Crust-mantle interactions during subduction of oceanic & continental crust. 10th International Eclogite Conference, Courmayeur (Aosta, Italy) - Post-conference excursions: September 9-10, 2013

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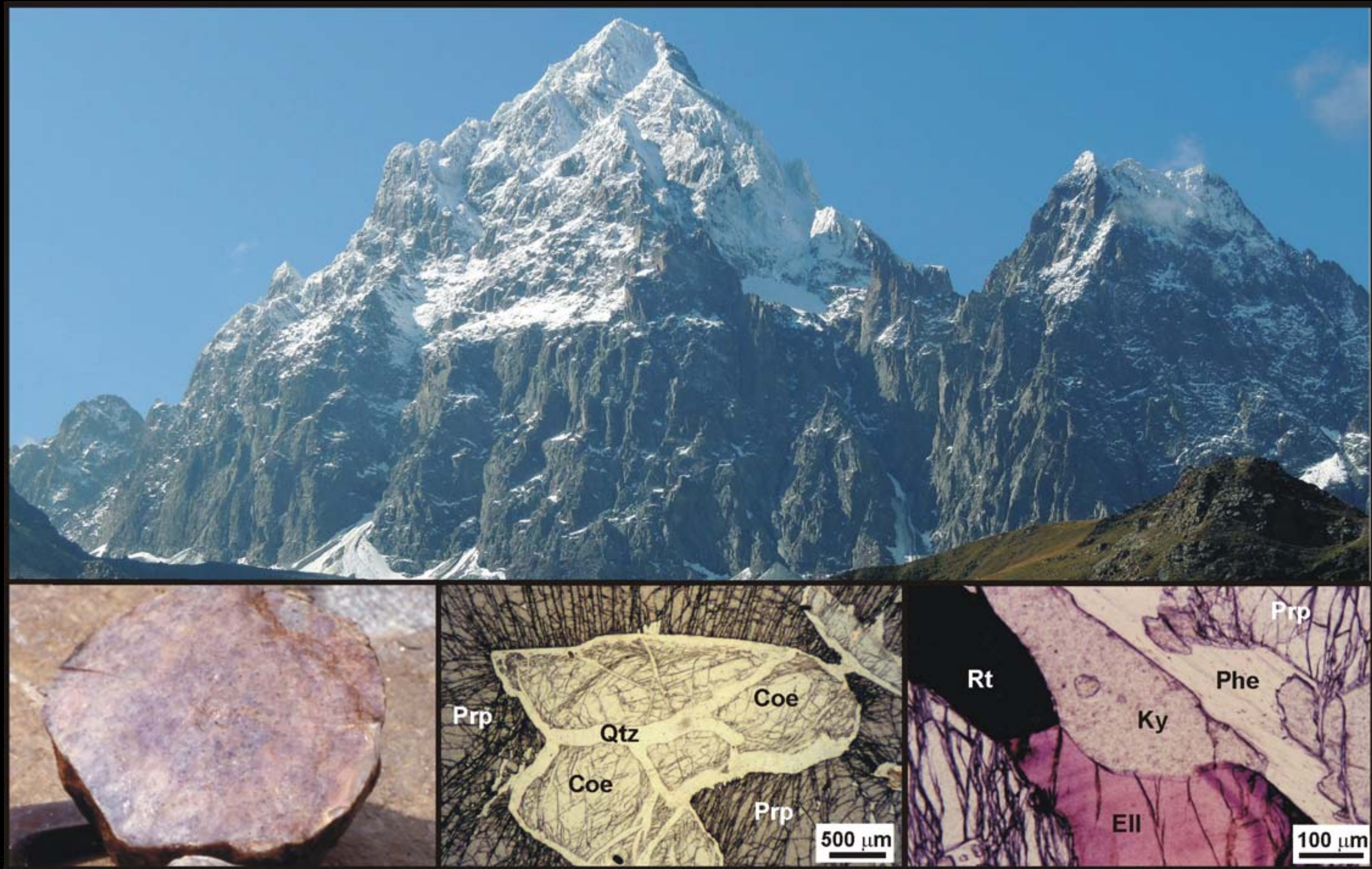
Geological Field Trips



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Periodico semestrale del Servizio Geologico d'Italia - ISPRA e della Società Geologica Italiana

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Day 1 - The Monviso meta-ophiolite Complex: HP metamorphism of oceanic crust & interactions with ultramafics

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Day 2 - Metasomatism from & to ultramafics: the UHP continental Brossasco-Isasca Unit

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Cover image – *Top*: Monviso (3841 m) and Visolotto (3348 m) as seen from Stop 1.1. *Bottom*: Pyrope megablast, ca. 20 cm across (left); relics of coesite included in a pyrope porphyroblast and partially converted to quartz along rims and fractures (centre); zoned ellenbergerite in association with kyanite, phengite, and rutile armoured in a pyrope megablast (right). Modified from Compagnoni et al. (2004).

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Introduction

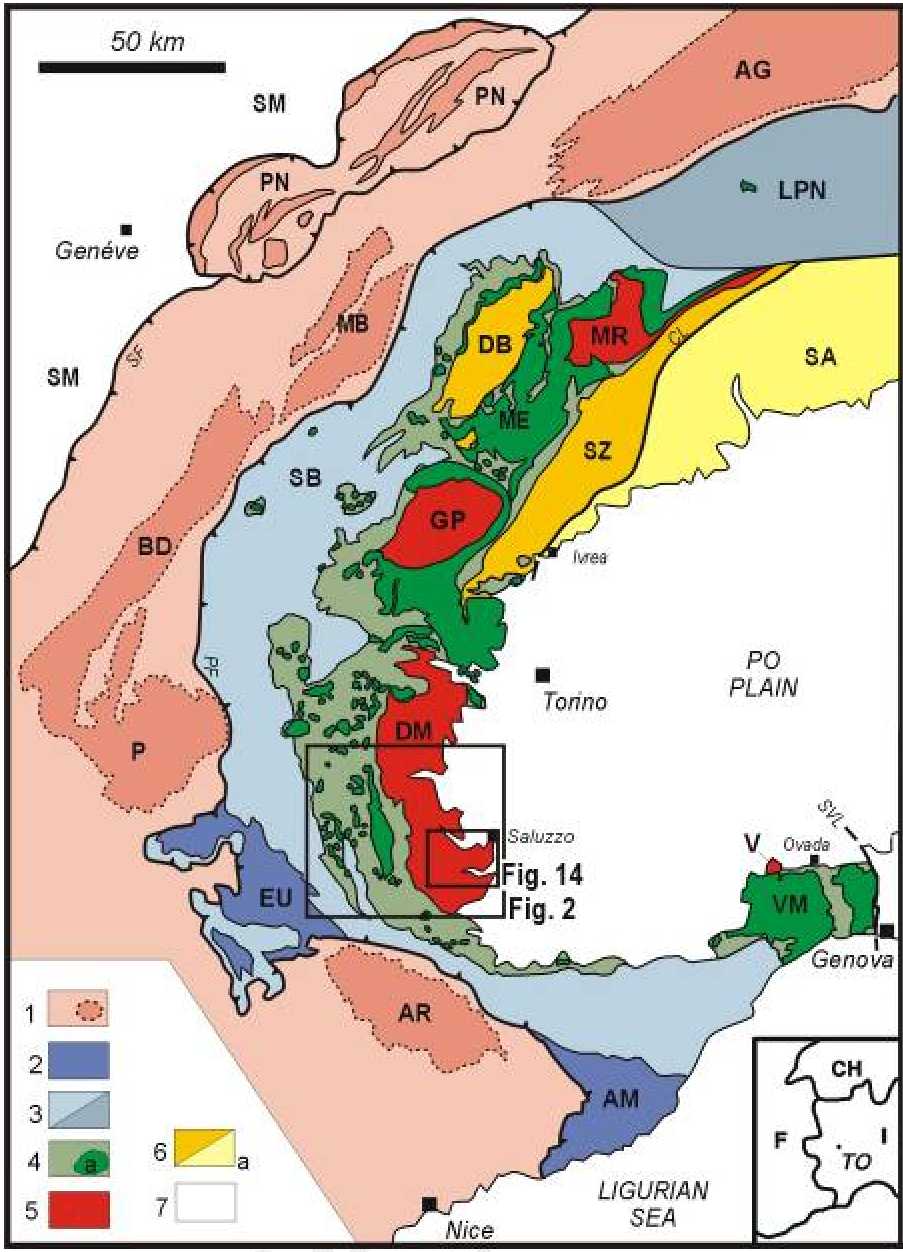
Tectonic outline of the Western Alps

The main tectonic and palaeogeographic domains of the Alps are: (i) the Helvetic-Dauphinois domain, (ii) the Penninic domain, (iii) the Austroalpine domain, and (iv) the Southalpine domain or Southern Alps (Fig. 1). Unlike these domains, where both basement and cover rocks are exposed, the Jura chain (separated from the main Alpine chain by the Swiss molasse basin) only consists of deformed cover sequences (Fig. 1).
In the western Alps, the Penninic domain is extensively exposed, while in central and eastern Alps it is hidden by the nappes of the Austroalpine domain and reappears only in the two tectonic windows of Engadine and Hohe Tauern.

The Helvetic-Dauphinois domain, which is subdivided in a series of nappes, consists of a Variscan (or older) crystalline basement and a post-Variscan, mainly Mesozoic sedimentary cover (Fig. 1). The crystalline basement, which consists of Variscan, or older, metamorphics intruded by post-Variscan mainly Permian (ca. 270 to 300 Ma) granitoids, is exposed in the so called "External Crystalline Massifs", which are (from N to S): Mont Blanc-Aiguilles Rouges, Belledonne, Pelvoux and Argentera.

The Penninic domain is a composite realm, which includes both continent- and ocean-derived units (Fig. 1). The continental units are the external Briançonnais Zone (or Grand Saint Bernard composite nappe), and the "Internal Crystalline Massifs" (ICM) of Monte Rosa, Gran Paradiso, Dora-Maira and Valosio. The ICM are overlain by the ocean-derived units of the Piemonte Zone.

(i) The continental Briançonnais - Grand Saint Bernard zone consists of a Variscan crystalline basement, a Permo-Carboniferous sequence and a Mesozoic to Eocene cover.
(ii) The ICM are tectonic windows within the Piemonte zone. They mainly consist of a Variscan amphibolite-facies metamorphic basement, intruded by Permian porphyritic granitoids, converted to augen-gneisses by the Alpine tectonics and metamorphisms.
(iii) The ocean-derived Piemonte Zone, also known as "Zone of Calcschists with meta-ophiolites" includes fragments of both the Mesozoic Ligurian-Piemontese oceanic lithosphere and its Mesozoic sedimentary cover, consisting of “calcescisti” (calcschists, or "schistes lustrés" in French) (see e.g.: Deville et al., 1992; Dal Piaz, 1999).



