of different rock types: (i) The ‘Isasca varied formation’ (Chopin *et al.*, 1991) comprises gneiss, marble and eclogite, and has a structural thickness of ~300 m. (ii) The ‘gneiss series’ (Chopin *et al.*, 1991), sometimes referred to as ‘country rock gneiss’ (Chopin & Schertl, 2000), is made up of metagranites, with a protolith age of *c.* 300 Ma (Tilton *et al.*, 1991; Gebauer *et al.*, 1997). The structure of the metagranites ranges from fine-grained orthogneiss over augengneiss to the undeformed (e.g. Biino & Compagnoni, 1992) Brossasco metagranite. The well-known coesite-bearing pyrope-quartzite is found intercalated as boudins in the gneiss series. The structural thickness of the gneiss series is estimated as ~500 m (Chopin *et al.*, 1991).

The mineral assemblage of the metagranites consists mainly of quartz, K-feldspar, plagioclase, biotite,

K-white mica, garnet and epidote/allanite. Accessories are apatite, titanite, tourmaline, zircon and rutile (Schertl *et al.*, 1991). Evidence of UHPM was first observed in pyrope quartzite and eclogite (Chopin, 1984; Chopin *et al.*, 1991). Later systematic research revealed coesite inclusions in xenoliths of the Brossasco metagranite and relict mineral assemblages of garnet, jadeitic clinopyroxene and phengite characteristic of UHP metamorphic conditions (Burchard, 1999) in the country rock gneisses. These findings support earlier assumptions (Chopin *et al.*, 1991; Biino & Compagnoni, 1992) that the entire gneiss series had been subjected to UHP conditions (Chopin & Schertl, 2000).

Field relations and sampling strategy

The results of detailed fieldwork with structural analysis of the HP and UHP metamorphic units in the southern Dora Maira Massif were published by Chopin *et al.* (1991), Henry *et al.* (1993) and Michard *et al.* (1995). These authors described a gently dipping mylonitic foliation with an ENE–WSW stretching lineation and top-to-the-west kinematic indicators, fully consistent with our observations in the field. The general orientation of the foliation and stretching lineation is displayed in the map (Fig. 1). The uniform trend of foliation and lineation in the study area suggests that the inhomogeneous deformation of the metagranites, localized into shear zones of variable length scale, took place within a coherent kinematic framework. Cross-cutting shear zones were not observed.

The present study is based on 15 oriented samples, selected from a larger number taken in the field. The sample locations are marked on the map (Fig. 1). The samples represent undeformed granites, granite gneisses and mylonitic shear zones, with a wide range of finite strain. The sample locations are spread over the entire outcrop area of the metagranites.

METHODS

The samples were cut normal to the foliation plane (*XY* plane) and parallel to the lineation direction (*X*). The quartz microstructure was investigated by optical microscopy in standard petrographic polished thin sections (~30 µm). The quartz CPO was determined by a combination of polarizing microscopy using an universal stage (Leitz Orthoplan) and by electron backscatter diffraction (EBSD) using a scanning electron microscope (SEM; LEO 1530) equipped with field emission gun and forescatter detector. The thin sections used for the EBSD technique were chemically polished using a colloidal silicon suspension (SYTON®) to minimize surface damage and then coated with carbon to limit charging effects. Operating conditions for EBSD analysis were an accelerating voltage of 25 kV and a working distance of 25 mm,

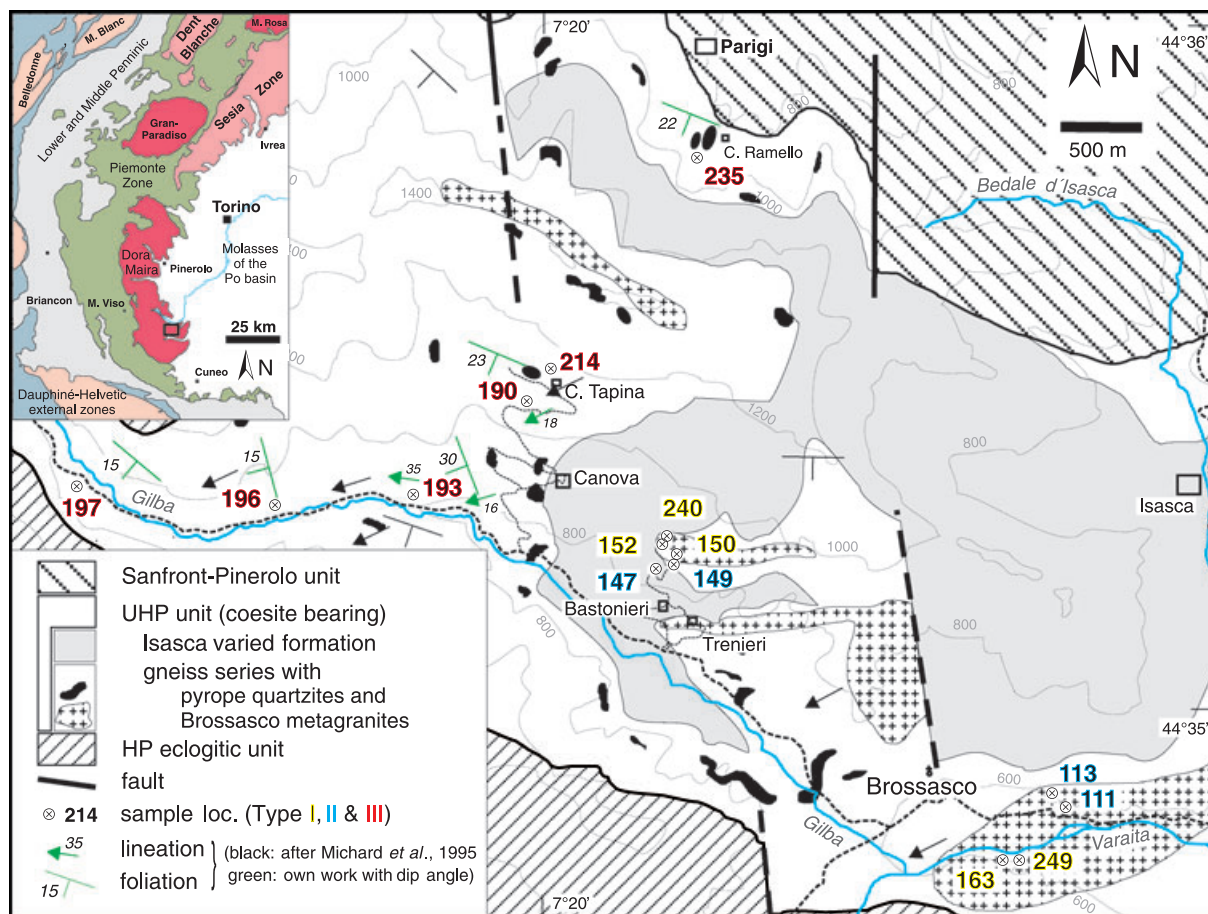


Fig. 1. Geological sketch map of the southern Dora Maira Massif after Henry *et al.* (1993) and Simon *et al.* (1997), showing the occurrence of metagranites in the UHP metamorphic unit and the sample locations. The black symbols without indication of the dip angle are taken from Michard *et al.* (1995); the green symbols with dip angles are observations made here.

with the section tilted over 70° . The EBSD patterns were identified and evaluated with the software CHANNEL 4 (Schmidt & Olesen, 1989). The orientation of the crystal axes is displayed with the program STERONETT (Duyster, 1997). Grain-size measurements were carried out by Digital Image Analysis (DIAna, 3.0; Duyster, 1997). The high-angle grain boundaries in quartz were traced by hand on an optical micrograph and then digitized by scanning. Analysis using DIAna yields the values of area, circumference, length of the long axis, orientation of long axis, length of short axis, and orientation of short axis for the individual grains. The long and short axes are defined as the longest and the shortest projected diameter of a particle, i.e. the longest and the shortest distance between parallel tangents on opposite particle interfaces, respectively. The longest and shortest axes are not necessarily at a right angle to each other.

