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Geometry and structural evolution of ultra-high-pressure and high-pressure rocks from the Dora-Maira massif, Western Alps, Italy

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Abstract—The crystalline nappes of the Dora-Maira massif, Western Alps, essentially made of continental material from the upper crust, show petrological relics of an ultra-high-pressure (UHP) to high-pressure (HP, 'cold' eclogite) Eoalpine metamorphism. They also display relics of UHP–HP structures, preserved in boudins and/or within large UHP porphyroclasts, in a retrograde, greenschist-facies regional deformation fabric.

The greenschist-facies overprint has the character of a shallow-dipping mylonitic foliation (S_m), bearing a penetrative stretching lineation (L_m) which roughly parallels the axes of coeval, isoclinal folds. Shear sense markers indicate a W-verging overthrust mechanism.

The UHP and HP relic structures are of variable nature. The coexistence of equant and inequant, either symmetric or asymmetric fabrics, indicates that the deformation at UHP–HP conditions was strongly heterogeneous and partitioned. This is also supported by the local preservation of Hercynian, magmatic fabrics. The UHP and HP deformation involved, at least locally, rotational components, although less intensive than during the latter retrograde stage.

The regional structural evolution is envisaged as follows: (i) the Eoalpine subducted crust was subdivided into lenticular bodies surrounded by UHP–HP shear zones. The main part of the exhumation processes remains unconstrained due to the sparseness and late rotation of the UHP–HP structural relics; conflicting models are possible depending on the interpretation of the early sense of movement (normal vs reverse) along the faults that limit the lens-shaped units; and (ii) the late, heterogeneous, regional greenschist deformation can be attributed to the Eocene collapse of the Alpine orogenic wedge.

INTRODUCTION

COESITE within crustal metamorphic rocks was first discovered in the southern Dora-Maira crystalline massif of the Western Alps (Chopin 1984). Several other occurrences have been discovered since in Alpine or pre-Alpine collisional belts, either in continental (Smith 1984, Okay *et al.* 1989, Smith & Lappin 1989, Wang *et al.* 1989, Yang & Smith 1989, Enami & Zang 1990) or oceanic rocks (Reinecke 1991). In such belts crustal rocks, even of continental origin, must have been subducted to a depth of 100 km, then uplifted and exhumed, yet preserving significant relics of the mineral associations formed at great depth.

In this paper we present structural geology data for the type-UHP (ultra-high pressure) rocks of the Dora-Maira massif together with the associated crustal rocks. Other aspects such as the petrology–mineralogy (Chopin 1987, Massonne & Schreyer 1989, Chopin *et al.* 1991, Kienast *et al.* 1991, Schertl *et al.* 1991) and isotopic geochemistry (Paquette *et al.* 1989, Tilton *et al.* 1989, 1991, Monié & Chopin 1991) of these rocks have already been thoroughly investigated. We particularly address the following question: do the UHP rocks preserve relics of structures imprinted during the UHP metamorphism, and/or relics of even older structures? Such structures can enlighten aspects of the mechanical behaviour of crustal rocks at subasthenospheric depth. The significance of these structural data for the exhumation history of the UHP rocks is also analysed.

GEOLOGICAL SETTING

The Dora-Maira Penninic massif

The Dora-Maira crystalline massif (Vialon 1966) is exposed in the internal part of the Western Alps, structurally beneath the ophiolitiferous 'Schistes lustrés' (Ligurian Tethys) nappe (Fig 1). As shown by Argand (1911, 1934), Michard (1965, 1967) and Borghi *et al.* (1984, 1985), the massif includes three major superposed continental units or complexes, which are from bottom to top: (i) the Sanfront-Pinerolo unit (SP) involving granitic and dioritic gneiss, Carboniferous (?) graphitic–conglomeratic schist, Permian–Lower Triassic metaclastic rocks and thin Middle Triassic carbonates; (ii) the main Dora-Maira Basement Complex (BC) consisting almost exclusively of polymetamorphic schist and orthogneiss derived from a pre-Alpine (Hercynian) basement, associated with some Permian–Triassic metasediments; and (iii) the Dronero-Sampeyre unit (DS, also named Jouglard-Selleries and Depot complexes in the northern part of the massif by Borghi *et al.* 1984), which mainly consists of some polymetamorphic schist and widespread Permian–Triassic meta-volcanoclastic rocks. Post-Triassic metasediments have recently been recognized on top of this unit, including Late Creta-

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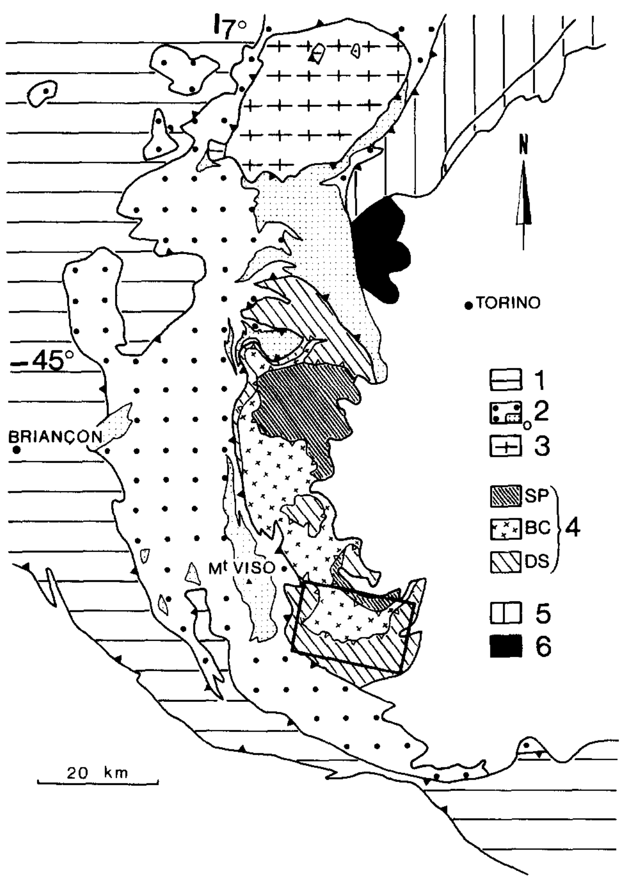

Fig. 1. The Dora-Maira massif in the Western Alps, with location of the studied area (framed). 1—Briançonnais zone, basement and post-Paleozoic cover (undifferentiated). 2—Schistes lustrés nappe; o—main ophiolitic massifs. 3—Gran Paradiso crystalline massif. 4—Dora-Maira crystalline massif (with SP—Sanfront-Pinerolo graphitic unit; BC—Basement Complex; DS—Dronero-Sampeyre unit). 5—Austroalpine and Sudalpine zones. 6—Lanzo ultramafic body.

ceous (?) olistolithic calcschists (Caron 1977, Marthaler *et al.* 1986, Philippot 1988, Henry 1990).

The Sanfront-Pinerolo unit is comparable with the money window in the Gran Paradiso massif (Compagnoni *et al.* 1974); both are part of the Middle Penninic Grand Saint Bernard (Briançonnais) nappe. The overlying Dora-Maira complex and Dronero-Sampeyre unit, and their equivalents in the Gran-Paradiso massif, are part of the Upper Penninic Monte Rosa nappe. The Grand Saint Bernard and the Monte Rosa nappes are commonly regarded as the European paleomargin (Argand 1991, 1934, Michard 1967, Lemoine *et al.* 1986, Coward & Dietrich 1989, Stampfli & Marthaler 1990), although different restorations of the European–Ligurian–Apulian transect have been proposed (e.g. Philippot 1988, Hunziker *et al.* 1989).

The UHP rocks in the southern Dora-Maira massif

Recently the investigated area has been mapped (1:50,000) by Henry (1990). The southern crystalline 'massif' was shown to consist of a heterogeneous stack of lens-shaped tectonic units, each a few hundred metres thick, upwarped by a late kilometre-scale antiform (Figs. 2 and 3). Each unit is characterized by a specific association of rock lithologies and/or by a specific metamorphic evolution (Fig. 4).

The Sanfront-Pinerolo unit is mainly represented by graphitic schists equilibrated under blueschist conditions (10–12 kbar, 500°C). The overlying Basement Complex (BC, Fig. 1) is represented in this area by a pile of three eclogitic crystalline units, numbered I–III from bottom to top (Figs. 2, 3 and 4). The coesite-bearing unit I, about 10 × 5 × 0.5 km in size, consists of continental crystalline and polymetamorphic material (pseudomorphs of pre-Alpine sillimanite and cordierite, Biino & Compagnoni 1991). It has been affected by UHP metamorphism (Chopin *et al.* 1991) and equilibrated at 28–35 kbar and 700–750°C. It underlies the 'cold'-HP in the following—eclogitic units II and III, containing lithologies roughly similar to those of unit I, equilibrated at about 16–20 kbar and 550–600°C, and 15–20 kbar and 500–600°C, respectively. On top of this high-grade pile (I–III), the Lower Varaita Ophiolitiferous Band (OB), the Dronero-Sampeyre volcanoclastic unit and the overlying Schistes lustrés nappe equilibrated under blueschist conditions near 10–12 kbar and 500°C. Some slices of the M. Viso ophiolites include eclogite relics, formed at about 12–15 kbar and 500–550°C (Lombardo *et al.* 1978, Nisio *et al.* 1987, Philippot 1988). The whole Dora-Maira pile shared a retrograde decompression history at intermediate temperature, from about 10 kbar and 500°C (blueschist–greenschist boundary, R1 on Fig. 4) to greenschist-facies conditions, close to 3 kbar and 350–400°C (R2).

The various rock-types within the UHP tectonic unit I form mostly lens-shaped bodies of metric to hectometric size. For example (see Figs. 2 and 11a), the pyrope-bearing quartzite itself (where coesite was discovered first) constitutes one to three levels of boudinaged lenses, about 0.5–5 m thick, either within fine-grained gneiss (Parigi gneiss) or along the contact between the gneiss and a varied metasedimentary and metavolcanic formation (Isasca series). On top of unit I, the Gilba mylonites represents another example of a lens-shaped rock-unit at a larger scale (300 m of maximum thickness).

Geochronological data

The last part of the ductile deformation history, coeval with greenschist-facies metamorphism, is dated at 45–30 Ma throughout the Dora-Maira pile (Fig. 4) by Rb–Sr and ^{40}Ar–^{39}Ar methods (Vialette & Vialon 1964, Scaillet *et al.* 1990, Monié & Chopin 1991). Similar ages are known from the overlying 'Schistes lustrés' units (Liewig *et al.* 1981, Dal Piaz & Lombardo 1985).