Simplified structural sketch map of the Western Alps.
1: Helvetic Domain and Prealpine decollement nappe system (PN). The dashed line contours the External Crystalline Massifs of the Helvetic Domain (AR: Argentera, P: Pelvoux, BD: Belledonne, MB: Mont Blanc - Aiguilles-Rouges, AG: Aar-Gotthard).
2: Helminthoid Flysch nappes (EU: Embrunais-Ubaye, AM: Maritime Alps, PN).
3: External Penninic Domain (SB: Grand St. Bernard Zone, LPN: lower Penninic nappes).
4: Piemonte Zone (VM: Voltri Massif); (4a) main meta-ophiolite bodies.
5: Internal Crystalline Massifs of the Penninic Domain (MR: Monte Rosa, GP: Gran Paradiso, DM: Dora-Maira, V: Valosio).
6: Austroalpine Domain (DB: Dent Blanche nappe, ME: Monte Emilius nappe, SZ: Sesia Zone); (6a) Southalpine Domain (SA: Southern Alps).
7: Swiss Molasse (SM), Po Plain Molasse and Piemonte-Liguria Tertiary basin.
CL: Canavese line; *SVL:* Sestri-Voltaggio line, *SF:* Subalpine frontal thrust, *PF:* Penninic frontal thrust. In the inset, the borders among France (F), Italy (I) and Switzerland (CH) are shown.

Fig. 1 – Simplified structural sketch map of the western Alps with location of tectonic maps reported in Figs. 2 and 14.

To SE of the Canavese line, the Southalpine Domain is exposed (Fig. 1), which was not involved in the Alpine orogeny. It includes the Ivrea Zone, characterized by pre-Alpine granulite- to amphibolite-facies metamorphics, and the Serie dei Laghi, which is composed of an amphibolite-facies basement, intruded by Permian granites ("Graniti dei Laghi") and overlain by Permian rhyolites and a Mesozoic sedimentary cover.

The Piemonte Zone is overlain by the Austroalpine Sesia Zone and Dent Blanche nappe system, which includes the main thrust sheets of Dent Blanche *s.s.*, Mont Mary, Monte Emilius and several minor outliers.
From a simplified NW-SE tectonic cross-section from the Pre-Alps to the Southern Alps through the Dent Blanche Nappe (with the top of the Matternhorn) and the Sesia Zone, it is evident that the orogenic wedge becomes progressively thicker going from the Jura to the Canavese Line and, consequently, the Alpine metamorphic grade is increasing from the external to the internal side of the western Alps.
This two-day excursion is devoted to show part of a cross-section within the Penninic domain west of Saluzzo (Fig. 1).

High to ultra-high pressure metamorphic rocks in the western Alps
In the western Alps, both pre-Alpine and Alpine eclogites occur, the former being much rarer than the latter.
The pre-Alpine eclogites were reported from the polymetamorphic basement of the External Crystalline Massifs of Mont Blanc, Aiguilles Rouges, Belledonne and Argentera, from the Briançonnais/Grand Saint Bernard Zone, and from the Serie dei Laghi. It is not yet conclusively ascertained if the lack of pre-Alpine eclogites in the internal Penninic Monte Rosa, Gran Paradiso and Dora-Maira massifs is a primary feature or if it is the consequence of the intense and pervasive Alpine polyphase recrystallization. The last interpretation could be supported by the local occurrence in the polymetamorphic basement (e.g., in the Gran Paradiso: Compagnoni & Lombardo, 1974) of meta-basics, where the presence of rounded aggregates of small euhedral Alpine eclogitic garnets suggest recrystallization after coarser-grained garnet of pre-Alpine age.
Two main Alpine metamorphic events were recognized all over the Penninic and Austroalpine domains of the western Alps: an older high- to ultrahigh-pressure (HP, UHP) metamorphism and a younger low pressure (LP) metamorphism, which overprints and extensively obliterates the older one.
The areal extent and zoning of the HP to UHP metamorphism is difficult to define at large scale, because it developed during subduction, long before the final nappe emplacement. Therefore, the once considered isograds of the HP metamorphism are really tectonic contacts between nappes, which experienced different

peak metamorphic conditions. On the contrary, the LP metamorphism postdates the main nappe emplacement and its isograds cut across the tectonic contacts between nappes.
All facies typical of the HP to UHP metamorphism are observed, i.e. coesite-bearing eclogite-, quartz-bearing eclogite-, epidote blueschist- and lawsonite blueschist-facies.
Coesite-eclogite facies rocks are reported from only two localities: the continental crust of the Penninic Brossasco-Isasca Unit, southern Dora-Maira Massif (Chopin, 1984) and the sedimentary cover of the metamorphic ophiolites of the Piemonte Zermatt-Saas Zone at Lago di Cignana, Upper Valtournenche, Val d'Aosta (Reinecke, 1989, 1991, 1998). As to the Brossasco-Isasca Unit, recent petrological data from both experiments (Hermann, 2003) and natural samples (Castelli et al., 2007; Groppo et al., 2007b) suggest that diamond-eclogite facies conditions were attained at about 750°C and $4.0<P<4.3$ GPa.
Quartz-eclogite facies rocks are the most widespread. They occur in most of the Sesia Zone and Monte Emilius of the Dent Blanche nappe system, in most of the internal Piemonte Zone, and in the Internal Crystalline Massifs of Monte Rosa, Gran Paradiso, Dora-Maira and Valosio.
Epidote blueschist- and lawsonite blueschist- facies rocks are extensively exposed in the external Piemonte Zone and in the Briançonnais Zone. In this area the carpholite-bearing assemblages are frequently found in rocks of appropriate bulk chemical composition (Goffé, 1984).

Day 1 - The Monviso Meta-ophiolite Complex: HP metamorphism of oceanic crust & interactions with ultramafics

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The Monviso meta-ophiolite Complex

The Monviso Meta-ophiolite Complex forms a NS-trending body, 35 km long and up to 8 km wide, structurally sandwiched between the underlying Dora-Maira thrust units and the other, dominantly metasedimentary, units of the ocean-derived Piemonte Zone (Fig. 2). In the western Alps, it is one of the best preserved relics of the oceanic crust formed during opening of the Mesozoic western Alpine Tethys that underwent eclogite-facies metamorphism during Alpine subduction (e.g. Lombardo et al., 1978; MONVISO, 1980; Philippot, 1988; Messiga et al., 1999; Schwartz et al., 2000).

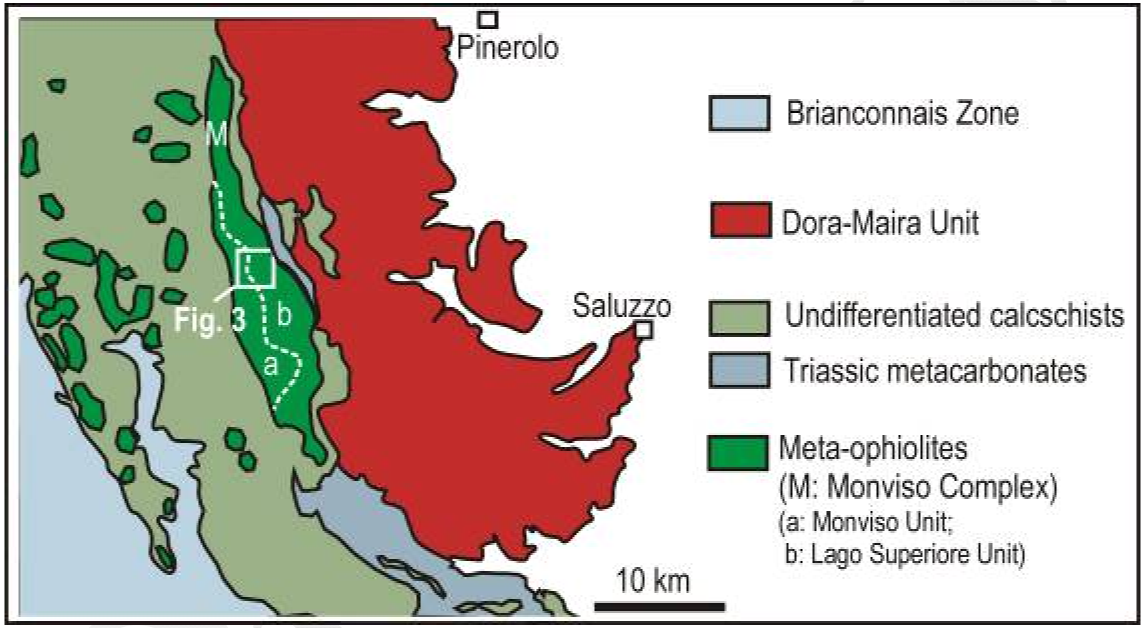


Fig. 2 - The Monviso Meta-ophiolite Complex in the frame of the Penninic domain of southwestern Cottian Alps, with location of the area visited during the field excursion (Fig. 3).

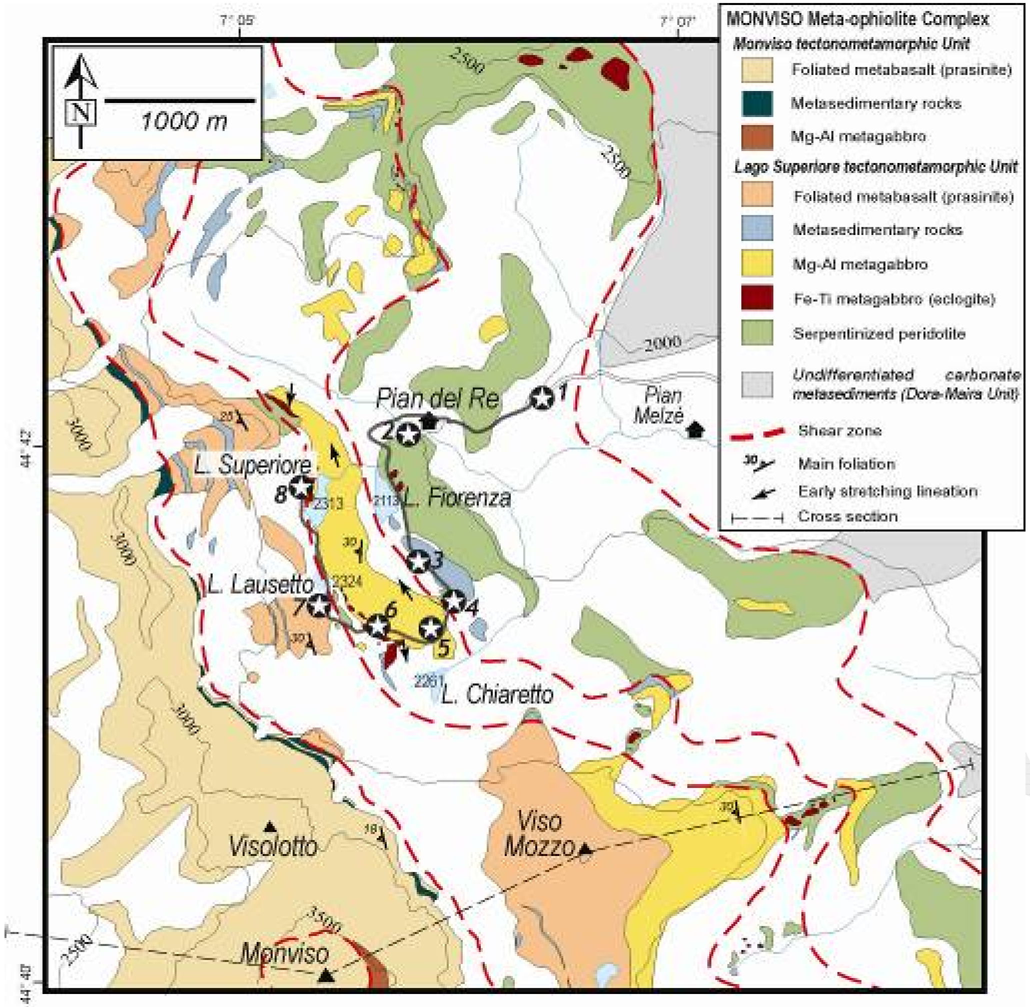
Tectonometamorphic and tectonostratigraphic units

The Monviso Meta-ophiolite Complex encompasses the whole lithological spectrum of the Piemonte-Liguria ophiolite rocks and comprises two major tectonometamorphic units, distinguished on the basis of Alpine H*P* peak metamorphic conditions (Fig. 3): the "Monviso Unit" to the west and the "Lago Superiore Unit" to the east (Angiboust et al., 2012a).