Grain shape is displayed following a method proposed by Schwarz *et al.* (1989). In this grain-shape diagram, the aspect ratio of single quartz grains (longest diameter b divided by short diameter a , normal to b) is plotted along the X -axis. The Y -axis shows the

ratio of radius r_1 /radius r_2 for a grain. Radius r_1 is that of a circle with a perimeter $U = 2\pi r_1$ equal to the true grain perimeter. Radius r_2 is the radius of a circle with an area $F = \pi r_2^2$ equal to the true grain area. The ratio r_1/r_2 is unity for a spherical grain with aspect ratio 1. For higher aspect ratios, the lowest possible ratio r_1/r_2 is that holding for an ellipse. The limiting curve, calculated according to $U = \pi[3/2(a + b) - (ab)^{1/2}]$, is shown in each grain-shape diagram. For a given aspect ratio, a simple grain shape plots close to the limiting curve, whereas a more complex grain shape with sutured grain boundaries plots well above the limiting curve, as illustrated in a schematic grain-shape diagram (Fig. 4).

MICROSTRUCTURAL RECORD

The metagranites of the gneiss series, assumed to originate from a uniform granitic protolith (Chopin & Schertl, 2000), are subdivided into three types based on the degree of deformation. Type I represents the Brossasco metagranite, which is essentially undeformed and shows the original magmatic fabric as well as

angular xenoliths of various metamorphic rock types. Type II represents small-scale shear zones of decimetre-scale thickness cutting through the undeformed type I metagranites. Type III comprises the augengneisses and fine-grained granitic mylonites with pervasive ductile deformation.

Type I

The undeformed porphyritic Brossasco-type metagranites can be subdivided into type Ia and type Ib granites, based on the grain size of quartz and evidence of late-stage inhomogeneous crystal plastic deformation. Type Ia metagranites (samples AL240, AL152, AL150) have been found in one locality north of Bastonieri (Fig. 1) and are described in detail by Compagnoni *et al.* (1994). Sample AL240 comprises more than 50% fine-grained quartz, which forms interconnected layers with equal-sized grains and an average grain size of $\sim 50 \mu\text{m}$ (Fig. 2a). Samples AL150 and AL152 contain 30–40% of quartz in fine-grained aggregates, that are not interconnected. The average grain size is $\sim 100 \mu\text{m}$ in sample AL150 and $65 \mu\text{m}$ in sample AL152. In all samples, the fine-grained quartz matrix displays a foam structure with straight or slightly curved grain boundaries and approximately 120° angles at the grain edges (Fig. 2a,b). In addition, sample AL152 contains a few areas where the quartz grains show undulatory extinction with subgrains and highly sutured high-angle grain boundaries. Some of these quartz grains are elongate with a palisade structure (Fig. 2c). They occur adjacent to fine-grained aggregates with a foam structure. Within the areas with a foam structure, the orientation of the long axes of the quartz grains is random. There is no shape-preferred orientation (SPO) (Fig. 3a), consistent with the lack of any sign of a foliation in these samples. The grain-shape diagram reveals only slight deviation from isometric grain shape (Fig. 4b). EBSD measurements of quartz *c*- and *a*-axes show a random fabric for all three type Ia samples (Fig. 5a).

The quartz microstructure in type Ib metagranites, when compared with type Ia metagranites, is characterized by a somewhat larger grain size and partly bimodal grain-size distribution (Fig. 3b). The two investigated samples AL163 and AL249 were taken from an outcrop near the bridge over Torrente Varaita

at the road from Brossasco to Venasca (for further details, see also Michard *et al.*, 1995). The bimodal grain-size distribution of quartz in sample AL163 is characterized by maxima at $\sim 0.2 \text{ mm}$ and $\sim 0.6 \text{ mm}$ (Fig. 2d). The distribution of the two different grain-size domains betrays no systematic relation to impurities or neighbouring phases. Although most of the quartz grains show straight or simply curved grain boundaries with 120° angles at the triple junctions, a considerable portion of quartz crystals also shows undulatory extinction and deformation lamellae. Locally, subgrains occur and grain boundaries can be sutured, indicating minor grain boundary migration. Fluid inclusion trails cutting grain boundaries represent healed transgranular microcracks.

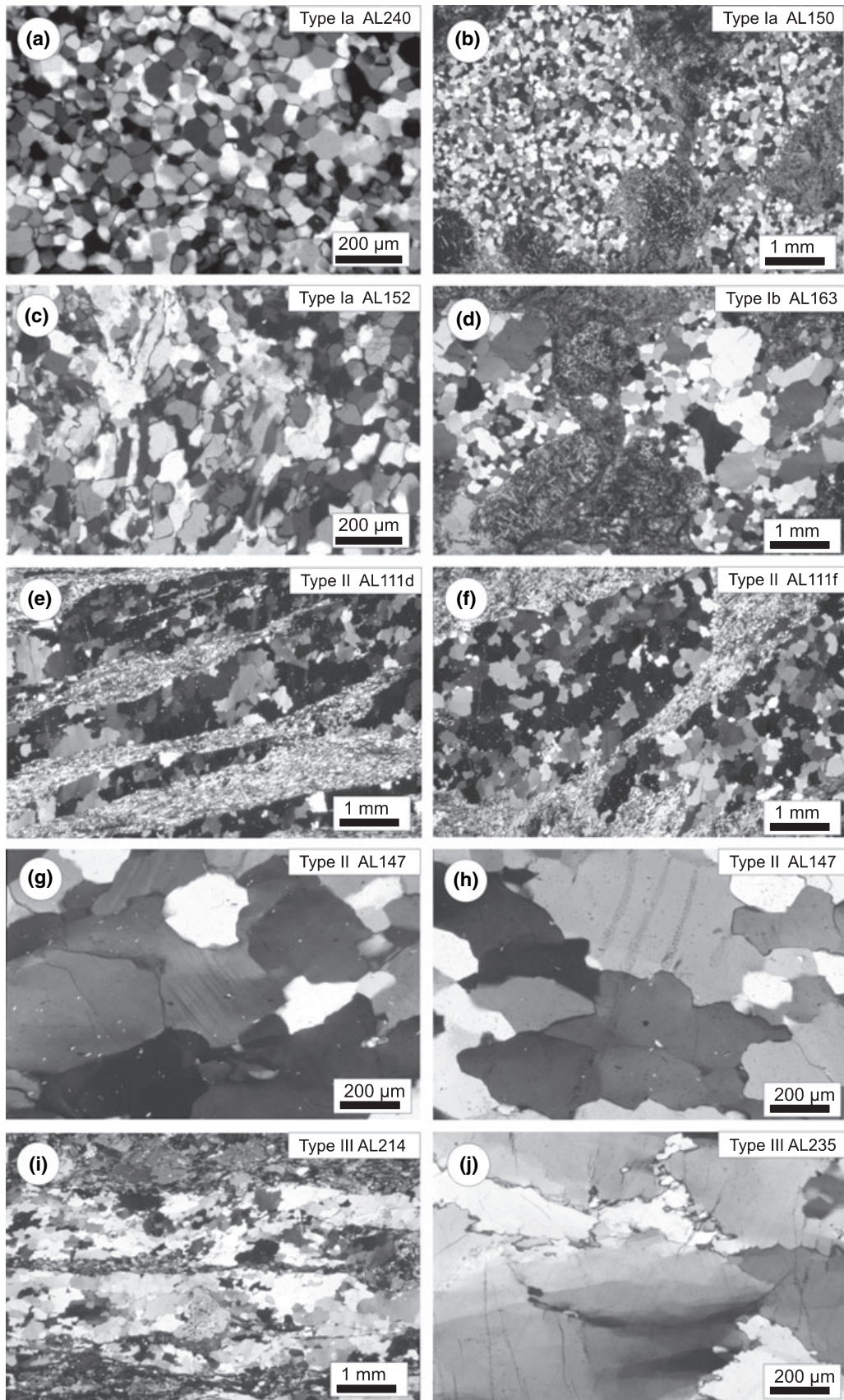
In sample AL249 the quartz grain size averages 0.6 mm . The quartz microstructure is highly heterogeneous with deformation bands and strongly sutured high-angle grain boundaries. An SPO of quartz or other minerals is not obvious. The grain-shape diagram reveals strong deviation from the ideal ratio of area to circumference, which would be maximum for elliptical grains. In both type Ib samples, the EBSD measurements of quartz *c*- and *a*-axes reveal no CPO (Fig. 5a).