Dates for the UHP–HP events are controversial. The highest-grade continental units I, II and III seem to have had their peak metamorphic conditions during the mid-Cretaceous or early Late Cretaceous. According to U–Pb zircon, Rb–Sr (Paquette *et al.* 1989) and ^{40}Ar–^{39}Ar (Monié & Chopin 1991) results, the initial cooling of unit I after UHP metamorphism is dated at 120–90 Ma. However, Sm–Nd and U–Pb ages of 38 Ma were ob-

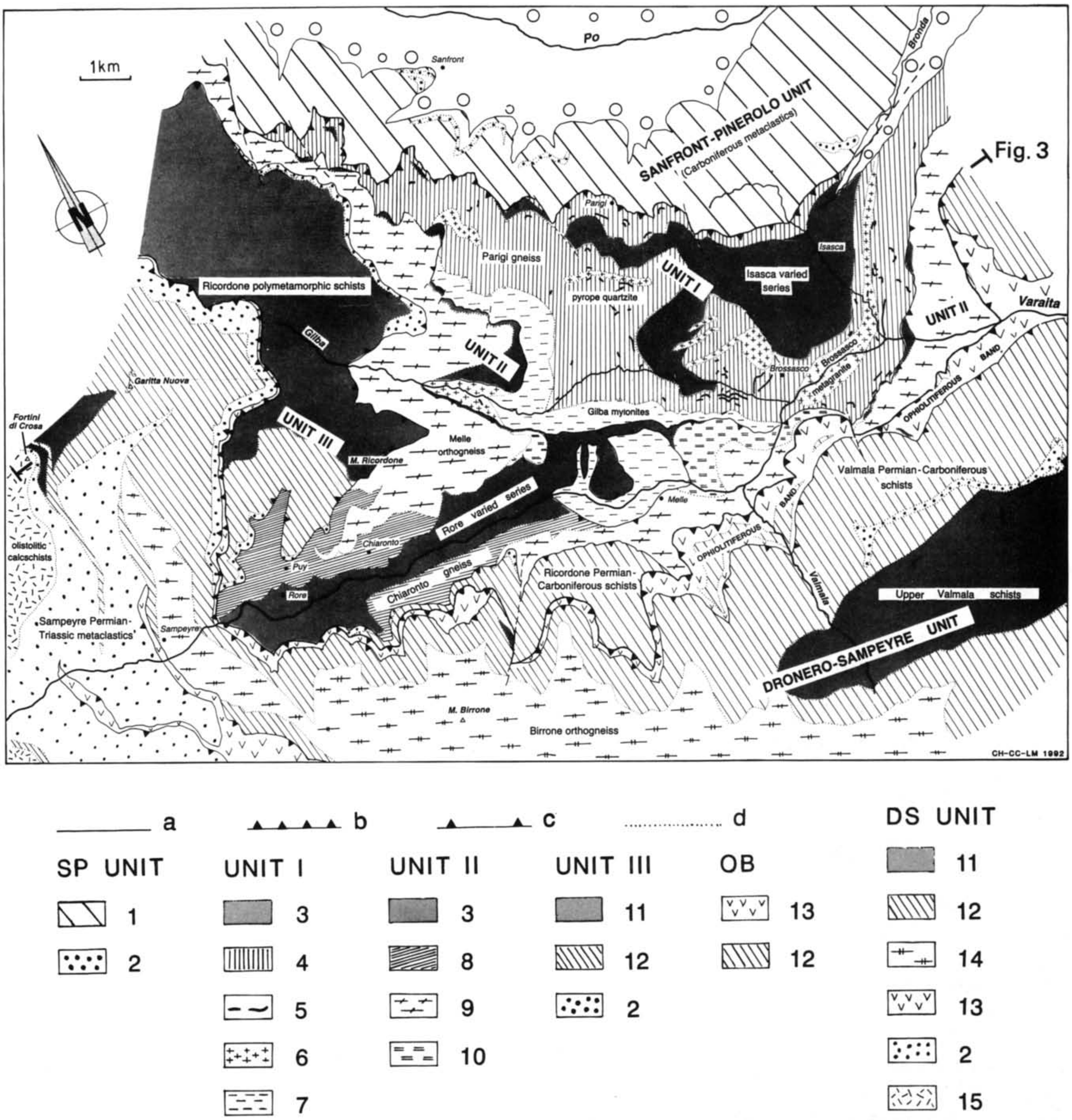

Fig. 2. Geological sketch map of the southern Dora-Maira massif. Location: Fig. 1. a—limit of the Quaternary alluvium (open circles); b—main tectonic contacts; c—subsidiary tectonic contacts; d—lithologic boundaries. 1—gneiss and Upper Carboniferous (?) graphitic schist; 2—Permian–Triassic to Middle Triassic quartzite and dolomitic marble; 3—varied series containing gneiss, metapelite, metabasite and marble; 4—Parigi fine-grained gneiss; 5—lenses of pyrope-bearing quartzite; 6—intrusive Hercynian Brossasco metagranite; 7—Gilba orthogneissic mylonite; 8—Chiaronto fine-grained, amphibole-bearing gneiss; 9—Melle orthogneiss; 10—leucocratic mylonite; 11—polymetamorphic staurolite-bearing schist; 12—Permian–Carboniferous metaclastic rocks; 13—Schistes lustrés slices (calc-schists with serpentinite lenses); 14—leucocratic Birrone Permian (?) orthogneiss; 15—olistolitic Cretaceous (?) calc-schists with mainly Permian to Triassic blocs (DS cover ?).

tained by Tilton *et al.* (1989, 1991) in the same unit and some of the same rocks. In the two overlying units, ^{40}Ar–^{39}Ar data on HP phengites suggest again a 115–90 Ma age for HP metamorphism, an age range supported by the laser fusion ^{40}Ar–^{39}Ar ages on zoned phengites in the northern extension of unit III (Pellice valley, Scaillet *et al.* 1990). The overlying Dronero-Sampeyre blueschist unit probably underwent peak metamorphism later than 90–80 Ma, if the alleged occurrence of a Late Cretaceous cover on it is correct. In the overlying ophiolitiferous nappes, the most recent data suggest a peak metamorphic age at about 50 Ma (Liewig *et al.* 1981, Monié & Philippot 1989). The age of blueschist metamorphism in the Sanfront-Pinerolo unit remains unknown. If one assumes that this unit is a mere extension of the Briançonnais basement, its peak metamorphism should be Eocene, hardly older than the regional greenschist overprint (Goffé & Chopin 1986).

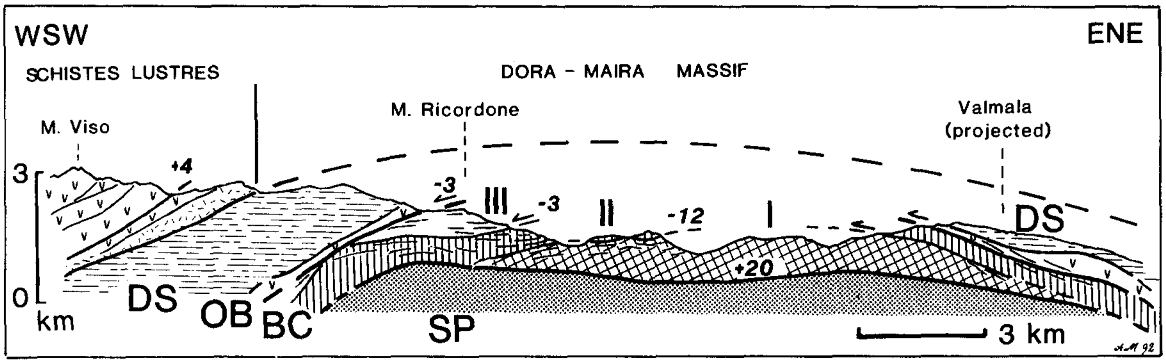

Fig. 3. Schematic cross-section of the southern Dora-Maira massif with emphasis on the metamorphic peak pressure conditions of each unit. Location: Fig. 2 (offset approximately in front of the M. Ricordone). U I—ultra-high-pressure, coesite-bearing unit; U II–III—high-pressure eclogitic units; SP, OB, DS—blueschist units, respectively, Sanfront-Pinerolo unit, Ophiolitiferous Band and Dronero-Sampeyre unit. The greenschist-facies foliation (S_m) is represented only locally. BC—Dora-Maira Basement Complex (units I, II and III). Disordered dashes—olistolitic calcschists; v-pattern—ophiolitiferous Schistes lustrés. Numbers on faults refer to pressure steps from lower to upper units, in kbar.

MAIN REGIONAL STRUCTURES: THE GREENSCHIST-FACIES OVERPRINT

The whole tectonic pile of the southern Dora-Maira massif shows a pervasive ductile fabric, irrespective of the protolith of the considered rock, its peak metamorphic conditions and its tectonic position. We will demonstrate that this main regional fabric (referred hereafter as S_m foliation and L_m stretching lineation) ultimately developed under greenschist-facies conditions (i.e. close to 3 kbar and 350–400°C, R2 on Fig. 4), with significant transposition of earlier UHP–HP minerals and structural elements. Most of the major, intrafolial folds, and the gently-dipping, low-angle ductile faults of the area formed or were reactivated during the greenschist event. High-angle shear bands and pseudotachylite bands (Henry 1990) developed later and are related to upwarping of the crystalline massif (Oligocene–Miocene time). It is convenient to begin with the major greenschist-facies structural imprint before entering the description of the UHP-HP and older structural relics.

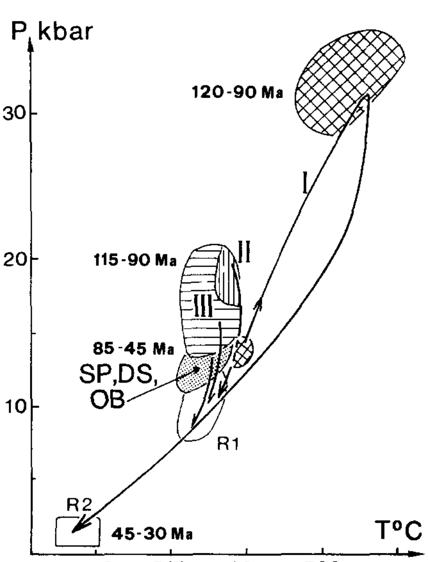

Fig. 4. Pressure–temperature paths of the southern Dora-Maira units, after Henry (1990) and Chopin *et al.* (1991). Abbreviations as on Fig. 3, with R1 (blueschist–greenschist boundary) and R2 (greenschist-facies) retrograde stages.

S_m *foliation*

In the studied area, the regional (or main) S_m foliation is a gently-dipping surface, arched in a half-dome all around the southern Dora-Maira periclinal (Figs. 5 and 6). This geometry is also conformal to lithological boundaries (compare Figs. 2 and 5), except in some fold hinges (F_{1m}, Fig. 7) at mesoscopic scale (see below, Folds).