Each tectonometamorphic unit is in turn subdivided into tectonostratigraphic units. The earlier tectonostratigraphy of the Monviso Meta-ophiolite Complex (Lombardo et al., 1978) has been renewed (Lombardo et al., 2002; Balestro et al., submitted) and, from west to east and from top to bottom, the following tectonostratigraphic units have been distinguished in the field excursion area (i.e. the Monviso - Pian Melzè area; Fig. 3):

(i) The Vallanta Unit: it is an overturned slab of largely re-equilibrated, eclogite-facies metabasalts, with small relics of a metasedimentary cover of carbonate-bearing micaschists. This unit is topographically and tectonically the highest one and caps the very top of Monviso. It is separated from the underlying Forciolline Unit by a narrow shear zone wherein sheared antigorite serpentinites occur.

(ii) The Forciolline Unit: it is an overturned sequence made up of metagabbros, massive and pillowed metabasalts, banded glaucophane-epidote metabasites derived from pillowed basalts and basalt breccias, and subordinate metasediments (Fig. 4). The Forciolline metagabbro is isotropic and coarse-grained, shows generally well-preserved igneous textures, and includes minor massive and layered troctolite (Compagnoni & Fiora, 1976). The rafts of serpentinized peridotite that are preserved in the metagabbro, suggest that it was emplaced into the oceanic mantle (Lombardo & Pognante, 1982). Petrographical and major-element data indicate derivation from olivine gabbro with TiO_2 of 0.3-0.4 wt% and Mg# [= 100xMg/(Mg+Fe_{tot})] ranging between 82 and 77 (Piccardo & Fiora in Lombardo et al., 1978). The magmatic clinopyroxene is a chromian diopside, with Mg# = 79-82. Massive and pillowed metabasalts stratigraphically overlie the metagabbros. This igneous association also includes some metabasalt dykes that cut both the metagabbros and the metabasalts (Fig. 4). The metabasalts are often Pl-phyric, and more rarely Ol-Pl-phyric. Their Mg# range between 65 and 51, and TiO_2 is between 1.2 and 1.8 wt% (Piccardo & Fiora in Lombardo et al., 1978). The metasediments that cover the metabasites consist of calcschists and manganous Qtz-rich micaschist (Lombardo et al., 1978). The latter is 2-5 m thick and can be followed almost continuously at the bottom of the Unit.



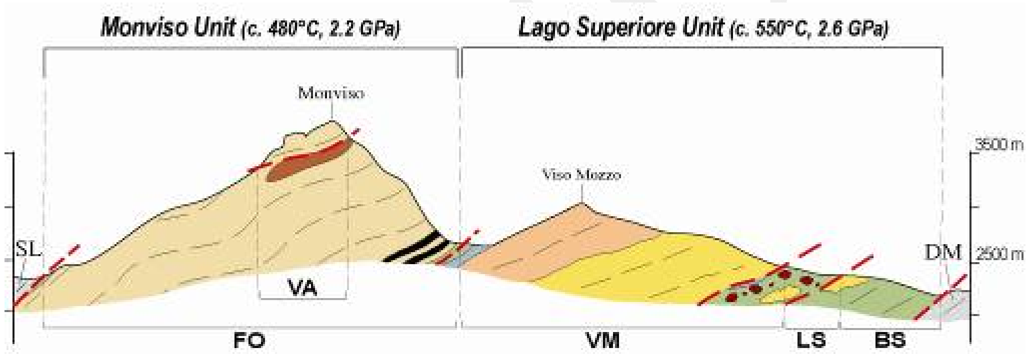


Fig. 3 - Geological map and cross section of the Monviso Meta-ophiolite Complex in the Monviso – Pian Melzè area, with the excursion route and location of Stops 1 to 8 (in *italics*) described in the text (modified from Angiboust et al., 2012a). The two main tectonometamorphic units (and associated peak *P-T* estimates from Angiboust et al., 2012a) and the related tectonostratigraphic units (Lombardo et al., 2002; Balestro et al., submitted) constituting the Monviso Meta-ophiolite Complex, are respectively shown above and below the cross section (SL: Schistes Lustrès; VA: Vallanta Unit; FO: Forciolline Unit; VM: Viso Mozzo Unit; LS: Lago Superiore Unit; BS: Basal Serpentinite Unit; DM: Dora-Maira).

(iii) The Viso Mozzo Unit: it mainly consists of banded glaucophane-epidote metabasites, in places with textural relics of pillow lavas and breccias (e.g. south of Viso Mozzo) that highlight their basaltic parentage. At the top, the metabasites are interbedded with m-sized layers of calcschists and subordinate Qtz-rich micaschists. These metasediments further crop out north of the Monviso-Pian Melzè area (Balestro et al., 2011) and represent the original cover of the basalt-derived metabasites. The omphacite-bearing Mg-Al metagabbros that crop out E of Viso Mozzo (previously included in the underlying Lago Superiore Unit; Lombardo et al., 1978), seem to have primary relationships with the metabasites and were therefore newly referred to the lower part of the Viso Mozzo Unit (Balestro et al., submitted) (Fig. 3). The contact with the overlying Forciolline Unit consists of a major shear zone (previously included in the Passo Gallarino Unit of Lombardo et al., 1978), made up of serpentinized peridotites and antigorite schists that host bodies of metagabbro. The latter mainly consists of a 100 m thick slab of interlayered Fe-Ti gabbro-derived eclogite and omphacite-bearing Mg-Al metagabbro, with minor metagabbronorite and metaplagiogranite (Lombardo et al., 1978; MONVISO, 1980; Fig. 3). The eclogite derives from a Fe-Ti gabbro protolith, with Mg# ranging between 39 and 32, and TiO_2 ranging between 4.0 and 5.0 wt%, whereas the metagabbronorite has Mg# of *c.* 70 with TiO_2 contents of *c.* 1.0 wt%. Other minor bodies of eclogite occur on the southwestern slope of Viso Mozzo (Schwartz et al., 2000) and have a composition (Mg# = 39-34, TiO_2 = 1.2-1.6 wt% and Al_2O_3 = 18.0-19.6 wt%) that suggests a Fe-oxide leucogabbro protolith.

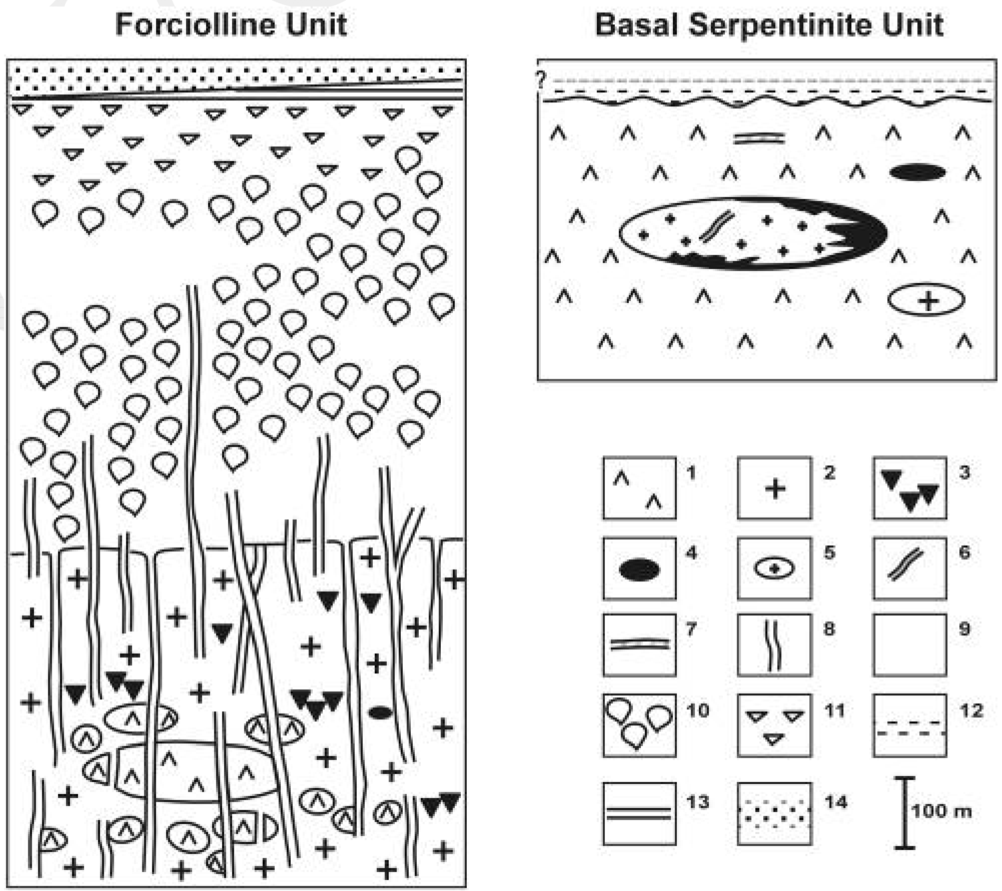


Fig. 4 - Schematic primary relationships of lithologies that make up the Forciolline Unit (left) and the southern Basal Serpentinite Unit (up right), representative of magma-rich and magma-poor units of the Monviso Meta-ophiolite Complex, respectively (Castelli & Lombardo, 2007). 1) Tectonite mantle peridotite; 2) gabbro body and gabbro dyke; 3) ultramafic and troctolitic cumulate; 4) Fe-Ti gabbro; 5) plagiogranite body; 6) plagiogranite dyke; 7) albitite; 8) basalt dyke; 9) massive basalt; 10) pillowed basalt; 11) basalt breccia; 12) ophicalcite; 13) metachert; 14) calcschist.