Type II

The type II metagranites are found within type I metagranites near Bastonieri and additionally in an old quarry east of Brossasco (Fig. 1). The shear zones with a thickness of a few tens of centimetres dissecting the weakly deformed metagranites have been described by Compagnoni *et al.* (1994). The foliation is defined by lenses of the metamorphic minerals microperthite, albite/oligoclase, muscovite, clinozoisite, chlorite, biotite and titanite replacing igneous K-feldspar, plagioclase and biotite (Compagnoni *et al.*, 1994). Moreover, an SPO of the sheet silicates with (0 0 1) subparallel to the foliation is obvious. Porphyroclasts of K-feldspar are mantled by fine-grained recrystallized feldspar forming strain shadows.

Sample AL111 is taken from a small shear zone and includes the shear zone boundary at one side of the sample (Fig. 6). Two thin sections from sample AL111 will be compared below. AL111d represents the centre of the shear zone and AL111f a position at a distance

Fig. 2. Optical micrographs of characteristic quartz domains in the different metagranites (types I–III). (a) Quartz foam structure in sample AL240 of the undeformed Brossasco granite (type Ia). (b) Quartz foam structure in sample AL150. The fine-grained quartz aggregate has replaced the original igneous quartz crystals (type Ia). (c) Palisade structure of quartz adjacent to foam structure in sample AL152 (type Ia). (d) Bimodal grain size distribution with predominantly straight or slightly curved grain boundaries and 120° angles at triple junctions in sample AL163 (type Ib). (e) Alternating mica and quartz layers in sample AL111 near the centre of a small shear zone (type II). The quartz crystals show undulatory extinction and sutured grain boundaries. (f) Different area of the same sample as shown in (e), about 7 cm from the centre of the shear zone, with a less pronounced foliation. High-angle grain boundaries in quartz are only slightly sutured; undulatory extinction is inconspicuous. (g) Quartz grains with deformation lamellae in sample AL147 (type II). (h) Transgranular fractures marked by fluid inclusion trails in quartz in sample AL147 (type II). (i) Quartz layers in augengneiss (sample AL214; type III) showing undulatory extinction with subgrains and sutured high-angle grain boundaries. (j) Low-angle grain boundaries in millimetre-sized quartz grains and sutured high-angle boundaries in sample AL135 (type III).



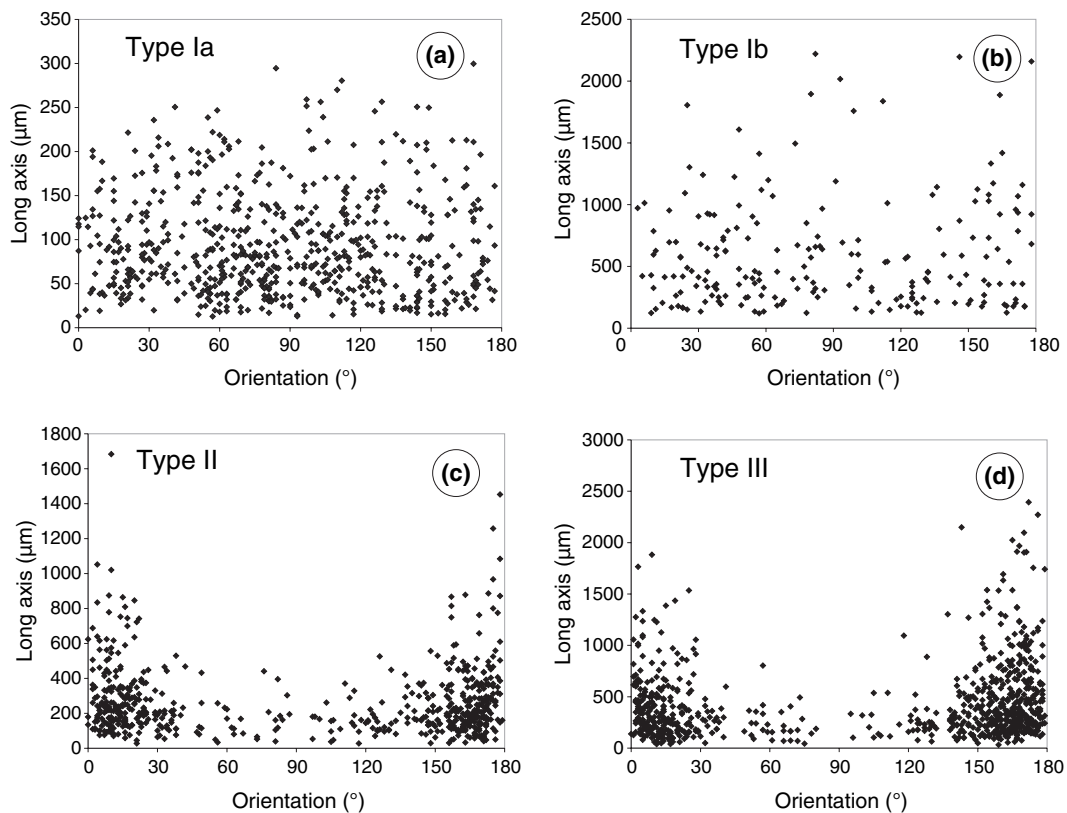


Fig. 3. Diagram showing the diameter measured along the long axis of quartz grains plotted against the orientation of the long axis relative to the X -axis measured in the XZ plane. For (a) type Ia rocks and (b) type Ib rocks a random orientation independent on grain size is indicated. For these rocks there is no shape preferred orientation of quartz. In contrast, a shape-preferred orientation is conspicuous for (c) type II and (d) type III rocks.