The S_m foliation shows a domainal, mylonitic structure, and can be regarded as the regional *XY* plane of strain (Treagus 1983). In metapelitic and gneissic rock types, S_m is assimilated to the shear plane (*C*) of an *S–C* foliation (Berthé *et al.* 1979). It is transected by shear bands C', and less frequently C'' (i.e. extensional crenulation cleavages, Platt & Vissers 1980) (Fig. 7, lower detail). Its progressive development is particularly clear in the Brossasco metagranite, where all transitional stages between undeformed granite and ultra-mylonite (White 1982) are observed at the outcrop scale. Within moderately deformed orthogneiss, C' shear bands are developed (Fig. 9a). Within the most deformed mylonite, the *S–C* structure is outlined by the preferred orientation of quartz grain boundaries within quartz ribbons (Fig. 9b), and by that of mica lamellae (brown–green biotite and low-Si phengite define *S* planes, Fig. 9c) in the quartz–albite–micas microlithons and in the micaceous films (*C* planes). Chlorite mainly occurs in C' (and occasionally C'') planes. In most gneisses from all units, large relic phengite crystals are deformed along the S_m planes (Fig. 9d), forming mica fishes (Lister & Snoke 1984). The core of these pre-S_m porphyroclasts shows a high phengitic substitution ($Si_{3.5}$), decreasing toward the rim to values similar to those found in the small syn-S_m white micas ($Si_{3.2}$). This evolution indicates development of the mylonitic fabric during a pressure decrease from about 10 to about 3 kbar (for methodological aspects and limitations, see Henry 1990, Chopin *et al.* 1991).

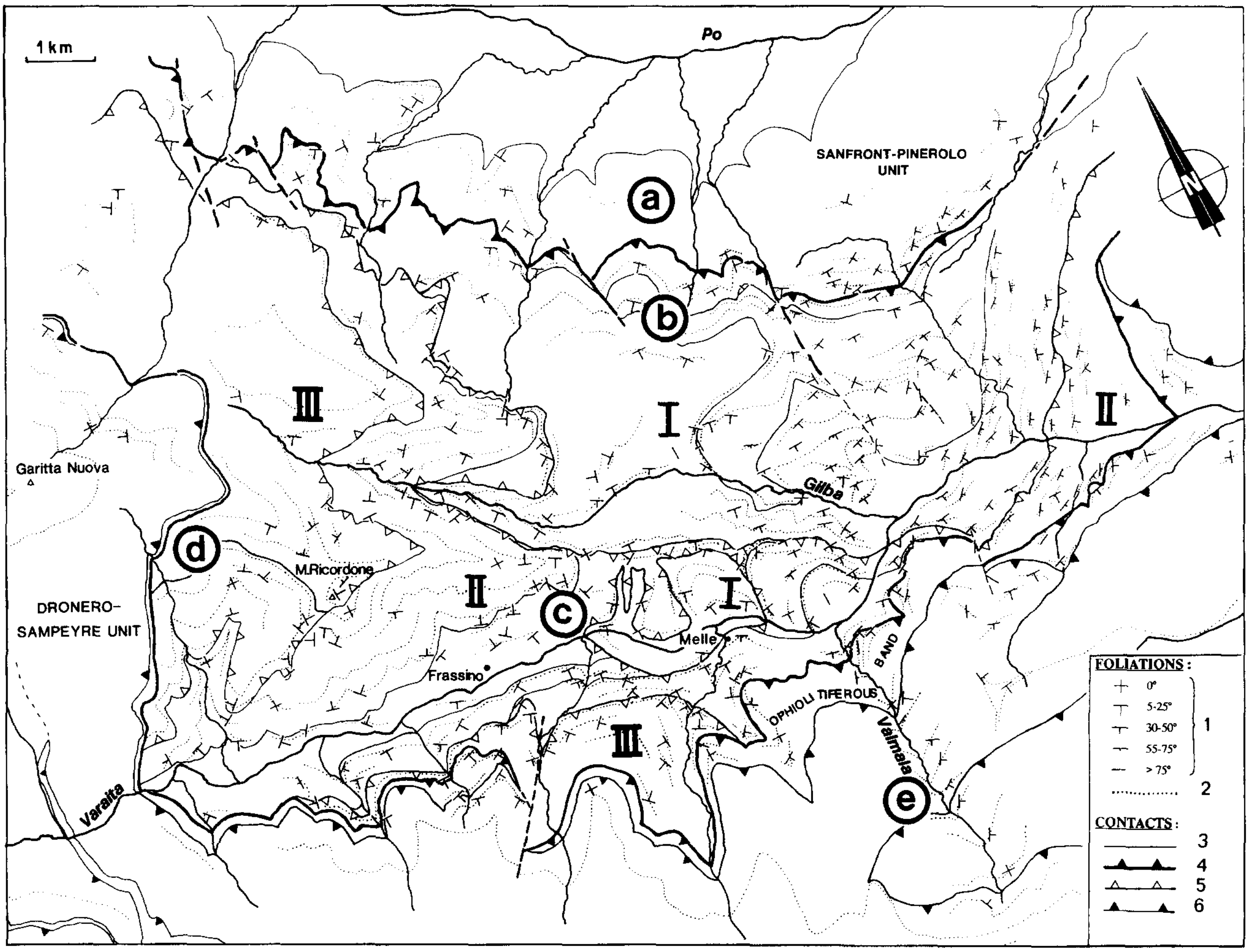

Fig. 5. Geometry of the greenschist-facies foliation (S_m) in the southern Dora-Maira massif. 1—dip of S_m; 2—trace of S_m on topography (approximated from dip and mapping of the lithologic layering); 3—pre-S_m lithologic contacts; 4—low-angle ductile faults bounding the Basement Complex; 5—as 4 but within the Basement Complex; 6—as 4 but in the overlying OB–DS units. Circled lettering—location of the stereonets of Fig. 6.

L_m *lineation*

A pervasive stretching lineation is found on S_m, which commonly coincides with a crenulation or intersection lineation. This intense stretching lineation represents the maximum elongation axis X of the deformation responsible for the greenschist foliation S_m. In most lithologies, this stretching lineation is marked by phengitic white micas, either re-oriented high-Si porphyroclasts or small low-Si grains in dynamic recrystallization tails, and by elongated quartz lenses or rods. On the S_m planes of the pyrope-bearing quartzite in the retromorphosed parts of the boudinaged outcrops, L_m is outlined by the stretching of chlorite pseudomorphs after pyrope (Fig. 9e); of phengite–margarite–albite pseudomorphs after kyanite; or else of albite–zoisite–phengite–chlorite pseudomorphs after jadeite. In all metapelitic rock types (either poly- or monometamorphic) throughout the tectonic pile, L_m is frequently marked by fibrous Fe-chlorite (e.g. within Permian silvery schists), consistent with the development of the L_m–S_m fabric in greenschist-facies conditions.

The L_m lineation shows a strong preferred orientation, close to N60–70°E throughout most of the southern Dora-Maira massif (Fig. 6). A different direction, about N30°E, is locally found (Frassino-Rore area, Fig. 6c), without one being able to distinguish it mineralogically or chronologically from the dominant E–W lineation.

Rotational strain regime and shear sense indicators

Asymmetric microstructures associated to the S_m–L_m greenschist fabric are widespread in most lithologies on XZ sections (parallel to L_m and normal to S_m). Most common are dynamic recrystallization tails on K-feldspar or phengite porphyroclasts (Fig. 9a), S–C and C–C' intersections (Figs. 9a–c) and preferred orientation of quartz grain boundaries in quartz ribbons (Fig. 9b). A strong crystallographic preferred orientation of quartz is observed in the XZ section of some mylonitic samples (uniform tint of polarization with auxiliary plate). These structures can be used as kinematic indicators (Berthé *et al.* 1979, Platt & Vissers 1980, Simpson & Schmid 1983, Lister & Snoke 1984, Passchier *et al.* 1990) and predominantly indicate a top-to-the-WSW shear direction (Fig. 8), independently observed by Philippot (1990) (except in the Frassino area, see be-

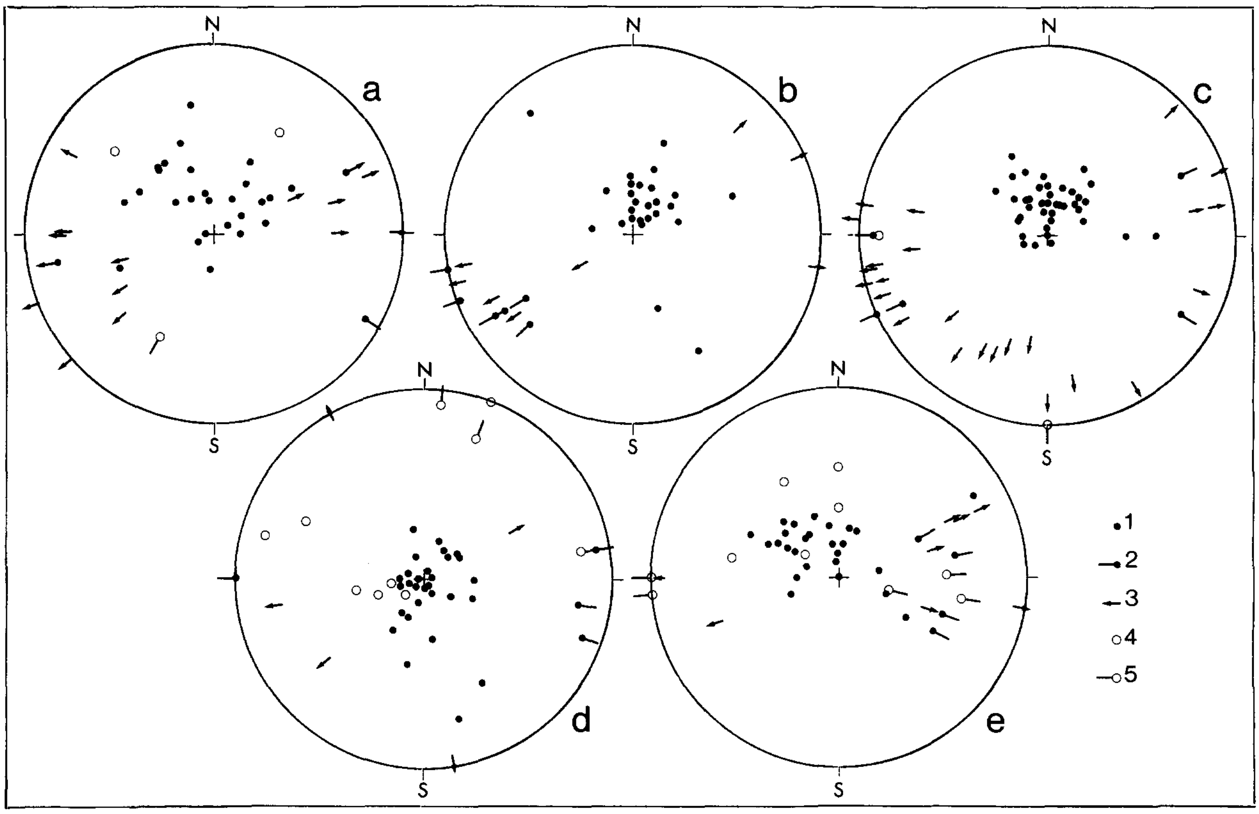

Fig. 6. Orientation of the structural elements related to the greenschist-facies evolution of some type-areas (location: Fig. 5), Wulff stereonet, lower-hemisphere. 1—S_m foliation; 2—syn-S_m (F_{1m}–F_{2m}) isoclinal fold axis; 3—L_m lineation; 4—F_3 crenulation fold (axial plane); 5—F_3 fold axis.

low). Ambiguous criteria are also found (Figs. 8 and 12b), which suggest that non-coaxial flow had locally a significant pure shear component. The finite strain ellipsoid remains poorly constrained, but the approximate strain features (oblate ellipsoids) given by Philippot (1990) are consistent with such a strain regime: no significant pressure shadows on porphyroclasts (e.g. Fig. 12a) in the *Y* direction (*XY* sections).