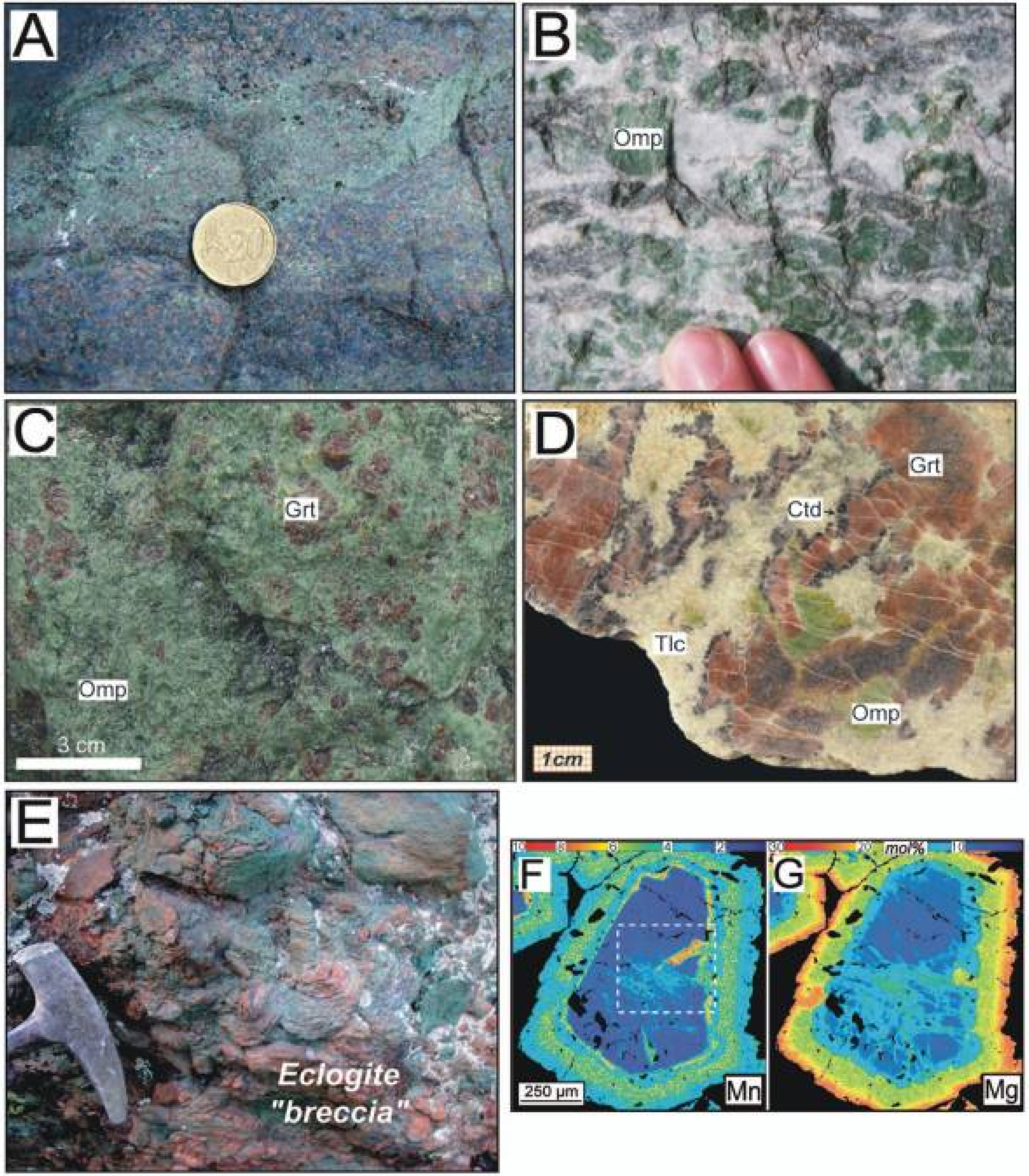


Fig. 5 – (A) Photograph of a Fe-Ti metagabbro from Passo Gallarino area showing variable degree of crystallization of dark blue amphibole. (B) Photograph showing the typical texture of Mg-Al metagabbro from Lago Superiore area. (C) Photograph of an eclogite sample, mainly consisting of garnet and omphacite, derived from metasomatic alteration of a Fe-Ti gabbro. (D) Photograph of a hand specimen of metatroctolite from the Mg-Al metagabbroic body (Lago Superiore area; Messiga et al., 1999) showing the coronitic formation of chloritoid (Ctd) between garnet (Grt) and talc (Tlc). (E) Photograph showing the typical texture of the eclogite breccia described in the shear zone at the base of the Mg-Al gabbro. (F, G) X-ray compositional mapping of an eclogite breccia sample showing multiple fracturing/healing events fossilized by garnet rims and mantles. Pictures E, F and G are from Angiboust et al., 2012b and picture D is from Angiboust et al., 2012a.

(iv) The Lago Superiore Unit: it is a discontinuous belt of serpentinized peridotites and of strongly deformed and recrystallized metagabbros that are separated from the overlying Viso Mozzo Unit by a narrow shear zone mainly made up of antigorite schists. The metagabbros are characterized by the presence of emerald-green chromian omphacite ("smaragdite"; Fig. 5B), and host (a) small bodies of ultramafic cumulates (original troctolites now transformed into magnesiochloritoid-bearing eclogites), (b) pods, layers and up to a few tens of meters thick bodies of Fe-Ti gabbro-derived eclogite, (c) transposed metabasaltic dykes and (d) m-sized layers of antigorite serpentinites and of well-preserved orthopyroxene-bearing metaperidotites. At the top of the Unit, a few m-thick metasedimentary cover locally occurs. The latter is made up of calcschists with layers of metabasites of ophiolite-detritus origin (Balestro et al., submitted). The analyzed metagabbros derive from olivine gabbro with TiO_2 = 0.2-0.5 wt% and Mg# = 83-78 (Piccardo & Fiora in Lombardo et al., 1978), whereas metatroctolite has Mg# = 88-84 (Messiga et al., 1999; Fig. 5D). No igneous relics are preserved in the eclogites that have Mg# ranging between 48 and 39, with TiO_2 of 5.8-4.6 wt% (Piccardo & Fiora in Lombardo et al., 1978; Rubatto & Hermann, 2003). The metabasaltic dykes have TiO_2 = 1.3-1.5 wt% and Mg# = 63-60 (Piccardo & Fiora in Lombardo et al., 1978).

(v) The Basal Serpentinite Unit: it is a thick sheet of antigorite serpentinite that can be followed both to the north (Balestro et al., 2011) and to the south (Lombardo et al., 1978) of the Monviso-Pian Melzè area. The serpentinite derives from a dominantly lherzolitic protolith, with only minor harzburgite and dunite, and hosts (a) large bodies of foliated eclogites deriving from both Mg-Al gabbro and Fe-Ti gabbro protoliths, (b) minor bodies of metaplagiogranite (i.e. the Verné plagiogranite-FeTi-oxide gabbro association: Castelli & Lombardo, 2007; Fig. 4), and (c) rare small boudins of fine-grained massive jadeitite (i.e. the Punta Rasciassa jadeitite; Compagnoni & Rolfo, 2003; Compagnoni et al., 2007; Compagnoni et al., 2012). Relics of ophicalcite and of a sedimentary cover mainly made up of calcschists were also found (Lombardo & Nervo in Lombardo et al., 1978; Mondino et al., 2004). The largest eclogite body occurs at the bottom of the Unit and has Mg# of 47 and TiO_2 = 2.8 wt% (Piccardo & Fiora in Lombardo et al., 1978). The upper part of the Unit consists of sheared antigorite serpentinite with blocks of Fe-Ti metagabbro (Fig. 5C), and to minor extent, lenses of Mg-Al metagabbroic and metasedimentary material, possibly representing a "tectonic mélange". However, this "mélange" is much thinner (between 50 and 100m thick) than suggested by Blake et al. (1995). Angiboust et al. (2011) demonstrated that this tectonic mélange can be continuously followed for 15 km at the base of the Mg-Al metagabbroic body (their "Lower Shear Zone"). Some of the eclogitic blocks embedded within antigorite schists along this shear zone are made of unusual eclogitic breccias consisting of cm-sized rotated angular fragments of mylonitic Fe-Ti metagabbro (now eclogite), cemented by a lawsonite-eclogite facies material (Angiboust et al., 2012b; Fig. 5e,f). This very unique fabric is interpreted as evidence for fossilized intermediate-depth seismicity during shearing under eclogite-facies conditions.

Igneous features of the Monviso Meta-ophiolite Complex

The tectonostratigraphic units of the Monviso Meta-ophiolite Complex belong to two different groups (Fig. 4):

(i) magma-rich units, in which a relatively thick basaltic layer caps gabbros and serpentinized peridotites (e.g. the Forciolline Unit);

(ii) magma-poor units, in which serpentinized peridotites are directly covered by a thin sedimentary sequence (e.g. the Basal Serpentinite unit).

Summing up, relict magmatic minerals, major-element compositions and compositional ranges of the metagabbros indicate derivation from protoliths of two types (Pl-Cpx±Ol±Spl±Opx Mg-Al gabbro and Pl-Cpx Fe-Ti gabbro) that were probably produced by fractional crystallization from tholeiitic melts at different fractionation stages (MONVISO, 1980). In this scenario, the metaplagiogranites (i.e. the Verné plagiogranite-FeTi-oxide gabbro association; Castelli & Lombardo, 2007) should correspond to later stages of the plutonic activity. Relict microstructures, major-element compositions and trace element data indicate as protoliths of the Monviso metabasalts both porphyritic (Pl- and Ol-phyric) and aphyric basalts of ocean-floor tholeiite affinity, with a fractionation trend mainly determined by removal of plagioclase and olivine (Lombardo et al., 1978; MONVISO, 1980). Available U-Pb zircon radiometric data suggest that the age of the magmatic activity is mostly in the range 163±2 Ma and 152±2 Ma (Rubatto & Hermann, 2003; Lombardo et al., 2002). Lombardo et al. (2002), noting this short duration of igneous activity in the Monviso Meta-ophiolite Complex and the short time span (from ca. 170 to ca. 150 Ma) for the entire Piemonte-Liguria Tethys, suggested an embryonic ocean (max 380 km wide; Piccardo et al., 2001; Schettino & Scotese, 2002) rather than a mature, slow spreading, Atlantic-type ocean model (e.g. Lagabrielle & Cannat, 1990).