of ~ 7 cm from the centre of the shear zone. Both are XZ sections, cut perpendicular to the shear plane and parallel to the shear direction. Two foliations oblique to the shear plane are observed in this shear zone (Fig. 6). Compositional layering exhibits an increasing angle to the shear plane with increasing distance from the centre of the shear zone. In the centre, the angle between shear plane and compositional layering is $\sim 8^\circ$, at a distance of 5 cm from the centre the angle has increased to $\sim 30^\circ$, and with further increasing distance the compositional layering gets more and more diffuse. At a distance of ~ 10 cm the foliation is no more discernible. Slightly elongate quartz grains define the second foliation oblique to the compositional layering and with a larger angle to the shear plane. In sample AL111d, the angle between the SPO of quartz and the shear plane is about $30\text{--}60^\circ$ (Fig. 6). The average quartz grain size decreases from the centre ($260\text{ }\mu\text{m}$) to the boundary ($190\text{ }\mu\text{m}$) of the shear zone. In the centre of the shear zone, quartz grains show a more pronounced undulatory extinction (Fig. 2e). The grain boundaries tend to be more irregular (Fig. 4d), with a wavelength of the sutures typically between 20 and $100\text{ }\mu\text{m}$ and an average ratio $r_1/r_2 = 1.2$. In sample AL111f, the high-angle grain boundaries are only slightly sutured (Fig. 4d; average $r_1/r_2 = 1.16$), with a

wavelength of $< 20\text{ }\mu\text{m}$. Quartz domains in both samples show small inclusions of sheet silicates with a length of $20\text{--}50\text{ }\mu\text{m}$. In sample AL111d, these inclusions show a clear SPO, in contrast to sample AL111f, where the orientation is random. EBSD measurements of quartz c -axes orientation show a single girdle with a maximum parallel to the shear plane and perpendicular to the slip direction in the centre of the shear zone. Further away from the centre, quartz c -axes define a cross girdle (Fig. 6).

Sample AL113 is taken from another shear zone from the same quarry. Quartz has an average grain size of $150\text{ }\mu\text{m}$ and a higher aspect ratio (average $a/b = 1.8$) compared with sample AL111 (average $a/b = 1.6$). The grain boundaries are almost straight or slightly curved. The quartz c -axes orientation indicates a single girdle perpendicular to the foliation plane. The a -axes are preferentially oriented parallel to the XZ plane. These orientation patterns are almost identical to those in sample AL111d.

Samples AL147 and AL149 are taken from a location near Bastonieri, immediately adjacent to outcrops of the undeformed Brossasco metagranite (see Fig. 1). The quartz grain size varies between $\sim 300\text{ }\mu\text{m}$ (AL147) and $\sim 150\text{ }\mu\text{m}$ (AL149). Grain shape is more irregular (Fig. 4e; average $r_1/r_2 = 1.35$) compared with the other

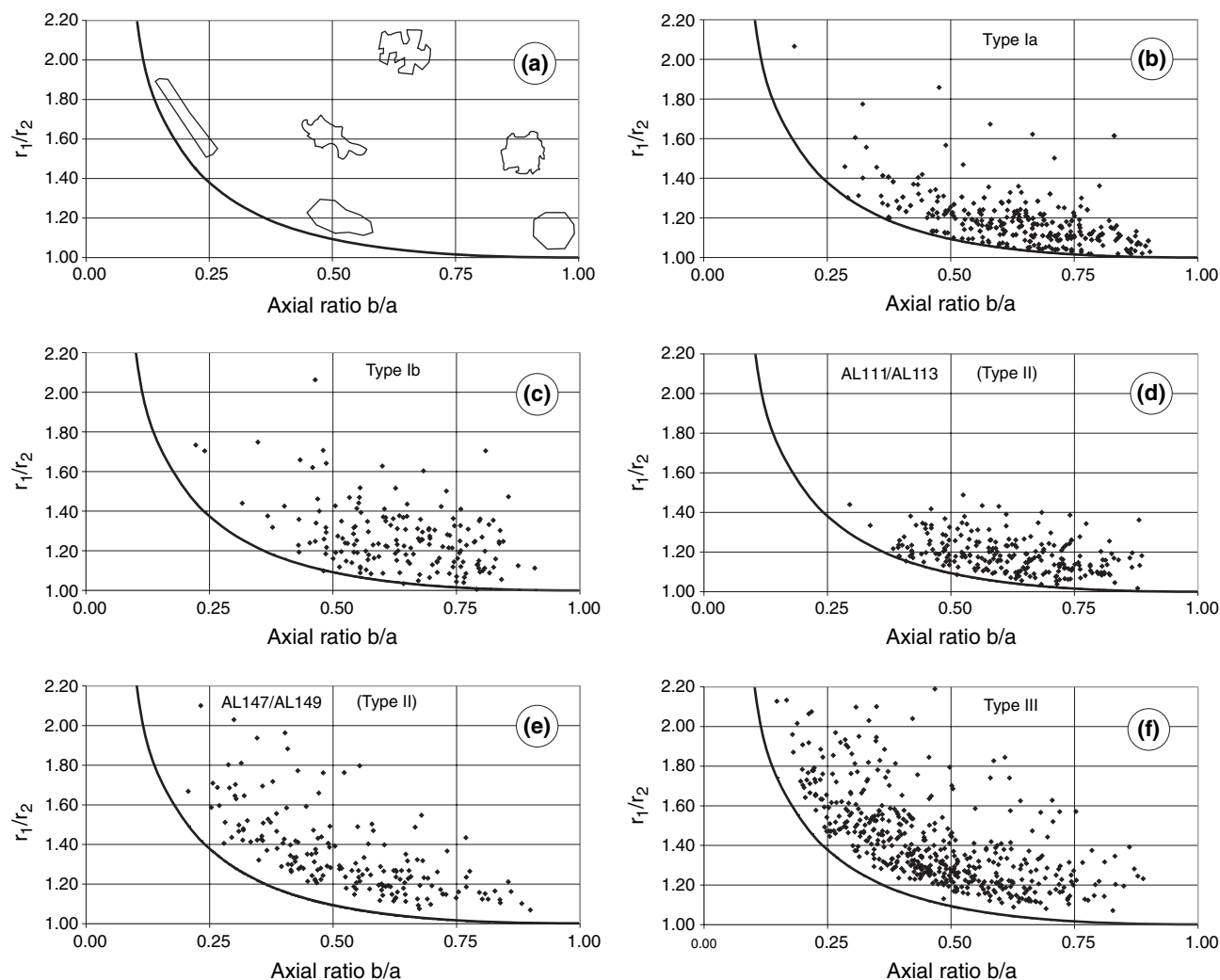


Fig. 4. Grain shape diagram (after Schwarz *et al.*, 1989) visualizing grain shape of quartz measured in XZ section. The parameters are described in the text. For each grain, the radius r_1 denotes that of a circle with equal area, and the radius r_2 that of a circle with equal perimeter. The aspect ratio is defined as b/a , with b denoting the diameter of a grain parallel to its short and a the diameter parallel to its long axis. The bolt curve shows the function holding for ellipses, which represents the limiting values. (a) The significance of the data points is visualized for some exemplary grain shapes. (b) Quartz grains in type Ia (270 measurements); (c) quartz grains in type Ib (172 measurements); (d) quartz grains in type II (sample AL111/113; 241 measurements); (e) quartz grains in type II (sample AL147/149; 180 measurements); (f) quartz grains in type III (489 measurements).

samples of type II (Fig. 4e,d). The larger grains show undulatory extinction, deformation bands and locally deformation lamellae (Fig. 2g). Fluid inclusion trails cutting several quartz grains indicate healed trans-granular microcracks (Fig. 2h). Both AL147 and AL149 show a marked SPO of quartz with the long axes parallel to the foliation. The texture in both samples displays a cross-girdle pattern of the quartz c -axes with a maximum of c -axes oriented parallel to the Y direction (Fig. 5b).