The intense, dominantly non-coaxial character of the greenschist deformation led us to interpret the pervasive L_m stretching lineation as the regional direction of transport (Burg *et al.* 1987, Ildefonse 1987, Schmid *et al.* 1987).

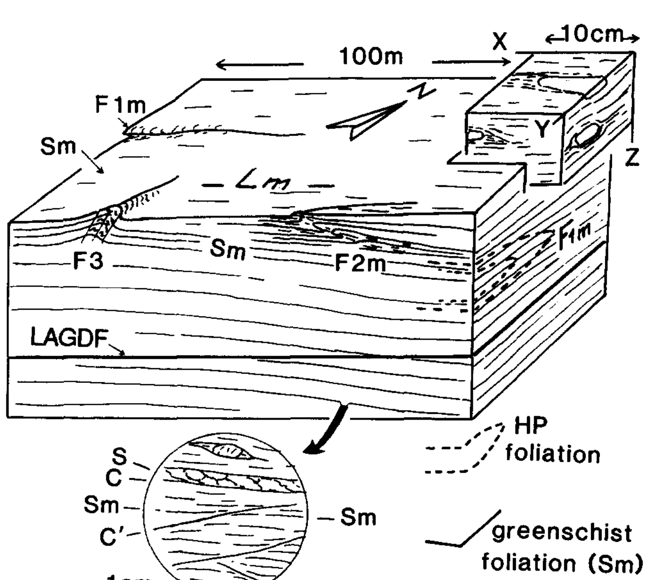

Fig. 7. Diagrammatic summary of the main regional greenschist-facies structures (see text).

A different, northeasterly sense of shear is shown in a restricted part of the studied area, where L_m tends to be oblique (N–S to N30°E, Frassino area, Fig. 8) with respect to the general regional trend of N60°E. This deformation is associated with the growth of the same greenschist minerals as the W-directed one. Ambiguous kinematic indicators are particularly frequent on the boundaries between the areas of W- and NE-directed vergence. We suggest that this local inversion in the syn-greenschist shear direction is the result of differential movements and deformation partitioning due to rheologic anisotropies (Lister & Williams 1979, Garcia-Celma 1982, Wheeler *et al.* 1987). Philippot (1990) also mentions opposite vergences of shear in the Dora-Maira complex, but partitioned on a larger scale.

F_{1m}, F_{2m} *and* F_3 *folds*

Two types of folds are commonly observed in close association with the main regional greenschist-facies fabric (F_{1m} and F_{2m}), while a last generation of folds (F_3), although still developed in greenschist-facies conditions, is characterized by different strain axes (Fig. 7). The earliest F_{1m} folds, found in the Chiaronto–Rore area, unit II (Fig. 9f), are similar folds which deform a pre-existing, HP foliation (garnet–blue amphibole–zoisite–clinozoisite–phengite–paragonite assemblage).

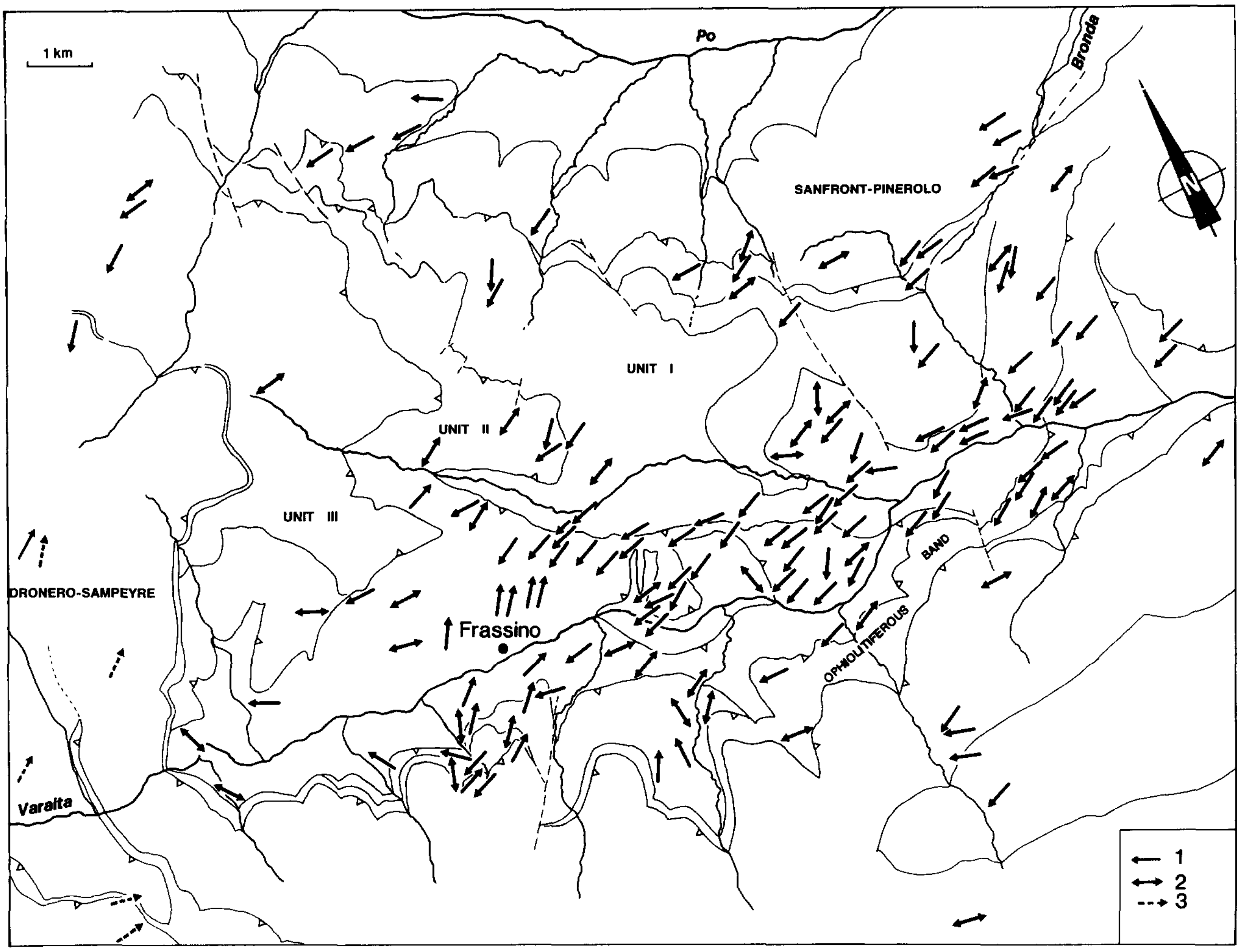

Fig. 8. Sense of shear during the regional greenschist stage, at the outcrop or sample scale. The arrows parallel the stretching lineation L_m, and point in the movement direction of the upper block. 1—unambiguous shear-sense (top-to-the-west); 2—ambiguous results; 3—data from Philippot (1990).

The S_m foliation, defined by an albite–green amphibole–chlorite–biotite assemblage, is axial-planar to these folds. In other places, the S_m foliation itself is affected by asymmetrical, conical folds, developed in otherwise planar mylonites. These F_{2m} folds show a greenschist-facies axial-planar foliation (S in Fig. 9g), and can be interpreted as drag folds developed contemporaneously with S_m. The progressive development of the greenschist-facies fabric is also demonstrated by the occurrence of superimposed folds in which large F_{2m} folds deform coaxial, isoclinal F_{1m} sheath folds (Parigi gneiss). The occurrence of sheath folds and the parallelism of most of the F_{1m}–F_{2m} axes with the coeval stretching lineation L_m (except in some areas of strong hinge curvature, e.g. Isasca marbles) characterizes areas of high shear strain (Burg *et al.* 1981, Lister & Williams 1983) during the greenschist-facies event.

The younger F_3, post-S_m folds show steeply dipping axial-planes, oblique to S_m and NNW–SSE to WNW–ESE-trending axes, oblique to L_m (Figs. 6d & e). These crenulation folds are associated with chlorite growth (Fig. 9h), and found mainly in the uppermost units of the southwestern area. When present, their asymmetry is variable, although mostly E- or N-verging. Their axial planes are locally reactivated into normal high-angle shear bands (see Philippot 1990).

Low-angle, gently-dipping faults (LAGDF)

The low-angle, generally gently-dipping ductile faults which limit the main tectonic units in the southern Dora-Maira massif were mostly mapped thanks to differences in the pre-greenschist peak metamorphism which affected the rocks below and above them (Chopin *et al.* 1991). In most cases, the S_m foliation roughly parallels these faults (Figs. 3 and 5). Since the S_m mylonitic foliation was probably subparallel to a shear plane during the greenschist-facies deformation, these faults probably operated contemporaneously as identically WSW-verging shear planes, with moderate amount of displacement along each of them. The sense of shear is confirmed locally by the occurrence of a weak obliquity between S_m and the low-angle faults, with the acute angle (S_m, LAGDF) opening westward above the fault (Figs. 3 and 5, M. Ricordone area, II–III and III–OB contacts; Valmala area, I–II and II–OB contacts).

Triple points between juxtaposed low-angle faults are seen locally (Figs. 2 and 5). The I–II and II–III faults were probably formed before the SP–BC and BC–OB faults, and could correspond to pre-greenschist faults (which delimited already lens-shaped units), reactivated during the greenschist-facies event. The SP–I–II triple point (west of Sanfront) is an example of an older fault cut by a younger one, related with the emplacement of the already juxtaposed units (I–II–III) of the Eoalpine, eclogitic Basement Complex onto the more external, and lower-grade Sanfront-Pinerolo unit. The II–III–OB–DS triple point (east of Sampeyre and south of Melle) can be related to the same event, but was probably reactivated later by high-angle normal shear bands, probably during the Oligocene–Miocene upwarping of the Dora-Maira massif (Philippot 1990).