The Alpine polyphase metamorphic evolution

The oceanic crust preserved in the Monviso Meta-ophiolite Complex experienced first an oceanic hydrothermal alteration (Nadeau et al., 1993) and then, during Alpine subduction, an eclogite-facies metamorphism (e.g. Lombardo et al., 1978; Nisio, 1985; Lardeaux et al., 1987; Nisio et al., 1987; Philippot & Kienast, 1989; Blake et al., 1995). The eclogitic stage was very pervasive and produced mineral assemblages characterized by Na-pyroxenes, Alm-rich garnets, lawsonite, and rutile. Some detailed structural, petrographic, fluid inclusions, and stable isotope studies on omphacite-bearing metamorphic veins from the Lago Superiore metagabbroic body, suggested that the veins formed by local circulation of fluids at eclogite-facies conditions (Philippot & Selverstone, 1991; Nadeau et al., 1993; Philippot, 1993). Recently, Spandler et al. (2011) suggested that some of these crack-seal veins may represent pathways for external fluids, attesting of larger-scale fluid circulation processes through the metagabbros during eclogite-facies metamorphism.

Twenty years of geothermobarometry yielded somewhat heterogeneous peak metamorphic conditions for the different tectonostratigraphic units. These *P-T* estimates range from 2.4 GPa and 620±50°C (for the rare Cr-rich magnesiochloritoid metatroctolite from the Lago Superiore Unit; Messiga et al., 1999; Fig. 5D) to 450±50°C and 1.2±0.3 GPa (for the Viso Mozzo Unit and for the previous Passo Gallarino Unit; Schwartz et al., 2000). Groppo & Castelli (2010) derived maximum *P-T* conditions of c. 550°C and 2.5 GPa for a lawsonite-bearing Fe-Ti eclogitic body from the southern sector of the Basal Serpentinite Unit, in line with previous thermobarometric estimates on metaplagiogranite from the same area (Castelli et al., 2002). Recently, Angiboust et al. (2012a) showed that the maximum *P-T* conditions of c. 550°C and 2.6 GPa (Fig. 6) are remarkably homogeneous across the whole Lago Superiore tectonometamorphic Unit for both metasedimentary and metagabbroic material. Slightly lower *P-T* estimates of c. 480°C and 2.2 GPa have been derived from the eclogitic rocks of the Monviso tectonometamorphic Unit (Angiboust et al., 2012a; Figs. 3, 6). Later exhumation produced re-equilibration under progressively lower pressures at epidote-blueschist- (c. 450°C, 1.0-1.5 GPa) and greenschist-facies (c. 400°C, 0.5-1.0 GPa) conditions, respectively, as suggested by the growth of blue amphibole (glaucophane to crossite), of amphibole + albite symplectites after omphacite, by the replacement of garnet by chlorite, of lawsonite by epidote, and of rutile by titanite (Lombardo et al., 1978).

Geochronological data on the Monviso Meta-ophiolite Complex suggest a Tertiary age for the Alpine HP metamorphism (Sm-Nd isochron age of 61±10 Ma from garnet + omphacite: Cliff et al., 1998; Ar-Ar age from mica of 50±1 Ma: Monié & Philippot, 1989; Lu-Hf isochron age from garnet of 49±1: Duchêne et al., 1997), which is in line with the radiometric ages obtained from other portions of the Internal Piemonte Zone. The most recent geochronological study of the Monviso Meta-ophiolite Complex is a U-Pb SHRIMP determination on zircons from a syn-eclogitic metamorphic vein, which yielded an Eocene age of 45±1 Ma (Rubatto & Hermann, 2003).

The internal structural coherency of the Monviso ophiolites has been over the last decades a matter of debate. Although the Lago Superiore tectonometamorphic Unit has been initially viewed as a relatively coherent sequence of oceanic lithosphere (Lombardo et al., 1978), some authors, noting that the eclogites record variable peak metamorphic pressures without connection with the size of tectonic blocks ranging from decimetre to kilometre scale, suggested that the Monviso Meta-ophiolite Complex does not represent a coherent fragment of the Tethyan lithosphere but rather an assemblage of tectonically juxtaposed slices that were subducted to different depths (Blake et al., 1995; Schwartz et al., 2000).

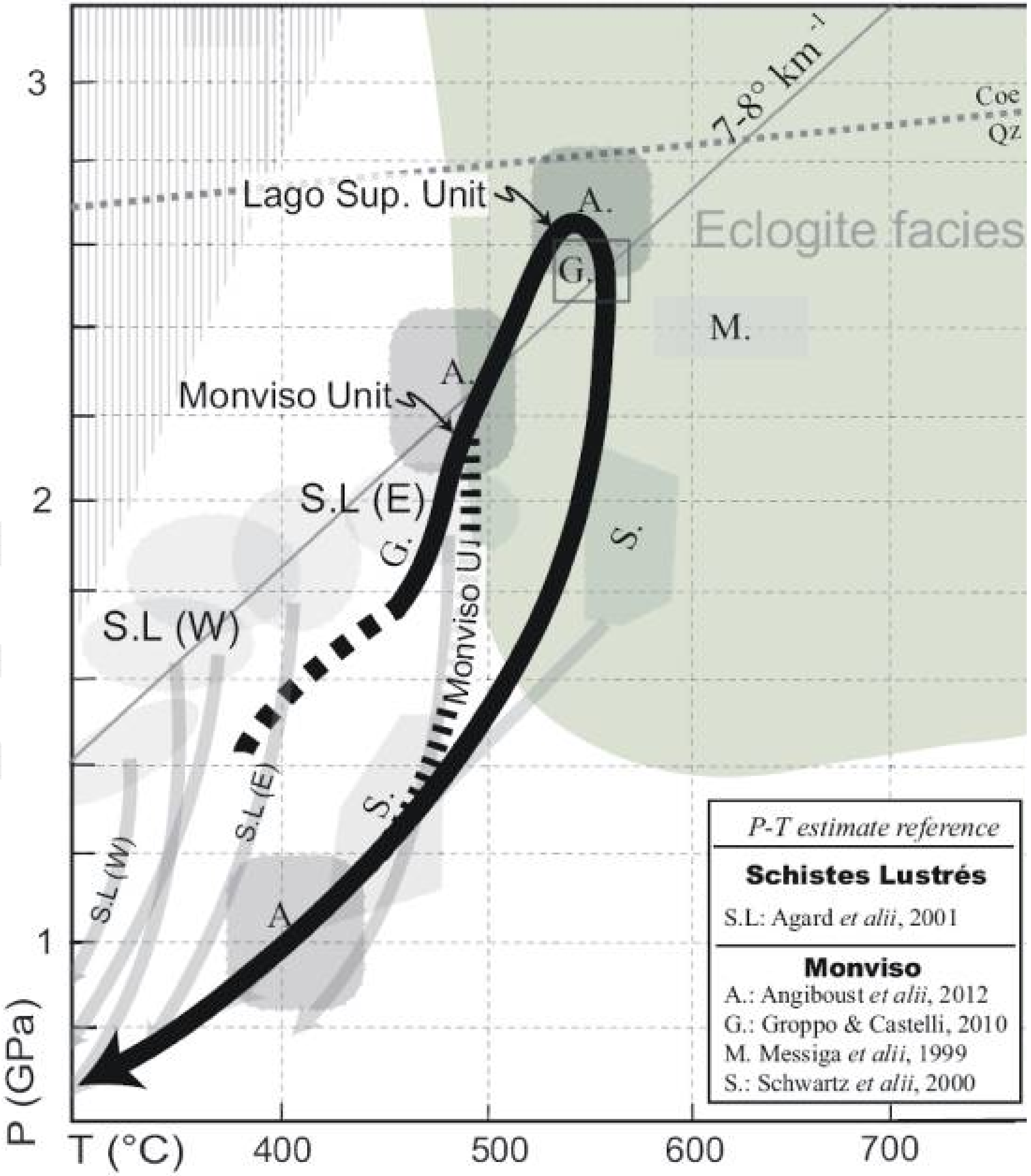


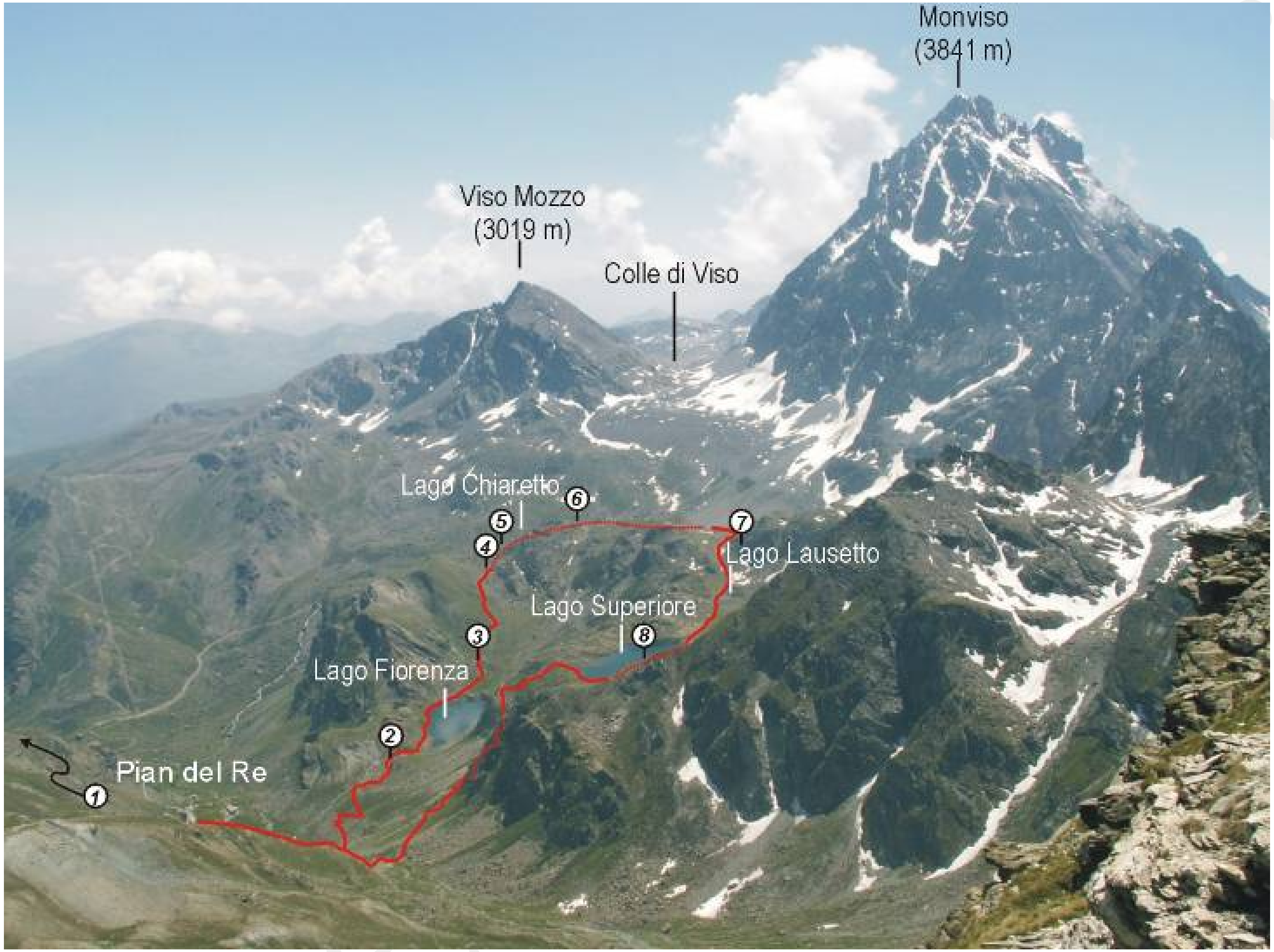
Fig. 6 - Compilation of P-T estimates of peak eclogite-facies conditions and of blueschist- to greenschist-facies re-equilibrations in the two tectonometamorphic units identified within the Monviso Meta-ophiolite Complex (modified from Angiboust et al., 2012a and references therein). Along the retrograde path, the *P-T* conditions become the same under blueschist-facies conditions. Peak metamorphic *P-T* estimates from the literature for adjacent tectonic units (such as the metasedimentary Schistes Lustrés) are also shown. Note that peak *P-T* estimates for the Monviso Meta-ophiolites are in line with the cold subduction gradient identified in Western Alps (7-8°C km^{-1}; Agard et al., 2001).