Type III

The structure of the type III metagranites varies between augengneiss and mylonite. The augengneisses

show a variable content of K-feldspar porphyroclasts of variable size, which represent the augen. The microstructures suggest that the feldspar augen and the adjacent strain shadows made up of fine-grained recrystallized feldspar developed from a core-and-mantle structure. The fine-grained recrystallized aggregates were weaker compared with the large clasts and deformation became localized in the mantle (Dell'Angelo & Tullis, 1989). Feldspar augen commonly develops under low- to medium-grade conditions (e.g. Debat *et al.*, 1978).

Samples AL193, AL196 and AL197 were taken along the Vallone di Gilba from east to west, AL190 and AL214 were taken north of Vallone di Gilba near Case Tapina, and AL235 was taken in direct contact

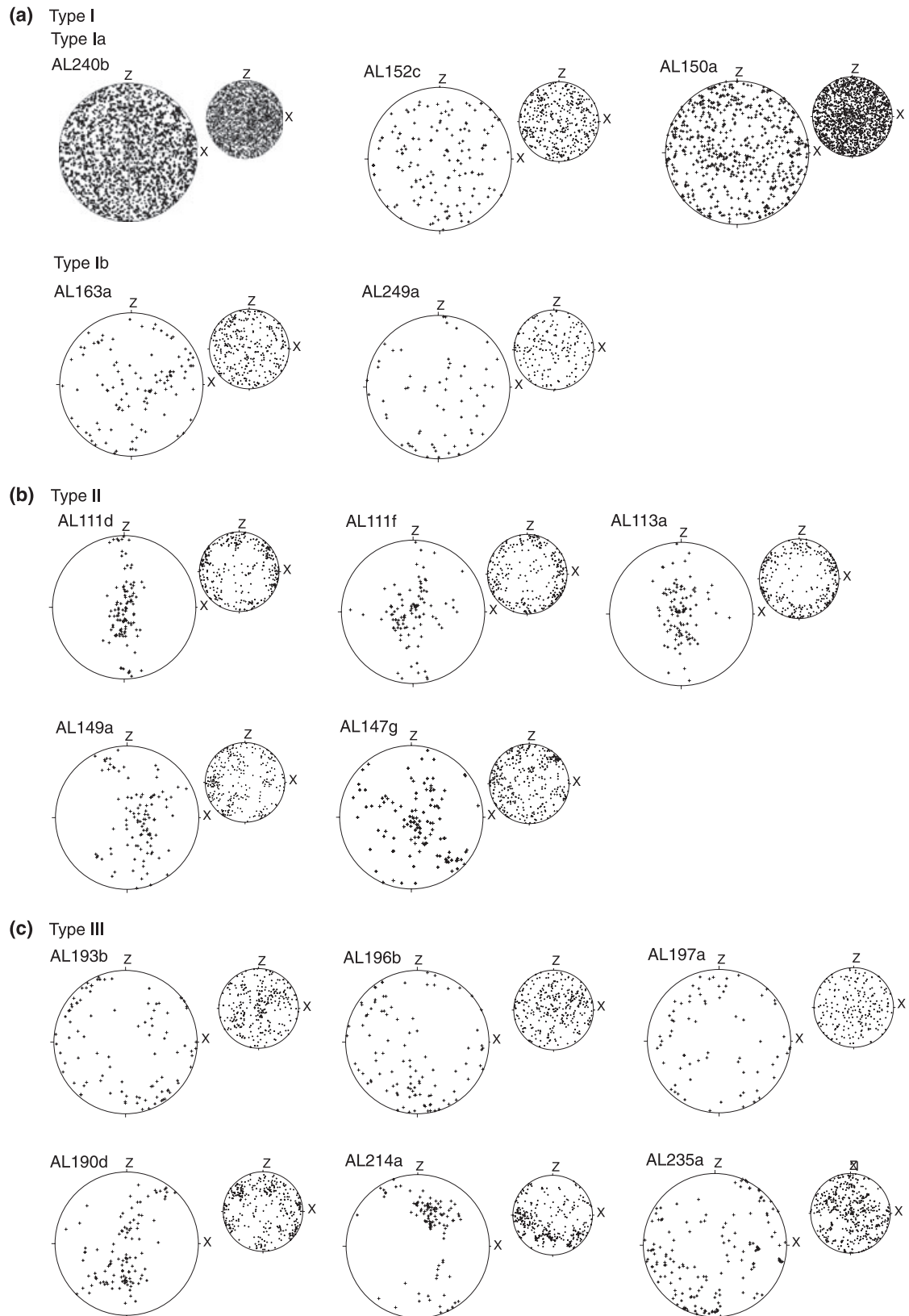


Fig. 5. Crystallographic orientation (texture) of quartz displayed in stereographic projections of *c*-axes (large diagram) and *a*-axes (small diagram). (a) Type I samples; sample numbers given in the diagram; (b) type II samples; sample numbers given in the diagram; (c) type III samples, sample numbers given in the diagram. See text for discussion.

with the pyrope-quartzite at the so-called Parigi outcrop (Fig. 1). Quartz in type III metagranites displays a strong SPO, with the long axes parallel to the lineation (Fig. 3d) emphasized by mica aggregates and feldspar augen (Fig. 2i).

The three samples from the Vallone di Gilba (AL193, AL196 & AL197) have low quartz contents between 15% and 25%. Single-phase domains of quartz are rare. Quartz grains are predominantly intermingled with feldspar grains. The average quartz grain size varies between 150 and 250 μm . The other samples contain between 30% (AL214) and 50% (AL235) of quartz with the average grain size varying between 0.3 and 0.45 mm. In these samples, quartz occurs in large single-phase domains and layers (Fig. 2i).

Quartz grain shape in all type III samples is highly irregular with sutured grain boundaries. In the quartz grain-shape diagram (Fig. 4f) most data points plot further away from the limiting curve for elliptical shape, when compared with type I and type II rocks. The axial ratio b/a is generally <0.5 .

A weak CPO of quartz is apparent in the gneisses from the Vallone di Gilba (Fig. 5c). An ill-defined point maximum of c -axes sub-perpendicular to the foliation is developed in samples AL193 and AL196, and a diffuse girdle in sample AL197. Both augen-gneiss samples from Case Tapina, AL190 and AL214, show distinct point maxima in their quartz c -axes orientation pattern and maxima in the corresponding girdle of a -axes orientations. The reason for the marked obliquity of the pattern in AL214 is not clear; it is probably an effect of strain inhomogeneity caused by a feldspar porphyroblast beyond the section plane.

In sample AL235, a lineation is not obvious and a different microstructure is evidenced compared with the other type III granite gneiss samples. Quartz forms lenses with a thickness of about a millimetre in the foliation plane. The grains are millimetre-sized with highly sutured grain boundaries, undulatory extinction, and abundant deformation bands as well as deformation lamellae (Fig. 2j). The c -axes orientations show a broad maximum at a high angle to the foliation (Fig. 5c).