RELIC UHP AND HP ECLOGITIC STRUCTURES

Although the greenschist regional structure attests of a widespread and intense deformation throughout the whole tectonic pile, strain partitioning however occurred, allowing the observation of Alpine, UHP–HP and even Hercynian relic structures. Relics of the structures that developed close to peak *P–T* conditions (UHP eclogite stage in unit I, HP eclogite stage in units II–III) are preserved within less overprinted rocks of the Dora-Maira Basement Complex, included in the most competent lithologies. In the coesite-bearing unit, these are the pyrope-bearing quartzite and the eclogites, both occurring in the form of boudins bounded by more ductile rocks, and also some almandine–kyanite-bearing quartz-rich metapelites. In the overlying units II–III, fabrics from the peak metamorphism are preserved in eclogitic metabasites and in some quartz-rich, chloritoid–kyanite-bearing Permian schists.

UHP relic structures (unit I): the pyrope-bearing quartzite

The pyrope–coesite-bearing quartzite is a rather heterogeneous rock. Its typical aspect in the thickest lenses (e.g. Val Gilba, Figs. 11a–c) is that of a nearly isotropic rock with almost idiomorphic, closely packed pyrope crystals, the diameter of which reaches frequently 10 cm. However, on the rim of the quartzitic lenses it can change into a white micaschist. The UHP structures are generally preserved in parts of the lenses that escaped the retrogressive mylonitization.

An UHP-eclogitic foliation (S_e), affected by the syn-S_m, F_{1m}–F_{2m} folds, is locally observed at the outcrop scale. In the lower Val Gilba, alternating beds of kyanite-bearing eclogite and white micaschist showing an early, bedding-parallel mineral orientation (flattening-stretching of quartz aggregates, garnet, pyroxene and kyanite crystals), were folded and partially recrystallized during the greenschist-facies event (Figs. 11a & b).

Typical UHP structures are also observed on the thin-section scale. The foliation that moulds the pyrope crystals (Fig. 11c) existed already during the UHP metamorphism, since the pressure shadows on the pyrope hosts include a talc–phengite–kyanite association (cf. Fig. 11g). In the quartzitic matrix of the pyrope-bearing rock, a 'palisade' fabric consisting of oriented elongated quartz grains is frequently observed (Figs. 10a & b). This unusual texture is thought to have formed by the retrogression of coesite, since the radiating pattern quartz crystals, normal to phengite–kyanite planes, is strongly comparable with the organization of quartz after coesite within the pyrope crystals (Fig. 10c). The 'palisade' fabric resembles the fibrous infilling of tension gashes and pressure shadows (e.g. Ramsay 1980). The progressive increase in spacing between micaceous layers of the 'coesitite' during the coesite–quartz transition (about 7% volume increase, Van der Molen & Van Roermund 1986) could have partly controlled the orientation of the quartz fibers (Fig. 10b), competitively with the quartz nucleation itself on the coesite–host mineral interface (Fig. 10a). If correct, the quartz–phengite–kyanite foliation involving quartz 'palisades' should proceed from a coesite–phengite–kyanite UHP foliation.

Within the pyrope megacrysts, the orientation of mineral inclusions (talc, kyanite, Mg-chlorite, ellenbergerite, rutile, zircon) is generally random, thus recording a static (non-syn-tectonic) crystallization at depth. However, an internal schistosity was observed within some of the pyrope crystals (Figs. 11d–f), and interpreted as an UHP relic fabric by Chopin (1986) and Schertl *et al.* (1991). This internal schistosity is planar or slightly curved, except in the most external part of some pyrope crystals, where its asymmetric curvature indicates the initiation of non-coaxial flow at the end of the pyrope growth. This UHP syn-crystalline deformation, involving slipping upon the shear planes, should be coeval with the formation of the cited UHP pressure shadows.

Fig. 9. The S_m–L_m greenschist-facies fabric and the associated and superimposed folds (units I, II and Dronero-Sampeyre). (a) Augen structure with C' shear bands, Brossasco orthogneiss. Polished *XZ* section. fK: K-feldspar phenocrysts moulded by alternating lenticular foliae of quartz (qz) and albite-micas (ab-m). Note the antithetic fracturing of a K-feldspar phenocryst in the middle left of the picture. (b) Brossasco ultramylonite. *XZ* section, crossed nicols, same abbreviations as (a). Note the preferred orientation of quartz grain boundaries in the quartz ribbons. White circle: location of (c). (c) Detail from (b): *S–C* structure in the microlithons of the Brossasco mylonite. Plane polarized light. qz—quartz; ab—albite; ph—phengite; bi—biotite; sph—titanite. (d) Relic phengite (phI) deformed as a mica fish, and transposed in the S_m foliation, parallel to later small phengite (phII) and biotite (bi) crystals, Parigi gneiss (unit I). *XZ* section, crossed nicols. Abbreviations: see (a), with gt—garnet; ep—epidote. (e) Retrogressed sample of the pyrope quartzite (unit I), section parallel to S_m plane; L_m is marked by elongated chloritized pyrope (pychl). (f) Tight intrafolial F_{1m} fold within fine-grained gneiss (Rore Formation, unit II). The horizontal axis of this fold trends approximately N130°. Hammer in front is 30 cm long. (g) Conical F_{2m} drag fold in the Gilba mylonites (unit I). *S*—axial planar foliation. (h) F_3 microfold in Permian micaschist (Dronero-Sampeyre unit): deformed phengite–chloritoid association (ph–ctd) in the S_m foliation, late chlorite (chl) in the hinge zones; ilm—ilmenite. Plane polarized light.

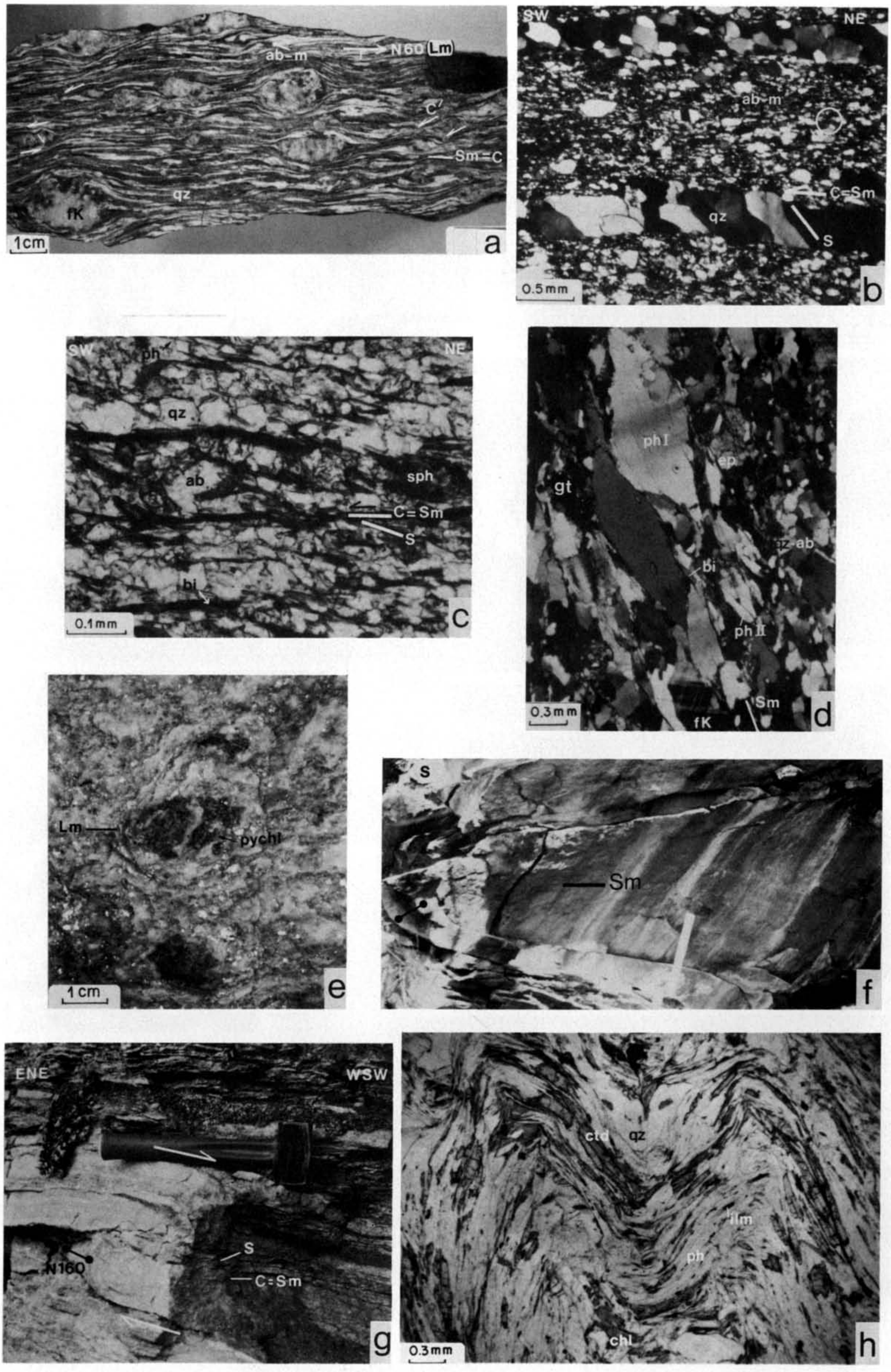
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Fig. 9.