According to Guillot et al. (2004), this interpretation is supported by the numerical modelling of Gerya et al. (2002) and Stöckert & Gerya (2005) who showed that a subduction channel with low viscosity (~ 10^{19} Pa·s) favours the formation of a large melange. If the Monviso Meta-ophiolite Complex represents the exhumed part of a subduction channel, the ages of eclogites between 61±10 Ma and 45±1 Ma could be related either to different blocking temperatures of isotope systems, or, most likely, to different events in the serpentinized channel, giving a duration of the channel of 5 to 15 Ma (Guillot et al., 2004).
However, this model is apparently in contradiction with new field and thermobarometric data (Groppo & Castelli, 2010; Angiboust et al., 2012a), demonstrating that both the Monviso tectonometamorphic Unit (to the west) and the Lago Superiore tectonometamorphic Unit (to the east; Fig. 3) from a metamorphic viewpoint can be considered as two relatively coherent bodies, which can be followed over 15 km along-strike in the field. These data suggest that the Monviso Meta-ophiolite Complex can be viewed as a stack of two very large tectonometamorphic slices (Fig. 3).

Comparison with other Alpine Meta-ophiolites

Similar peak metamorphic conditions and exhumation patterns have been recently proposed for other portions of the Piemonte-Liguria domain, such as the Zermatt-Saas ophiolite (Angiboust et al., 2009) or Alpine Corsica (Vitale Brovarone et al., 2011b). Additionally, some similarities in the structure (such as the presence of large continuous crustal slices; Angiboust & Agard, 2010) or the preservation of almost undisturbed ocean-continent transitions (Zermatt-Saas, Beltrando et al., 2010; Corsica, Vitale Brovarone et al., 2011a) suggest that the Alpine H*P* deformation of oceanic units is apparently not chaotic ("mélange-like") but rather localized along shear zones preserving km-scale, coherent volumes detached at lawsonite-eclogite facies conditions (500-550°C, 2.2-2.6 GPa; Angiboust et al., 2012a). Such detachment may potentially occur in response to the entrance of the thinned continental margin in the subduction zone (Lapen et al., 2007; Agard et al., 2009) and/or be due to key dehydration reactions at depth (Bucher et al., 2005; Groppo & Castelli, 2010). In any case, the presence of a thick, highly buoyant serpentinized sole could have substantially favoured the preservation of such a large volume of dense, eclogitized oceanic crust (Hermann et al., 2000; Guillot et al., 2004; Angiboust & Agard, 2010; Angiboust et al., 2012c).

Day 1 – Itinerary



During the excursion we will visit the area south of the headwaters of the Po River, from Pian del Re to Lago Lausetto and Lago Superiore (Figs. 3, 7), where the Basal Serpentinite, Lago Superiore and Viso Mozzo tectono stratigraphic units (all belonging to the Lago Superiore tectonometamorphic Unit) are best exposed. Difficult access and lack of time unfortunately prevent direct observation of the upper Forciolline and Vallanta tectonostratigraphic units. The field trip follows a path which rises from 2000 m up to about 2400 m a.s.l. and consists of the following eight stops.

Fig. 7 – Panoramic view of the itinerary from M. Meidassa, with locations of Stops 2 to 8.

Route: By bus from Revello to the Valle Po up to Crissolo. By minivan from Crissolo to Pian del Re, the starting point of the field trip.

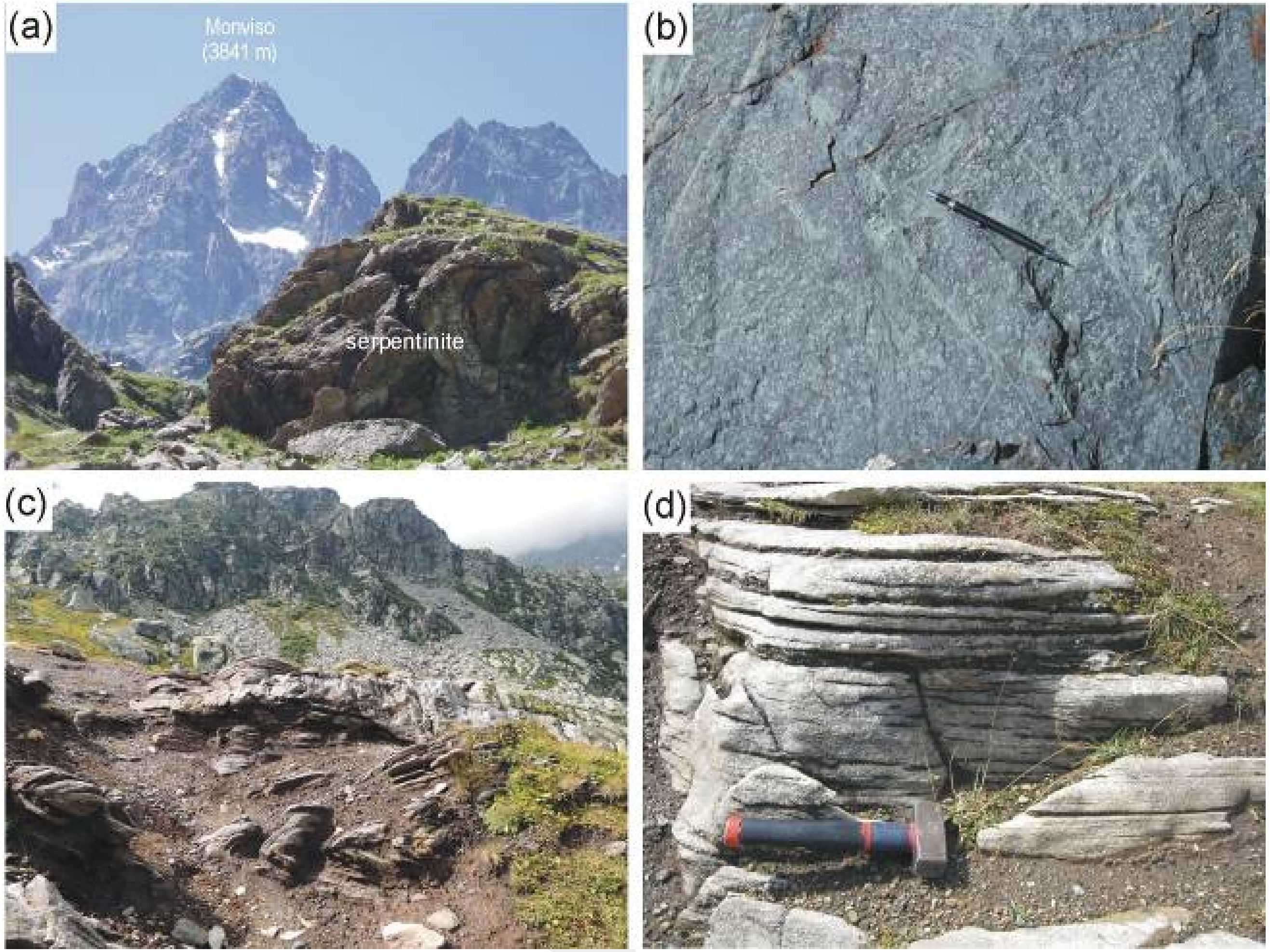
STOP 1.1: General view of the Monviso Meta-ophiolite Complex and serpentinite of the Basal Serpentinite Unit

Casa Cantoniera along the road from Pian Melzè to Pian del Re
(Coord: N 44°42' 12'' E 7° 06' 30'', elev. 1950 m a.s.l.)

From Crissolo to the small basin of Pian Melzè, the valley is cut in the belt of Middle Triassic dolomite marble and of calcschist. This Triassic belt separates the Monviso Meta-ophiolite Complex from the underlying Dora-Maira thrust units. The Stop is located between Pian Melzè and Pian del Re, at km 7.5 of the road to Pian del Re. A general view of the internal structure of the Monviso Meta-ophiolite Complex can be observed looking southward to the ridge trending NE-SW from Monte Grané (2314 m) to Viso Mozzo (3019 m) and Monviso (3841 m). Along this section are exposed, from left to right and from bottom to top: (i) antigorite serpentinite of the Basal Serpentinite unit, cropping out at Monte Grané and in the lower ridge above Pian Melzè; (ii) smaragditic metagabbros (foot of Viso Mozzo and low ridge above Pian del Re) of the Lago Superiore unit (see Stop 1.5); (iii) the Viso Mozzo unit metabasites and the Colle di Viso serpentinite, which crop out in the low between Viso Mozzo and Monviso. This serpentinite sliver separates the two main tectonometamorphic units composing the Monviso Meta-ophiolite Complex (the Monviso tectonometamorphic Unit to the west and the Lago Superiore tectonometamorphic Unit to the east). The pyramid of Monviso and the sharp ridge leading NW to Visolotto and Punta Udine are cut in the Vallanta and Forciolline units, respectively (Fig. 3).
In this stop, nice examples of serpentinite and serpentinized peridotite blocks from the Basal Serpentinite Unit may be examined. They contain typical rodingites derived from metasomatic alteration, coeval with the serpentinization process, of original basaltic dykes. The rodingitized dykes consist of a fine-grained massive garnetite core bounded by two Mg-chlorite-rich selvages.
Similar lithologies may be observed along the path from the parking place of Pian del Re to the Po River source (see Stop 1.2), lined by two alignments of m-sized blocks, also derived from the Basal serpentinite Unit.

STOP 1.2: Serpentinite of the Basal Serpentinite Unit

Along the path from Pian del Re to Lago Fiorenza
(Coord: N 44° 41’ 57’’ E 7° 05’ 36’’, elev. 2075 m a.s.l.)