DISCUSSION

P-T history of the investigated samples

When comparing the record of different metagranites of the Dora Maira Massif, the first question is whether all investigated rocks were subjected to similar UHP metamorphic conditions at some stage of their history and whether they followed similar *P-T* paths. This leads to the question of whether the metagranites form a coherent tectonic unit, notwithstanding their variable record.

Ultrahigh-pressure metamorphism in the stability field of coesite is unequivocal in the case of the unde-

formed Brossasco granite (type I metagranite). The peculiar quartz microstructures are interpreted to be a result of the coesite-quartz phase transformation (Biino & Compagnoni, 1992; Renner *et al.*, 2001; Stöckhert, 2002; Lenze *et al.*, 2005). In addition, preserved relict coesite has been identified within a xenolith in the Brossasco granite by Snoeyenbos (1995; cited in Chopin & Schertl, 2000).

As a consequence, the mesoscopic shear zones within the Brossasco granite (type II metagranite) must have experienced the same *P-T* history as their undeformed host rock (type I metagranite). The situation is less obvious for the huge volume of type III metagranites, which are not juxtaposed with the Brossasco granite (Fig. 1). However, a similar *P-T* history is suggested by the following observations: (i) relics of UHP metamorphic mineral assemblages (clinopyroxene with garnet and phengite) were identified in some granite gneisses (Chopin & Schertl, 2000), (ii) the type III metagranites are intimately associated with the spectacular UHP metamorphic pyrope-quartzites (Chopin *et al.*, 1991; Schertl *et al.*, 1991; Henry *et al.*, 1993), and (iii) the characteristics of the original magmatic rocks appear to be very similar to those of the Brossasco granite, suggesting a uniform plutonic protolith (Chopin & Schertl, 2000).

Although not providing a definite proof, the above observations suggest a common history and similar *P-T* paths for all investigated samples. The metagranites are therefore proposed to originate from the same Carboniferous (Tilton *et al.*, 1991; Gebauer *et al.*, 1997) plutonic body and to form a coherent crustal slice, notwithstanding the variable record of late-stage deformation and overprint.

Quartz microfabrics in the undeformed Brossasco granite (type I)

In the undeformed Brossasco granite, the undistorted original magmatic quartz domains are comprised of fine-grained aggregates, with a grain size between 50 and 100 μm . This reduction in grain size cannot be a simple consequence of deformation, as the original magmatic structure is preserved. Furthermore, the fine-grained quartz aggregates do not show a crystallographic preferred orientation (Fig. 5a). A CPO would be expected where the small grain size is the result of dynamic recrystallization during deformation by dislocation creep. Finally, most quartz aggregates show a well-developed foam structure (Fig. 2a,b), indicating grain growth driven by the reduction of interfacial free energy (e.g. Evans *et al.*, 2001). As static recrystallization can weaken but not obliterate a CPO (Otani & Wallis, 2006), dynamic recrystallization caused by deformation with subsequent annealing can be excluded. Nearly identical microstructures (Fig. 2a,b,d) were found in experimental studies on the coesite-quartz transformation (Wirth & Stöckhert, 1995; Mosenfelder & Bohlen, 1997; Schönbohm, 2003; Lenze

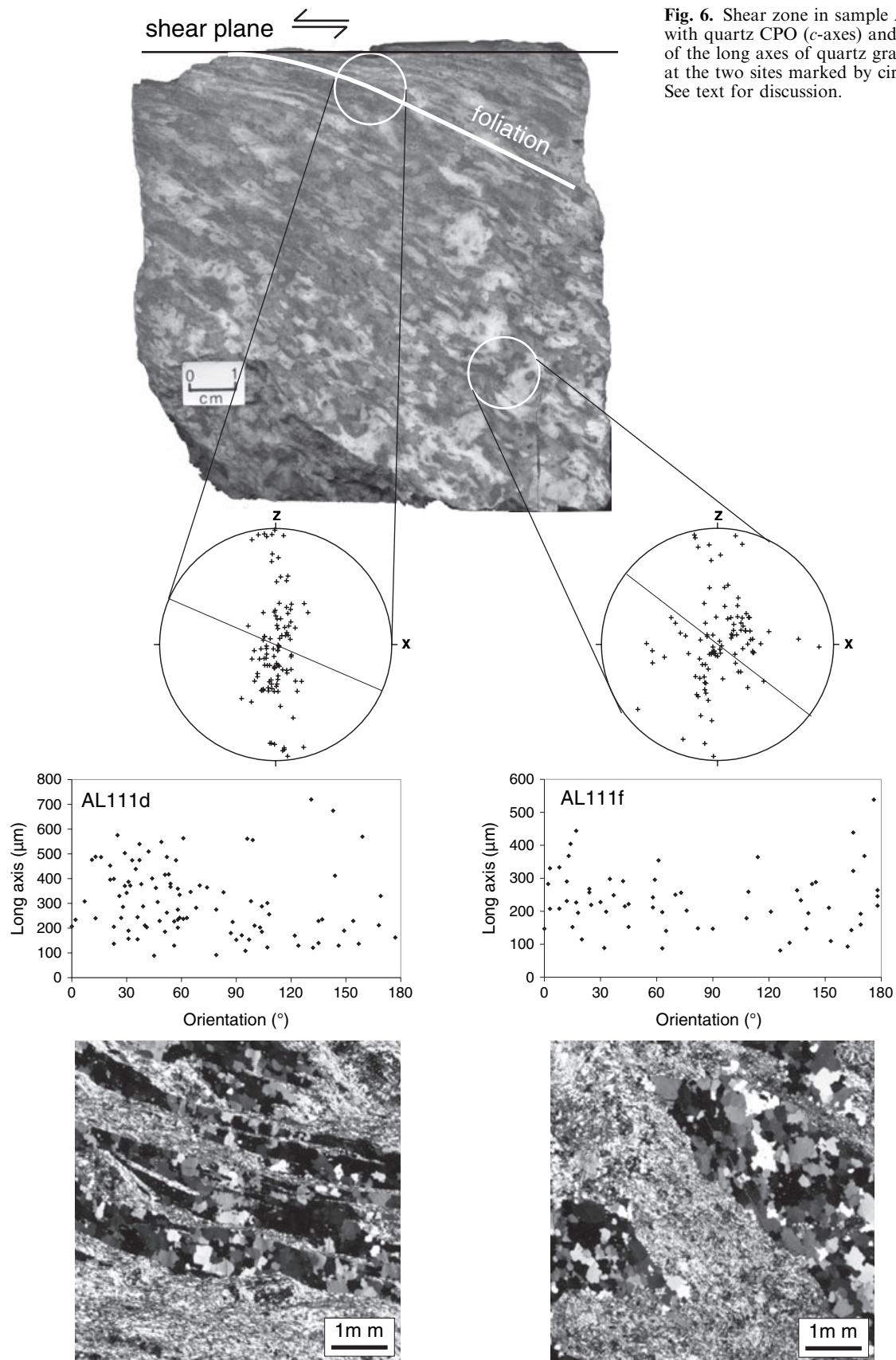


Fig. 6. Shear zone in sample AL111 (type II) with quartz CPO (*c*-axes) and SPO (orientation of the long axes of quartz grains) measured at the two sites marked by circles in the image. See text for discussion.

et al., 2005). Based on the microstructures observed in samples quenched at various stages of transformation, Lenze *et al.* (2005) proposed the involvement of several stages: (i) nucleation of quartz and rapid growth along cracks and grain boundaries of coesite; (ii) plastic deformation of quartz driven by inhomogeneous volumetric strain; (iii) recrystallization of plastically deformed quartz; and (iv) static grain growth during annealing after complete transformation. A similar sequence of processes is expected to have resulted in the peculiar quartz microfabrics of the Brossasco (type I) granite. Lenze *et al.* (2005) have argued, referring to their observations in UHP metamorphic pyrope quartzites and jadeite quartzites of the Dora Maira Massif, that evidence of high differential stress at high temperatures implies rapid phase transformation comparable with laboratory time scales. In those rocks, quartz displays a foam microstructure with a grain size similar to 0.5 mm, which is at least one order of magnitude larger than that in the typical Brossasco (type I) granite. As the kinetics of grain growth are strongly dependent on temperature (e.g. Evans *et al.*, 2001), this contrast possibly indicates a somewhat lower temperature at the time of the coesite-quartz transformation in the Brossasco granite, or differences in the concentration of H₂O or water-related defects, which have been found to affect the rate of transformation (Mosenfelder & Bohlen, 1997; Schönbohm, 2003; Lathe *et al.*, 2005).