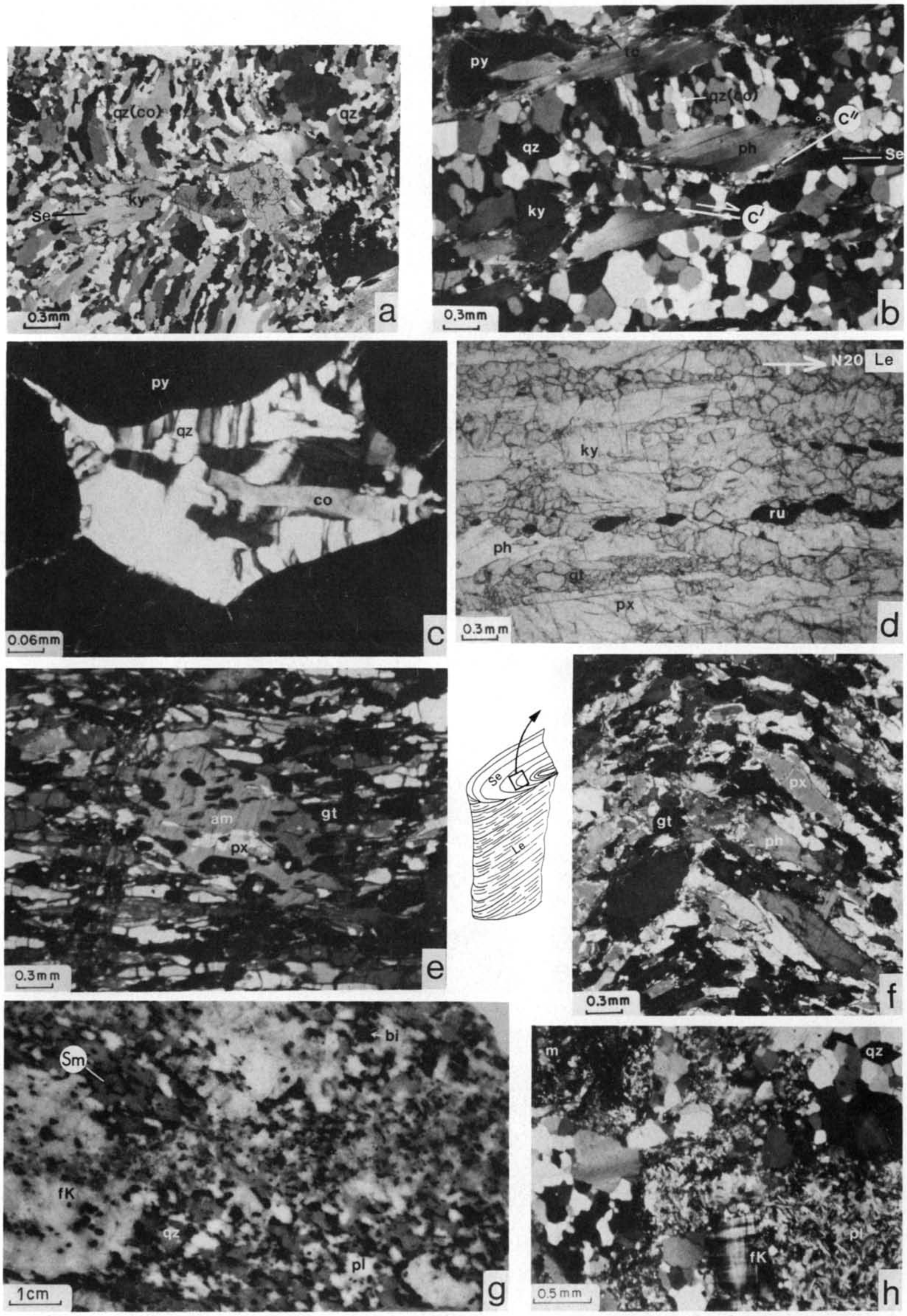
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Fig. 10.

Some pyrope crystals, although preserving their euhedral habitus with UHP talc–phengite pressure shadows, change into a 'cellular' radiating fabric made up of a HP kyanite–phlogopite–quartz assemblage (Fig. 11h) (Schertl *et al.* 1991) at the beginning of the retrogressive evolution. In the matrix around the pyrope crystals as well as in the associated jadeite–kyanite-bearing layers, close to the 'palisade' relics of the UHP foliation, the quartz texture (Fig. 10a) indicates a grain boundary migration recrystallization at high temperature and low strain rate (Hobbs *et al.* 1976, Schmid & Handy 1991), hence at the very beginning of the retrograde evolution. In the mica-rich samples, a high-Si phengite stretching lineation is found normal to the intersection of conjugate, although inequally developed, C'–C'' planes (Fig. 10b). This may denote a regime of combined flattening and shearing strain (i.e. with both pure shear and simple shear components) during the progressive deformation of these rocks at depth. These structures developed after the crystallization of the large phengite porphyroclasts (peak of UHP metamorphism), but still under high-grade, eclogitic conditions, with complete preservation of the pyrope–phengite–talc–kyanite assemblage.

Other UHP relic structures (unit I)

In the almandine–kyanite-bearing metapelites, the only relics of the UHP fabric are nearly undeformed garnet, kyanite and high-Si phengite crystals scattered in the greenschist-facies fabric (Figs. 12a–c), itself outlined by a quartz–low Si phengite–paragonite–albite association. Within almandine crystals, the inclusion trails frequently define an internal foliation (Fig. 12d). This foliation is pre- to syn-UHP, since the HP staurolite–chloritoid–paragonite assemblage (core of the garnet) grew before the UHP coesite (rim of the garnet) (see Chopin *et al.* 1991, fig. 6, where this chronology is established within a garnet showing a concentric pattern of inclusions).

The quartz–mica-rich, kyanite-bearing eclogites show an L–S or (mostly) $L > S$ fabric (Flinn 1965) that predates the greenschist-facies fabric of the neighbouring rocks. The structure is defined by large elongated kyanite and omphacite crystals, together with garnet and rutile aggregates (Fig. 10d). The constrictional finite strain ellipsoid thus revealed should record, among varied possible strain histories, one of the two following main processes; either convergent flow in a simple shear deformation or occurrence of a coaxial deformation during the UHP–HP stage.

Fine-grained (250–500 μm) basic eclogites display a plano-linear fabric ($S > L$, Flinn 1965) defined by the flattening-stretching of the eclogitic minerals (garnet, pyroxene), together with zoisite and amphibole. One notes the presence of late-eclogitic poekilitic amphiboles (Fig. 10e), the composition of which varies from blue or colourless Na-rich core to greenisch Ca-rich rim. They roughly parallel the eclogitic lineation. This could result either from mimetic growth on the primary omphacite lineation or, more probably, from syntectonic crystallization during the decompression path.

Syn-eclogitic isoclinal F_e folds were observed in both types of eclogites. The largest example, shown in the Val Gilba (3 km northwest of Brossasco), has 10 m long limbs. The omphacite–garnet ± kyanite S_e foliation and L_e lineation are folded, but the eclogitic minerals remain stable and free of internal deformation in most of the hinge zones (Fig. 10f). Folding probably occurred by shearing on the S_e plane, as shown by dispersion of the deformed lineation on a great circle of the Wulff stereonet (Hansen 1971).

Relics of HP eclogitic structures (units II and III)

In the HP eclogites (metabasites) themselves, we observe the same structures as in the basic eclogites of unit I (S_e–L_e fabric, F_e folds). In chloritoid–kyanite-bearing Permian micaschists, the quartz-rich layers present a sub-equant HP fabric. It is characterized by the random disposition of chloritoid and kyanite in two perpendicular planes. The quartz grains show an annealing polygonal texture. In other samples, the fabric is lineo-planar, with a micaceous foliation and stretching lineation involving reorientation and deformation of unaltered chloritoid and kyanite crystals. In their quartzitic matrix, syn-tectonic recrystallization at still high temperature is marked by the development of serrated grain boundaries. The presence of conjugate symmetrical shear planes suggests that flow was coaxial there during the HP deformation.

Strain regime during the UHP and the HP eclogitic metamorphism

From the structural relics described above it is difficult to restore the strain regime and kinematic axes that

Fig. 10. Eoalpine UHP and HP fabrics (a–f), and pre-alpine texture (g & h) in the coesite-bearing unit I. (a) Typical 'palisade' texture of quartz after coesite (qz(co)) and later dynamically recrystallized quartz (qz), in the matrix of the pyrope-bearing quartzite, Parigi, Val Po. Crossed nicols. S_e: trace of the eclogitic UHP foliation, also marked by kyanite grains (ky). (b) 'Palisade' texture (upper centre) and subequant fabric of quartz in the pyrope-bearing quartzite, Rocca Dritta, Val Po. *XZ* section, crossed nicols. Abbreviations: see (a), with ph—phengite; py—pyrope; tc—talc; C'–C''—conjugate, asymmetric shear-planes, associated with the eclogitic foliation (S_e). (c) Coesite–quartz inclusion in a pyrope crystal from Rocca Dritta, east of Parigi. Crossed nicols. Abbreviations: see (b). (d) Eclogitic foliation and lineation (L_e) within a quartz-rich eclogite from Rocca Ussa, right bank of lower Val Gilba. Plane polarized light. L_e is outlined by pyroxene (px) and kyanite (ky) blasts, together with rutile (ru) and garnet (gt) aggregates; ph—phengite. (e) Poekilitic late-eclogitic amphibole (am) in a metabasite from Isasca area. Crossed nicols. Abbreviations: see (d). (f) Cross-section of the hinge zone of a syn-eclogitic F_e fold in an eclogite from the Isasca formation. Crossed nicols. Abbreviations: see (d). (g) Primary magmatic texture in the Brossasco metagranite, as seen on a polished section; bi—biotite; pl—plagioclase; fK—K-feldspar; qz—quartz; S_m—incipient greenschist-facies foliation. (h) Same sample as (g), thin section, crossed nicols. Abbreviations: see (g), with m—micaceous aggregate of biotite—phengite, after magmatic biotite.