The first outcrop of serpentinite (Fig. 8a) can be reached with a few minutes’ walk along the path leading southward from Pian del Re to Rifugio Quintino Sella. The path starts near the springs traditionally considered as the sources of the Po River. Most of the lithologies cropping out along the path may be observed, as boulders and blocks, around the springs.

Fig. 8 – (a,b) Stop 1.2: Serpentinite of the Basal Serpentinite Unit. (a) Outcrop of serpentinite above Pian del Re; (b) Massive serpentinite with relics of the peridotitic clinopyroxene. (c,d) Stop 1.3: Calcschists and micaschists (Basal Serpentinite Unit). (c) Folded calcschists and impure marbles. (d) Foliated quartzite.

The serpentinite, deriving from primary spinel-lherzolite, is either massive (Fig. 8b) or sheared. It mainly consists of antigorite, with variable amounts of clinopyroxene, brucite, Mg-rich chlorite, Ti-clinohumite, metamorphic olivine and chrysotile; the ore minerals are magnetite, FeNi-alloys and sulphides. The peridotite

clinopyroxene, locally preserved as dusty relics crowded with very fine-grained magnetite, has been altered to antigorite ± magnetite, or partly recrystallized to fine-grained diopside. The Mg-rich chlorite occurs as lens-like aggregates, most likely pseudomorphous after primary spinel. Brucite appears to develop at the expense of metamorphic olivine.
Three different generations of veins crosscutting the serpentinite have been observed: (i) olivine + magnetite ± FeNi-alloys + Ti-clinohumite veins; (ii) brucite ± Fe-rich chrysotile veins; (iii) lizardite veins.
A few tens of meters to the east from the stop, a m-thick rodingitic metagabbro dike is exposed within the serpentinite.

STOP 1.3: Calcschists and micaschists (Basal Serpentinite Unit); Cr-rich magnesiochloritoid-bearing eclogite loose blocks (Lago Superiore unit)

Along the path, 500 m SSE of Lago Fiorenza
(Coord: N 44° 41’ 28’’   E 7° 05’ 55’’, elev. 2290 m a.s.l.)

The path crosses, between Lago Fiorenza and the small pass at the upper end of the Lago Fiorenza valley (Colletto Fiorenza), a folded elongated body of metasediments tectonically embedded within serpentinite of the Basal Serpentinite Unit. This ca. 50 m-thick sliver belongs to an eclogite-facies shear zone which lines the bottom of metagabbro (Lago Superiore Unit) and passes through Colletto Fiorenza (Stop 1.4). The metasediments consist of impure marble, calcschist and micaschist (Italian: "calcescisti"; French: "schistes lustrés") (Fig. 8c) grading to quartzite (Fig. 8d). Rare cm-thick layers of metabasite are interbedded with metasediments.
Impure marbles are massive and mainly consist of granoblastic calcite with minor, randomly oriented, flakes of white-mica (phengite ± paragonite), roundish dolomite crystals and quartz aggregates, and accessory Mg-chlorite, rutile and graphite. Very fine-grained aggregates of chlorite + white mica, dark because of the presence of dusty graphite, are pseudomorphs after former lawsonite.
Calcschists usually exhibit a well-developed foliation, defined by the preferred orientation of white-mica flakes. They usually consist of carbonate minerals (calcite and ankerite), white-mica (phengite ± paragonite), quartz and chlorite (both Mg- and Fe-rich chlorite), but locally may contain clinozoisite/epidote and chloritoid. The common accessory minerals are graphite, sulphides, apatite, titanite, rutile and tourmaline. Peak *P-T* estimates based on Raman Spectroscopy of Organic Matter (Beyssac et al., 2002) and pseudosection

modelling provided temperatures of 550°C ± 25°C and pressures of c. 2.6 GPa for this metasedimentary material (Angiboust et al., 2012a).
Quartzite is characterized by major amounts of phengite and garnet, but carbonates (mainly ankerite), clinozoisite/epidote, Mg-chlorite and, interestingly, blue-amphibole also occur. Accessory minerals are titanite, rutile and apatite. The epidote grains are commonly zoned with an allanite core overgrown by a clinozoisite/epidote rim. Garnet is partially altered to Fe-chlorite and stilpnomelane, and zoned blue-amphibole (glaucophane to crossite) breaks down to green amphibole + albite. Microstructural relationships indicate two different parageneses in the quartzite: an older one consisting of quartz, phengite, garnet, blue-amphibole, ankerite and allanite; a younger one composed of Fe-chlorite, epidote, albite, green-amphibole and stilpnomelane.
The Cr-rich magnesiochloritoid eclogite boulders occurring near Stop 1.3 are described after introducing the smaragdite metagabbro of Lago Chiaretto (Stop 1.5).

STOP 1.4: Serpentinite with tectonic inclusions of eclogitic rocks (Basal Serpentinite Unit)

Colletto Fiorenza
(Coord: N 44° 41’ 22’’ E 7° 05’ 54’’, elev. 2265 m a.s.l.)

The tectonic contact between calcschists and the overlying metagabbro is marked by a zone from a few m- to a few tens of m-thick of strongly sheared serpentinite and talc-carbonate-amphibole-schist, with tectonic inclusions of banded eclogitic rocks.
The serpentinite-metagabbro contact is characterized by the presence of a heterogeneous rock (Fig. 9a), mainly consisting of tremolite-actinolite (coarse-grained crystals arranged as radiated aggregates; Fig. 9b), talc and carbonate in largely variable amounts from place to place.
At the very contact with serpentinite, also the metagabbro is strongly sheared. Under the microscope it consists of omphacite porphyroclasts (pseudomorphous after magmatic pyroxene and partly recrystallized to fine-grained omphacite) and garnet porphyroblasts overgrowing a very fine-grained mylonitic foliation, defined by omphacite and minor Mg-chlorite.
A fine example of banded eclogitic rock (Fig. 9c), which is possibly the fold hinge of a refolded fold, may be observed further on along the path leading to Lago Chiaretto. The rock consists of interbedded layers (from a

few mm to a few cm in thickness) of Mg-chlorite eclogite, fine-grained omphacitite, and Mg-chlorite-omphacite garnetite. Mg-chlorite, garnet and omphacite, which are in stable association, represent an unusual eclogitic assemblage (Kienast, 1983). Garnet commonly has atoll-like habit. Accessory minerals are rutile and rare Fe-rich epidote usually overgrowing allanite. The eclogitic assemblage is locally retrogressed, with development of Fe-chlorite after garnet, albite + chlorite + dark-green amphibole after omphacite and titanite after rutile.

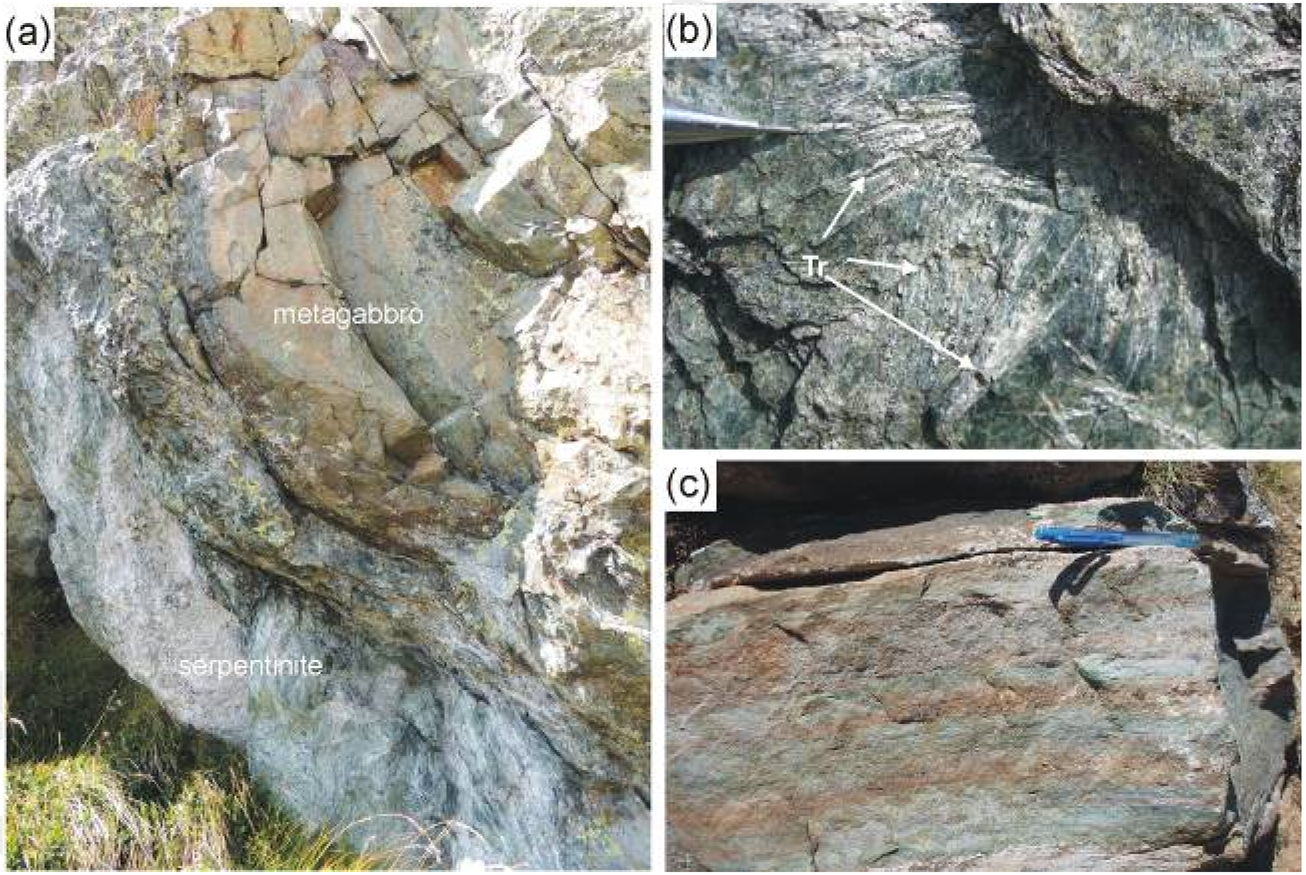


Fig. 9 – Stop 1.4: (a,b) Contact between serpentinite and metagabbro at Colletto Fiorenza. The serpentinite-metagabbro contact is characterized by the presence of a heterogeneous rock, mainly consisting of tremolite-actinolite (coarse-grained crystals arranged as radiated aggregates; b), talc and carbonate. (c) Banded eclogite consisting of interbedded layers of Mg-chlorite eclogite, fine-grained omphacitite, and Mg-chlorite-omphacite garnetite.

STOP 1.5: Smaragdite metagabbro with transposed dykes of metabasalt (Lago Superiore Unit)

Between Colletto Fiorenza and Lago Chiaretto
(Coord: N 44° 41’ 16’’ E 7° 05’ 53’’, elev. 2280 m a.s.l.)