An additional hint to the former presence of coesite in the Brossasco (type I) granite is the observation of relict palisade quartz within the fine-grained polygonal quartz matrix (Fig. 2c). These palisade structures in quartz aggregates formed from coesite are observed in many UHP metamorphic rocks (e.g. Chopin, 1984) and were also found in laboratory experiments (e.g. Wirth & Stöckhert, 1995; Mosenfelder & Bohlen, 1997; Schönbohm, 2003), albeit on smaller length scales.

A distinction into type Ia and Ib metagranites is made to take into account the variable degree of late inhomogeneous crystal-plastic deformation superimposed on the annealed quartz aggregates. Otherwise, type Ib metagranites show the same preserved magmatic fabric and the same random crystallographic orientation inside the fine-grained quartz aggregate with foam structure as type Ia metagranites.

Notably, the fine-grained foam microstructure of quartz without CPO is exclusively observed in the undeformed Brossasco granites (type I) and no sample taken from the mylonitic shear zones (type II metagranites) or augengneisses (type III metagranites) shows similar quartz microfabrics. Such microfabrics would be expected for type II and III metagranites as well, if their deformation had taken place during burial in the subduction zone and UHP metamorphism, and prior to the transformation of coesite to quartz during exhumation. The fine-grained quartz aggregates with foam microstructure and lacking a CPO are restricted

to metagranites which remained undeformed through their entire burial and exhumation history.

Quartz microfabrics in shear zones within the Brossasco granite (type II)

Type II metagranites are the result of ductile deformation localized in shear zones within the Brossasco granite. In principle, this inhomogeneous deformation of the granite body could have taken place at any time between solidification in a magma chamber in the Carboniferous and exhumation into the upper crust in the Late Oligocene. As UHP metamorphism is indicated for the undeformed host rock (type I metagranites), the material in the decimetre- to metre-thick shear zones must have experienced the same conditions causing temporary replacement of quartz by coesite. Coesite is not preserved, however, and the quartz microfabric in the shear zones differs markedly from that in the adjacent undeformed Brossasco (type I) metagranite.

In type II metagranites the quartz microstructure is characterized by the abundance of low-angle grain boundaries, sutured high-angle grain boundaries and a pronounced SPO with high aspect ratios. In addition, the pole figures (Fig. 5b) reveal a CPO, which is somewhat variable for the investigated samples.

Most obvious for samples AL111d and AL113 (Fig. 5b), the preferred orientation indicates the activation of prism $\langle a \rangle$ and rhomb $\langle a \rangle$ slip (Lister & Price, 1978; Hobbs, 1985; Law, 1990). For these samples, the single-girdle c -axes patterns indicate simple shear (Burg & Laurent, 1978; Lister & Price, 1978; Bouchez *et al.*, 1979). Cross-girdle patterns in samples AL111f, AL147 and AL149 may not only indicate a more important component of pure shear, but were also attributed to lower temperature and lower strain (Schmid & Casey, 1986). In the polyphase rocks with centimetre-sized feldspar crystals, stress inhomogeneities around rigid feldspar porphyroclasts can be the cause for a weak CPO (e.g. Lister & Price, 1978; Handy, 1990) or a more complicated pattern, as observed in samples AL147 and AL149. When the feldspars are not cut by the thin section, the effect of such porphyroclasts on the CPO in adjacent quartz domains cannot be identified. With the restrictions related to the grain-scale inhomogeneity of the strain field in polyphase rocks, the quartz microfabrics indicate deformation by dislocation creep accompanied by recrystallization and recovery. Predominance of prism $\langle a \rangle$ and rhomb $\langle a \rangle$ slip point to moderate temperatures (Lister & Price, 1978; Hobbs, 1985; Law, 1990), consistent with the greenschist facies conditions inferred from the mineral assemblage (Chopin *et al.*, 1991).

In sample AL113, quartz shows a CPO and SPO, although with a lower grain aspect ratio compared with the other samples. In addition, most high-angle grain boundaries are nearly planar or simply curved.

The microstructure indicates grain growth driven by the reduction of interfacial free energy (e.g. Cotterill & Mould, 1976; Evans *et al.*, 2001), also referred to as grain boundary area reduction (e.g. Barnhoorn *et al.*, 2004) or grain boundary energy-driven grain boundary migration (GBEGBM; Bons & Urai, 1992). This process is expected to proceed during annealing at low stress, after reduction of stored strain energy by recovery and recrystallization, at temperatures allowing effective grain boundary migration. Bons & Urai (1992) used experiments and computer simulations to demonstrate that post-tectonic GBEGBM can partially or completely destroy an SPO, eventually creating a foam structure. A true foam structure is not developed in sample AL113, but the microstructure indicates a more pronounced effect of GBEGBM compared with the other samples. This observation suggests that the shear zones in the Brossasco granite (type II metagranites) did not necessarily form simultaneously during a single stage of tectonic activity; instead, they record slightly differing stress–temperature histories.

Quartz microfabrics in granite gneisses (type III)

The granite gneisses (type III metagranites) make up a large portion of the UHP unit of the Dora Maira Massif (Fig. 1). Their microfabrics are similar to those of the shear zones (type II metagranites) within the otherwise undeformed Brossasco granite. The quartz microstructures characterized by sutured high angle grain boundaries, SPO, and low-angle grain boundaries indicate deformation by dislocation creep, as does the CPO in the majority of the samples. In three samples of the Valle Gilba (AL193, AL196 & AL197), a CPO is missing or at least very weak (Fig. 5c). In these samples, quartz is dispersed in aggregates of limited size and the orientation patterns are combined from several isolated domains within a certain area. For such fabrics, diffuse orientation patterns are supposed to arise from heterogeneous flow at the grain scale (Law, 1990). The mechanisms leading to the asymmetry of the texture in samples AL190 and AL214 (Fig. 5c) are not understood. The CPO compares well with patterns described by Schmid & Casey (1986; their samples CC1 & R405), which are interpreted to result from $\langle a \rangle$ slip only. The quartz textures described by Schmid & Casey (1986) are found in rocks deformed at greenschist facies conditions.