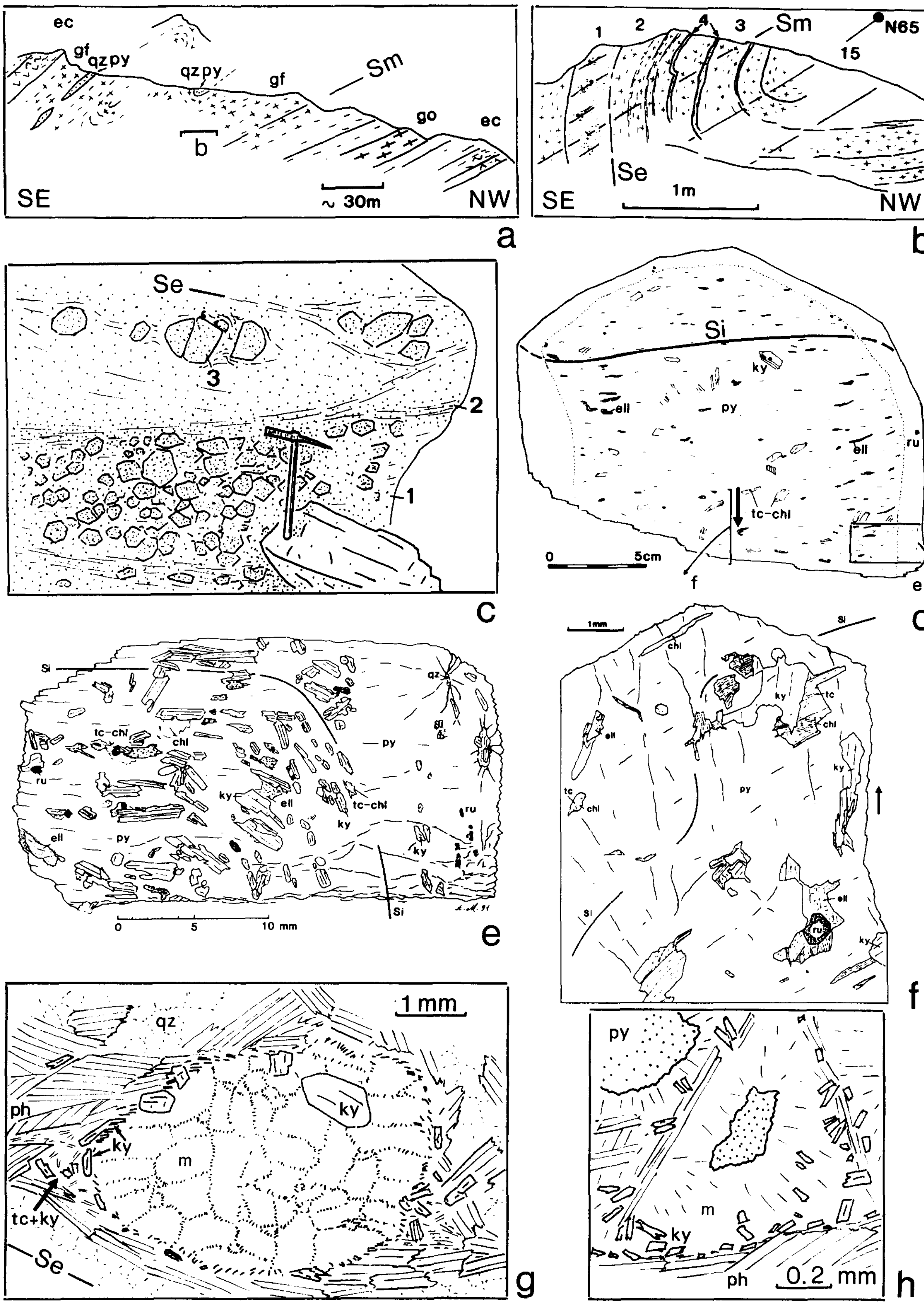

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Fig. 11.

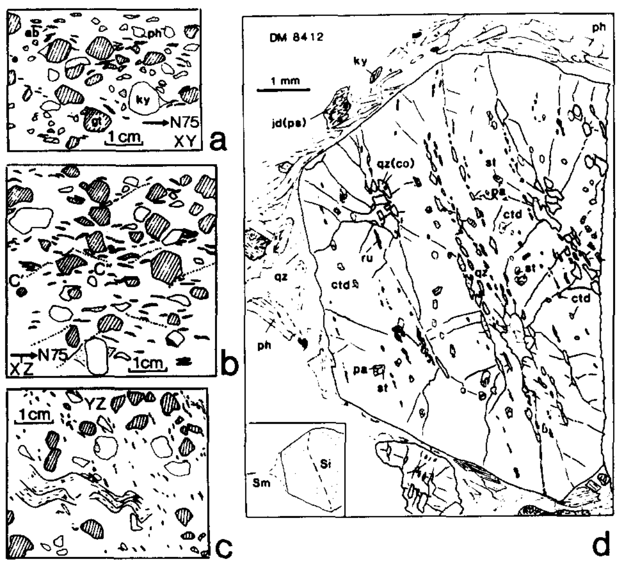

Fig. 12. Structures in an almandine–kyanite-bearing metapelite, Isasca series (unit I). (a–c) Sketch of the greenschist-facies fabric in the *XY* (a), *XZ* (b)and *YZ* (c) kinematic planes, with scattered HP-UHP blasts of almandine garnet (gt), kyanite (ky) and phengite (ph). ab—albite pseudomorph after jadeite; *C′* and *C″*—symmetrical shear planes. (d) Detail of (c): internal foliation within an almandine garnet, defined by prograde HP inclusions (see text for explanation) of chloritoid (ctd), staurolite (st), paragonite (pa) and rutile (ru) and UHP inclusion of coesite (co), replaced by quartz. S_i—prograde and peak eclogitic internal foliation; S_m—greenschist-facies foliation.

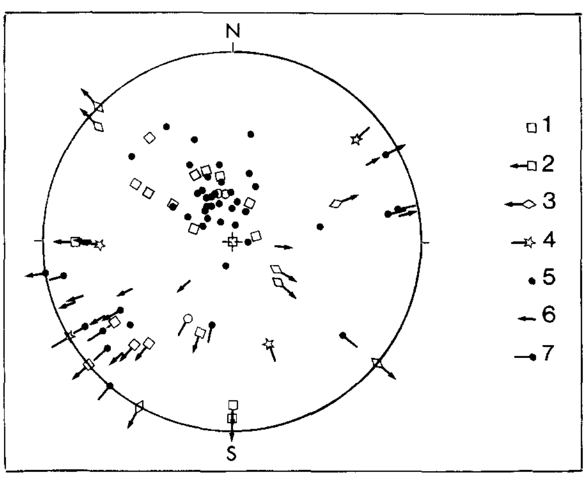

Fig. 13. Dispersion of the eclogitic structural elements in unit I. The orientation of the greenschist-facies structural elements is represented for comparison. Wulff stereonet, lower-hemisphere projection. 1—eclogitic foliation; 2—eclogitic lineation; 3—HP lineation (bluish amphibole, high-Si phengite); 4—HP fold axis; 5—S_m foliation; 6—L_m lineation: 7—syn-S_m fold axis.

prevailed around the peak metamorphism in the Basement Complex units. However, the following might be emphasized: (i) the observed geometry of the eclogitic structures are similar in the UHP unit I and in the HP units II–III; in unit I, we are also unable to recognize distinct strain regimes for the UHP-peak and the HP early retrograde stages; (ii) the eclogitic structural relics document a strongly partitioned strain. They include locally sub-equant fabrics, e.g. in the pyrope megacrysts of unit I (UHP inclusion pattern or early retrograde, cellular pseudomorphs), or in the chloritoid–kyanite-bearing micaschists of unit III. However, most eclogitic structures involve a foliation (S_e) and a mineral lineation (L_e), which define either $L > S$ or $S > L$ fabrics (Flinn 1965); (iii) UHP or HP mylonitic structures were not observed; minor shear bands are scarce and generally symmetric (C'–C''), which suggest coaxial flow in the corresponding rock bodies; and (iv) in other places, non-coaxial flow occurred, as documented on various scales by asymmetric inclusion trails within pyrope megacrysts, or by decametric syn-eclogitic folds (F_e).

It is impossible to restore the syn-eclogitic, kinematic axes. The number of L_e lineation measurements is quite restricted, partly because these lineations are difficult to distinguish from younger, greenschist-facies lineations. Where the lineation is defined by obvious UHP or HP minerals (e.g. garnet–omphacite lineation within eclogites) its orientation is variable (Fig. 13), probably due to variable bulk rotation of the eclogite boudins during late greenschist-facies deformation.

LOCAL PRESERVATION OF PRE-ALPINE TEXTURES

Additional information on the UHP strain regime can be drawn from the local preservation of pre-Alpine textures in rocks from unit I. Some primary magmatic textures are famous since they were described by Stella (1895) in the porphyritic Brossasco metagranite. In places, the granitic texture is found almost undeformed (Fig. 10g), although the magmatic minerals (except K-feldspar) have been recrystallized into new grains (Fig. 10h).

Mineral relics of a high-pressure stage are found in the metagranite: jadeite in K-feldspar sites (Biino & Compagnoni 1992), high-Si phengites in feldspar and plagioclase sites ($Si_{3.5}$: Henry 1990, Chopin *et al.* 1991). The presence of these relics implies pressures of about 15 kbar, by far less elevated than those inferred from

Fig. 11. Structures in the pyrope-bearing quartzite formation (unit I). (a) Geometry of the pyrope-bearing quartzite in a cross-section, 1 km west of Brossasco, right bank of the Gilba river. qzpy—pyrope-bearing quartzite; ec—quartz–kyanite-bearing eclogitic gneiss; gf—fine-grained gneiss; go—orthogneiss (Brossasco metagranite). (b) Detail of (a): folding of the UHP foliation (S_e) in the pyrope-bearing quartzite formation. 1—kyanite-bearing phengitic quartzite; 2 and 3—more or less quartz-rich fine-grained gneiss; 4—quartzo-phengitic layers; S_m—greenschist-facies foliation, axial-planar to the fold. (c) Close-up of another pyrope-bearing quartzite outcrop, Val Gilba area. 1—subequant quartzic zone, with closely packed and nearly undeformed pyrope megacrysts; 2—more micaceous layers; 3—boudinaged pyrope; S_e—UHP foliation. (d) Cross-section of a pyrope crystal from Parigi (Val Po). py—pyrope; chl-tc—chlorite–talc association; ell—ellenbergerite; ky—kyanite; ru—rutile. (e–f) Close-up of crystal (d) in two perpendicular sections; the trace of the UHP internal foliation (S_i) is outlined by curved inclusion trails. (g) General aspect of a HP pseudomorph after a pyrope crystal from Rocca Dritta (east of Parigi). Abbreviations: see (d), with ph—phengite; S_e—eclogitic foliation. (h) Close-up of the cellular HP retrograde fabric around a piece of relic pyrope, with kyanite neoblasts surrounding a phlogopite–microcrystalline quartz association (m) (cf. Schertl *et al.* 1991).

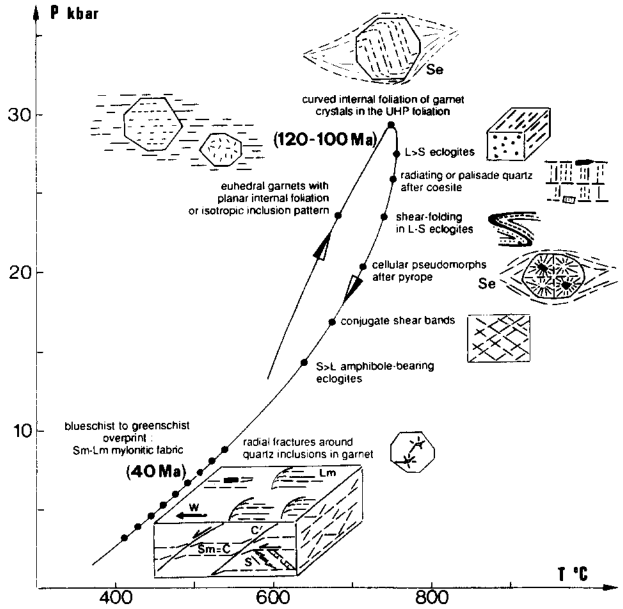

Fig. 14. Sketch of the (P–T–t) path with superimposed microstructural evolution for the Dora-Maira UHP rocks (unit I). The position of the successive fabrics on the retrograde path is approximate.

coesite-bearing parageneses. However, the available chronological and cartographic data suggest that the Brossasco metagranite shared the same deep burial history as the enclosing rocks from unit I: (i) the mapping pattern suggests an initial emplacement in the form of sills intrusive in the Isasca polymetamorphic series and in the Parigi gneiss, and the metagranite likely intrudes the Isasca–Parigi boundary (Fig. 2); (ii) mineralogical relics of contact metamorphism are locally present (Compagnoni & Hirajima 1991); and (iii) a pre-Alpine, 303 ± 1 Ma isotopic age of the Brossasco granite has been obtained by U/Pb dating of zircon (Paquette personal communication 1989).