Microfabrics along a strain gradient in a type II metagranite (shear zone)

Shear zones allow correlation of fabrics and shear strain. In the present study, this approach provides a test of whether the shear zone developed at a late stage after transformation of coesite to quartz, or whether it formed earlier during subduction or UHP metamorphism in the stability field of coesite. In the latter

case, a correlation between the CPO and the shear strain would not be expected. The marked difference in fabric asymmetry between two domains in sample AL111 with low and high shear strain, respectively, is shown in Fig. 6. In a detailed analysis, Gapais & Cobbold (1987) described the gradual rotation of the overall quartz fabric pattern accompanying the rotation of the schistosity towards the shear direction with increasing shear strain. Similar correlations between CPO asymmetries are observed in sheared ice (Hudleston, 1980) and in peridotites (e.g. Nicolas & Poirier, 1976). Moreover, quartz *c*-axes patterns are observed to develop from a single girdle to a cross girdle with increasing distance from the centre of a shear zone, hence the decreasing shear strain (Schmid & Casey, 1986). The correlation of shear strain with the quartz CPO pattern documented in Fig. 6 indicates that the shear zone formed in the stability field of quartz at a late stage of exhumation and not during burial or UHP metamorphism in the coesite stability field.

Evidence for minor late-stage deformation of quartz at high stress

Apart from type Ia metagranites, where the small grain size precludes the respective observations in the optical microscope, deformation lamellae in quartz are observed in many samples. Deformation lamellae form in the regime of low-temperature plasticity at high stress (Drury, 1993). In addition, healed cracks marked by fluid inclusion trails are widespread, recording brittle deformation at moderate temperatures (Smith & Evans, 1984). This minor deformation reflects late-stage processes in the upper crust with very low strain.

Exhumation history and related deformation

All shear zones in the Brossasco granite and all granite gneisses show quartz microfabrics indicating deformation by dislocation creep and a corresponding CPO, which is found to correlate with shear strain. In contrast, quartz crystallized at the expense of coesite in the undeformed Brossasco granite has a foam microstructure and no CPO. This indicates that deformation in the shear zones and granite gneisses invariably occurred after transformation of coesite to quartz during exhumation. The quartz microfabrics suggesting predominantly $\langle a \rangle$ slip and the mineral assemblages indicate greenschist facies conditions (Henry *et al.*, 1993; Compagnoni *et al.*, 1994). Hence, all significant deformation took place in the middle crust at temperatures of about 400 ± 100 °C, at a late stage of exhumation.

Among the models proposed for the exhumation of UHP metamorphic rocks, the return by back flow in a narrow subduction channel (Cloos & Shreve, 1988) provides an explanation for rapid exhumation (e.g. Avigad *et al.*, 1993; Gerya & Stöckhert, 2002; Gerya *et al.*, 2002; Jolivet *et al.*, 2003). In the case of the Dora

Maira Massif, the age of UHP metamorphism is given as 35.1 ± 0.9 Ma (at 110 ± 10 km depth) and that of a stage after decompression (at 35 ± 5 km depth) as 32.9 ± 0.9 Ma (Rubatto & Hermann, 2001). This yields a minimum exhumation rate of about 34 km Myr^{-1} . (Rubatto & Hermann, 2001). Various mechanisms are envisaged for this process, most of which having in common some kind of strain localization in a low-viscosity material allowing rapid exhumation of large rigid slices (e.g. Ernst, 1971; Michard *et al.*, 1995; Stöckhert, 2002; Jolivet *et al.*, 2005).

The nature of these HP and UHP shear zones is still poorly determined and there is not much unequivocal record preserved in UHP metamorphic terranes. HP and UHP metamorphic shear zones are, for instance, described in much detail by Terry & Heidelberg (2004) from the Norwegian Caledonides, and reported from the Dabie Sulu Belt in China by Zhao *et al.* (2005). In contrast, the shear zones in the UHP metamorphic Brossasco granite and the neighbouring granite gneisses in the UHP metamorphic Dora Maira Massif can be ruled out as candidates for localized deformation during burial and exhumation at deep levels. For all these rocks, deformation has taken place in the stability field of quartz at greenschist facies metamorphic conditions. At the time of deformation, these rocks resided in the middle crust with a normal geotherm. In numerical simulations (Stöckhert & Gerya, 2005; Gerya & Stöckhert, 2006), deformation of UHP metamorphic continental crust at similar conditions is predicted after extrusion from a subduction channel at an active continental margin.

Although there is no definite proof, the Brossasco granite and the adjacent granite gneisses very likely have a common origin and formed a coherent crustal slice during subduction and exhumation. This is indicated by the close spatial correlation, the petrographic similarity, and the presence of various relics of UHP metamorphic overprint in the gneisses and their inclusions. Indications of notable deformation during subduction, during UHP metamorphism, and during early exhumation were not identified. In contrast, there is compelling evidence that the mesoscopic shear zones formed late after the rocks had returned into a mid-crustal environment. The present outcrop area of the metagranites (Fig. 1) suggests that the diameter of rigid slices transported in subduction channels could be up to kilometres.

SUMMARY AND CONCLUSIONS

The undeformed UHP metamorphic Brossasco-type granites reveal a broad spectrum of quartz microstructures, ranging from fine-grained quartz aggregates with a foam microstructure to aggregates of coarser grains with irregular grain shape. There is no CPO of quartz. The microstructural differences are suspected to reflect the specific *P–T* conditions of the coesite–

quartz transformation in the individual rock, and possibly differences in the type and content of incorporated H_2O and H_2O -related defects. The fine-grained aggregates with a foam microstructure correspond to those produced in coesite–quartz transformation experiments on laboratory time scales (Lenze *et al.*, 2005).

Shear zones in the metagranites and augengneisses are characterized by a quartz microfabric with highly irregular grain shape, sutured high-angle grain boundaries, undulatory extinction and low-angle grain boundaries, indicating dynamic recrystallization and recovery during deformation by dislocation creep. There is a marked CPO of quartz in all samples from shear zones and augengneisses. The patterns are quite variable in detail; however, probably due to the inhomogeneous strain field caused by the centimetre-sized feldspar porphyroclasts. Invariably, the patterns show the predominance of $\langle a \rangle$ slip on basal, prism and rhombohedral planes. In combination, the microstructures and the CPO indicate deformation at greenschist facies conditions.

Notably, there is no record of any significant deformation of the granites during burial and exhumation at HP and UHP conditions. All deformation has taken place at a late stage, after exhumation into a mid-crustal environment with a normal geotherm. Based on petrographic similarity, field relations and relics of UHP metamorphism, the granite gneisses and the undeformed UHP metamorphic Brossasco granite are suspected to represent a coherent body with a common history. If this is true, kilometre-sized slices of continental crust can be buried to depths of 100 km and exhumed to a mid-crustal position, without internal deformation. All deformation during subduction and early exhumation must have been localized in weak shear zones of considerable spacing, none of which apparently being represented by the deformed portions of the granite exposed in the Dora Maira Massif. Furthermore, the observations indicate that deformation became increasingly distributed after exhumation into a mid-crustal environment, presumably because the peculiar deformation mechanisms allowing localization into weak shear zones at (U)HP conditions had become ineffective.

ACKNOWLEDGEMENTS

We thank H.-P. Schertl for providing invaluable information on outcrops and petrology of the Brossasco-Isasca unit, and R. Neuser for assistance during SEM work. Financial support by the Deutsche Forschungsgemeinschaft within the scope of collaborative research center 526 'Rheology of the Earth – from the Upper Crust into the Subduction Zone' (SFB 526) is gratefully acknowledged. Constructive reviews by two anonymous referees, in one case with very detailed and valuable comments, and the helpful suggestions of the editor are greatly appreciated.

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Received 27 September 2006; revision accepted 22 February 2007.