If correct, we have to accept that primary magmatic textures can be preserved in a granitic mass, despite having been buried down to a 100 km-depth. The structurally preserved parts correspond to metre-size lenses limited by mylonitic shear zones and are located in the thickest parts of the deformed sills (Val Varaita, Brossasco; Val Po, 3 km southwest of Sanfront). The mylonites around these sub-equant bodies bear the imprints of retrograde synkinematic greenschist-facies recrystallization. We were unable to recognize any trace of an older phase of movement. The presence of undeformed metagranite bodies bounded by mylonites reinforces the concept of a strongly heterogeneous strain during the UHP metamorphism of unit I. This is consistent with the concept of incomplete eclogitization of essentially dry, crustal material at depth greater than 50 km, as documented by Austrheim (1987), Jamtveit *et al.* (1990), T. Andersen *et al.* (1991) and T. B. Andersen *et al.* (1991) in the Norwegian Caledonides.

DISCUSSION

The structural data in this paper can be used to tentatively constrain two major problems related to deep seated tectonics: (i) the kinematics in continental rocks buried at mantle depth, which may differ from that prevailing at high crustal level, and which is regarded as critical for subduction dynamics (T. B. Andersen *et al.* 1991); and (ii) the kinematics of the uplift of these rocks during their progressive unloading, i.e. the problem of the exhumation tectonics of HP–LT rocks (e.g. Platt 1986, 1987, Behrmann & Ratschbacher 1989).

Changing kinematics with depth?

Vuichard (1989) reported some features (ambiguous sense of shear, lack of sigmoidal internal foliation within garnet porphyroblasts) in the Eoalpine Sesia zone, which suggest a coaxial flow regime during the eclogitic metamorphism of this area. In their study of the UHP eclogite-bearing Western Gneiss Region (Norway), T. B. Andersen *et al.* (1991) inferred that the subducted continental crust experienced a near vertical, non-rotational constrictional flow at mantle depth, which they ascribed to slab-pull by a heavy and cold root. They note that the early HP coaxial fabrics were overprinted by extensional simple shear as the deep crust (in which the eclogitized material had already been emplaced by buoyant 'eduction') reached middle and upper crustal levels. The structures recorded in the southern Dora-Maira BC material, and particularly in the UHP unit I (Fig. 14), allow us to discuss the potential generalization of these conclusions.

We have found relics of constrictional ($L > S$) or symmetric (conjugate C'–C'' shear bands) structures formed under UHP or HP eclogitic conditions which would support the hypothesis of a dominantly irrotational flow regime within the continental rocks buried (i.e. most probably subducted) at mantle depth. However, we concurrently observed relictual eclogitic structures that attest of a rotational flow, such as curved internal foliation in pyrope megacrysts, or intrafolial syn-eclogitic shear folds. Moreover, the preservation of almost undeformed granitic lenses within the deeply buried pre-Alpine material implies that they were already bounded by shear zones during the syn-eclogitic deformation. Therefore, our data support the view of a strongly partitioned strain within the deeply buried continental crust, as already observed in the eclogitic Sesia zone by Lardeaux *et al.* (1982), and possibly as a consequence of a pre-subduction lenticular structure of the extended crust (Hamilton 1987). Our data do not confirm the concept of an exclusively coaxial flow in crustal rocks buried at mantle depth. Moreover, due to the strong and uncontrolled reorientation of the eclogitic fabric during the blueschist–greenschist evolution, checking the hypothesis of a slab-pull-related vertical strain during the eclogitic (particularly UHP) evolution is not possible.

As for the strain developed at shallower, crustal depth, under blueschist- then (and mainly) greenschist-facies conditions, we found it was still partitioned, allowing the preservation of older fabrics within almost unstrained, lenticular bodies. However, such bodies are sparse and small, and the widespread greenschists-facies overprint suggests an intensive, and almost homogeneously distributed strain during the final unloading of the Dora-Maira pile. Although ambiguous shear-sense criteria are observed, deformation is mostly characterized at that time by asymmetric fabrics at various scales (cf. Fig. 7), intrafolial shear folds (including sheath folds) and a quartz crystallographic preferred orientation, suggesting a rotational flow with a strong simple shear component. The dip of the shear planes, as restored taking into account the late upwarping of the massif, was nearly horizontal. The shear sense was mostly top-to-the-WSW, but admitted local or regional inversions. These data support the concept of a collapse of the Alpine orogenic wedge under its own weight, with a dominant, foreland-directed flow (e.g. Platt 1986, 1987, Behrmann & Ratschbacher 1989).

Structural constraints on exhumation modelling

After the discovery of coesite in the southern Dora-Maira massif (Chopin 1984), the difficulties in modelling the exhumation of Alpine eclogitic rocks were increased. Several papers successively dealt with this new geodynamic issue. Goffé & Chopin (1986) and Gillet & Goffé (1988) focused on the 'understacking' (or underplating) mechanism as a way to account for the observed *P–T–t* paths, and for the preservation of HP–LT mineral assemblages, but in their model exhumation relies on isostatic rebound and erosion. In contrast, Butler (1986) and Platt (1986) developed models in which the exhumation of the Eoalpine eclogites heavily relies on extensional tectonics. These models are based on two lines of arguments: (i) the highest presently observed erosion rates seem too small to cause the necessary exhumation rates (e.g. Draper & Bone 1981); (ii) HP–LT rocks are frequently overlain by a lower-grade tectonic 'lid' much thinner than their original overburden. The fault between the high-grade rocks and their low-grade 'lid' is seen as a gently-dipping, normal fault ('omission' of intermediate-*P* rocks), which would originate either by collapse of the orogenic wedge during continuous plate convergence (Platt's 1986 model), or by a plate divergence episode intervening between the Eoalpine and Mesoalpine constrictional events (Butler's 1986 model, inspired by the 'metamorphic core complex' models). Deville (1990) and Blake & Jayko (1990) favoured Butler's approach, while Ballèvre *et al.* (1990), Philippot (1990) and Avigad (1992) favoured Platt's. Wheeler (1991) concluded his study of the northern Dora-Maira massif suggesting that the exhumation of the Eoalpine eclogite units resulted from two successive processes: (i) buoyant extrusion of a 'pip' of eclogitized crustal rocks up to crustal levels, also admitted by Platt (1987); and (ii) collapse and extensional thinning of the orogenic wedge, with coeval foreland-directed thrusts and hinterland-directed normal faults.

The ductile faults bounding the eclogitic rock units in the Dora-Maira massif, with appropriate chronology, metamorphic *P–T* conditions and direction of movement can help to constrain these models. In the studied area, low-angle, gently-dipping (LAGD) faults do occur, that bound tectonic units of various metamorphic grades (Figs. 3 and 4). The I–II and II–III ductile faults are good candidates as normal faults responsible for 'omission' in a normally-ordered metamorphic sequence. In the case of the I–II contact, the 12 kbar *P*-gap would indicate the excision of a 40 km-thick rock section. Blake & Jayko (1990) and Avigad (1992) suggested that these faults and other similar faults in the Monte Viso–Val Po section are indeed normal faults since they are gently dipping to the west, and bear top-to-the-west kinematic indicators, and that they were responsible for the exhumation of the coesite-bearing units during the Eocene, Mesoalpine (Lepontine) event.

However, the normal vs reverse nature of the southern Dora-Maira LAGD faults cannot be determined on the ground of their present dip since, as reported above, they were arched during Oligo-Miocene time after they operated under greenschist-facies conditions, i.e. after the development of the only visible kinematic indicators (cf. Figs. 3 and 7). When restored in their syn-greenschist position, they are nearly horizontal, and the movement along the I–II fault during the greenschist-facies event could hardly be large enough to account for the excision of a 40 km-thick rock section. Moreover, the recorded *P–T–t* paths suggest that the juxtaposition of units, I, II and III within the Basement Complex (BC) was most probably achieved in HP eclogitic- to blueschist-facies conditions, prior or during their common emplacement onto the blueschist-facies Sanfront-Pinerolo unit (Fig. 4, and Chopin *et al.* 1991). The anteriority of the I–II fault with respect to the SP–BC is supported by their geometrical relationship, as discussed above. During the greenschist event, all the pre-existing faults were reactivated in the collapsing wedge, with a dominant foreland-directed flow. The SP–BC fault has probably operated in the same sense just before, as a contractional thrust (Mesoalpine emplacement of the Upper Penninic units onto the Middle Penninic). In contrast, we have no structural indication to constrain the movement along the I–II or II–III faults during the Eoalpine–Mesoalpine span of time, not even the original dip of these faults. They could have been parts of large gently-dipping normal faults (cf. Butler's model), of out-of-sequence contractional thrusts cutting through a deformed metamorphic structure (cf. Cowan *et al.* 1989, Roure *et al.* 1990), or else of steeply-dipping faults bounding buoyant slices of (partly) eclogitized continental material (cf. Ernst 1988, Wheeler 1991). Speculation on this point, i.e. on the early (and main) stages of the exhumation of the UHP and HP eclogite units, is beyond the scope of this paper.

CONCLUSION

The structural study of the southern Dora-Maira massif, in the area where coesite was discovered, leads to the following conclusions. The observed structures reflect a strong strain partitioning at UHP and HP conditions as well as during a late greenschist-facies deformation event. This partitioning allowed the preservation of weakly or even non-deformed rock bodies (e.g. Brossasco metagranite). The (indirect) evidence of the existence of UHP–HP shear zones bounding already lens-shaped units, gives us a picture of a deeply buried deep crustal segment. Because of the sparseness of UHP and HP structural relics, it proved impossible to recognize distinct flow regimes at UHP and HP conditions. Moreover, the corresponding kinematic axes remain unknown, due to the uncontrolled rotation of these relics; therefore, the early, Mid-Late Cretaceous stages of the exhumation processes are not constrained by precise structural data. On the contrary, the late, widespread greenschist-facies event is characterized by a well expressed, basically W-verging non-coaxial flow that can be referred to the Eocene collapse of the alpine orogenic wedge. During this event, the low-angle, gently dipping faults which limit the lens shaped units (particularly I–II and II–III contacts within the BC) are reactivated as mylonitic shear planes with moderate amount of displacement. The Eoalpine–Mesoalpine throw and sense of movement on these faults (normal vs reverse), critical for exhumation modelling, remains uncertain.

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