The smaragdite metagabbro, which locally includes transposed cm- to dm-thick dykes of metabasalts, is exposed along the path about 200 m south of Colletto Fiorenza (Fig. 10a). The metagabbro shows a well-developed foliation and consists of a whitish matrix, in which bright-green porphyroclasts of "smaragdite" (up

to two cm in length) are preserved (Fig. 10b,c). Under the microscope, the smaragdite metagabbro appears to consist of pale-green Cr-omphacite porphyroclasts, enclosed in a foliated matrix of zoisite, garnet, Mg-chlorite, tremolite, talc, rutile, or their greenschist-facies retrogression products, albite, tremolite/actinolite, Fe-rich chlorite, clinozoisite/epidote, quartz and titanite. Pseudomorphous reactions are common and thus the original igneous mineralogy may be inferred from the metamorphic assemblages: (i) the magmatic clinopyroxene has been replaced by a single crystal of Cr-omphacite ± tremolite ± talc (="smaragdite"), often surrounded by an aggregate of fine-grained omphacite; (ii) the plagioclase by jadeitic pyroxene + zoisite; (iii) the olivine by an aggregate of talc ± tremolite, usually surrounded by an irregular garnet rim; (iv) the Ti-bearing opaque ore (possibly titanomagnetite) by a rutile aggregate. Locally, a layering is observed in the smaragdite metagabbro, suggesting a *cumulus* structure: one of such layers, about 10 cm thick, has the composition of a troctolite and consists of omphacite, chloritoid and rutile, occasionally found in textural equilibrium with garnet and talc.
The greenish-gray to pinkish metabasaltic dykes are very fine-grained and locally preserve remnants of a porphyritic structure. Microscopically, the metabasalts show homogeneous grain-size and mainly consist of omphacite, garnet, clinozoisite and minor glaucophane, quartz, Mg-chlorite. Rutile is the typical accessory mineral. Locally, metabasaltic dykes are crosscut by late albite veins and the eclogitic assemblage is deeply altered to greenschist-facies albite, Fe-chlorite, actinolite and titanite. Scanty paragonite porphyroblasts (partially transformed to albite) have overgrown the foliation.
Pseudomorphous replacements occurring in the smaragditic metagabbro of Lago Chiaretto are similar to those described by Kienast & Messiga (1987) in Cr-rich magnesiochloritoid eclogites found as boulders (Messiga et al., 1999; Fig. 5d) near Stop 1.3. They are characterized by dark, rounded pseudomorphous aggregates of chloritoid ± talc ± omphacite ± magnesite after igneous olivine set in a whitish to emerald-green groundmass mainly consisting of omphacite + chloritoid ± paragonite, garnet and talc after igneous plagioclase, with scattered relics of igneous clinopyroxene and spinel. Some boulders preserve igneous layering textures, with emerald-green bands (enriched in Cr-omphacite) corresponding to layers with high modal clinopyroxene. The bulk-rock compositions of such unusual layers are similar to spinel-bearing metatroctolite and troctolite from northern Apennine ophiolites (Messiga et al., 1999) or from the Pennine Alps of Switzerland (Zermatt-Saas meta-ophiolite, Allalin metagabbro; Bucher & Grapes, 2009).

STOP 1.6: Layered eclogitic sequence (Lago Superiore Unit)

Along the path from Lago Chiaretto to Lago Lausetto
(Coord: N 44° 41' 16'' E 7° 05' 53'', elev. 2280 m a.s.l.)

West of the smaragdite metagabbro, along the path leading to Lago Lausetto and Rifugio Giacoletti, layered eclogitic rocks polished by the glacier are exposed. They consist of alternating layers of eclogite and metagabbro from a few cm- to several dm-thick (Fig. 10d). Eclogite largely prevails over metagabbro, but intermediate lithologies can also be found. The sequence is folded by open folds, trending N40W with sub-horizontal axes. Structural and chemical data indicate that melanocratic layers (now eclogite) derive from primary FeTi-oxide gabbro, whereas leucocratic layers from Mg-Al gabbro protoliths similar but not identical to typical ophiolitic Mg-gabbros. In the field, the two lithologies are closely associated and interlayering appears to be a primary feature, later folded and transposed by a mylonitic foliation mainly defined by the parallel alignment of rutile trails. Locally, mm- to cm-sized omphacite porphyroclasts are preserved, which are remnants of the original igneous pegmatoid structure. The metagabbros are discontinuously covered by a few m thick metasedimentary sequence that is made up of calcschists with abundant levels of ophiolitic *detritus*.
Microscopically, the eclogite shows homogeneous grain-size ranging from 0.2 to 0.5 mm; locally, porphyroclasts of zoned omphacite, pseudomorphous after magmatic clinopyroxene, are included in a foliated matrix consisting of garnet and omphacite, minor blue amphibole, rutile, lawsonite pseudomorphs (now consisting of lozenge-shaped epidote/paragonite aggregates) and phengite with accessory abundant apatite, titanite, opaque ores, allanite and zircon. Porphyroblasts of blue amphibole (usually crossite), and phengite clearly overgrow the rock foliation. The eclogitic foliation is cut by different generations of veins characterized by coarse-grained zoned omphacite; crossite + Fe epidote ± albite; crossite + albite + titanite; albite + Fe-epidote + Fe-chlorite ± dark-green amphibole ± stilpnomelane.
At the eclogite-facies peak, the Mg-Al metagabbro consisted of porphyroclastic omphacite (pseudomorphous after magmatic clinopyroxene) enclosed in a matrix of zoisite, fine-grained omphacite, minor garnet and rutile, and accessory apatite and quartz. Porphyroblasts of blue amphibole (ranging in composition from glaucophane to crossite) and phengite locally have overgrown the mylonitic foliation. Most metagabbros, however, are strongly retrogressed and mostly consist of the greenschist-facies mineral assemblage albite, clinozoisite/epidote, actinolite, Fe-chlorite and titanite.

In conclusion, three main parageneses, different in age and mineralogy, may be recognized in the eclogitic sequence: (i) omphacite, garnet, rutile, opaque ores, zircon; (ii) glaucophane/crossite, phengite, clinozoisite, titanite; (iii) albite, Fe-chlorite, green amphibole, clinozoisite, Fe-rich epidote ± stilpnomelane. The main rock foliation formed before the development of the second mineral assemblage.

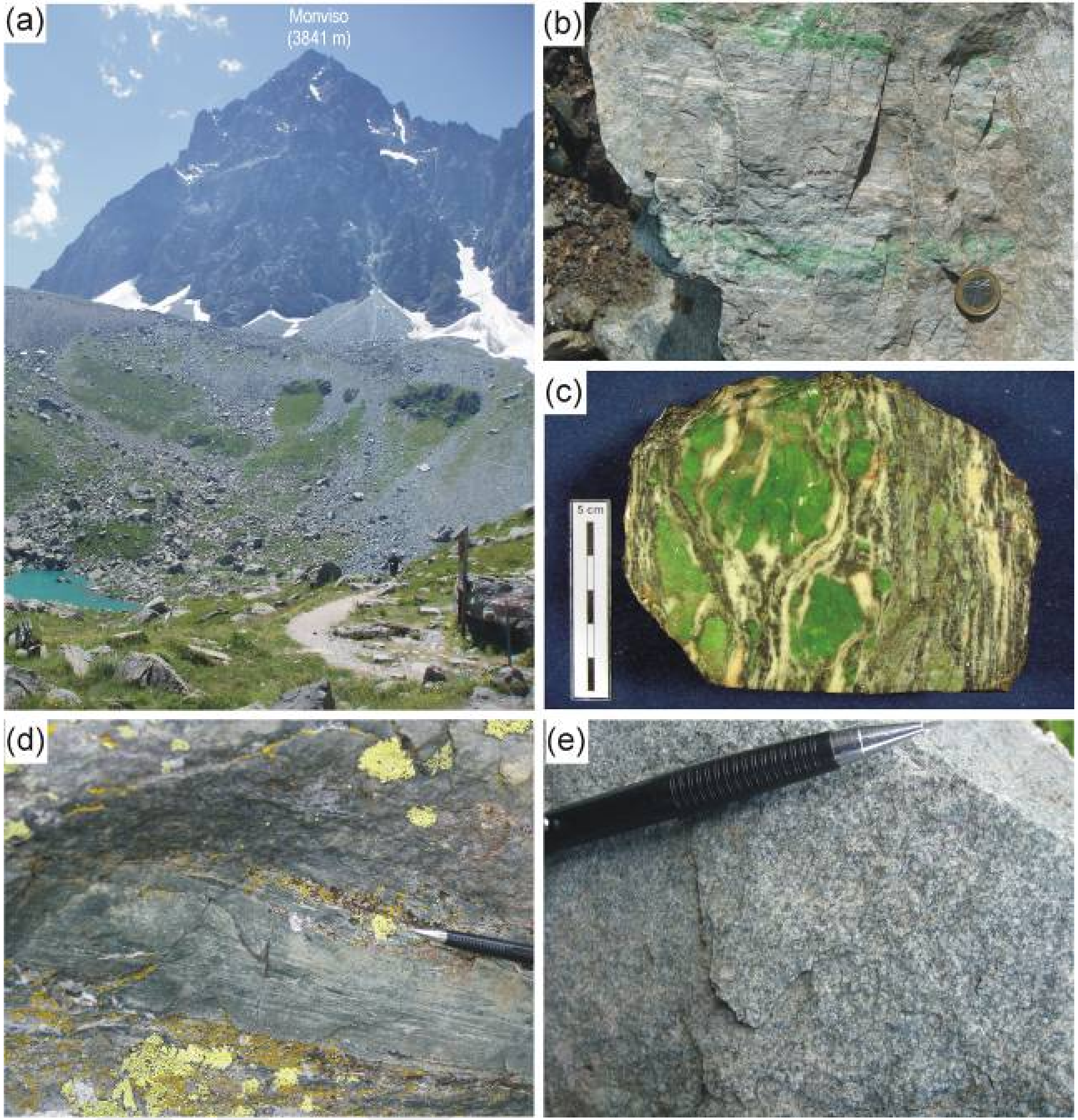


Fig. 10 – (a,b,c) Stop 1.5: Smaragdite metagabbro at Lago Chiaretto. (a) View of Lago Chiaretto; (b,c) Examples of smaragdite metagabbro. (d) Stop 1.6: layered eclogitic rocks exposed along the path from Lago Chiaretto to Lago Lausetto. (e) Stop 1.7: prasinite cropping out at Lago Lausetto.

STOP 1.6a (optional): FeTi-oxide metagabbro blocks in the serpentinite from an eclogite-facies shear zone (Lago Superiore Unit)

Northern side of Lago Chiaretto (300 m SE of Stop 1.5)

In the cliff *c.* 200 m north of Lago Chiaretto are observed rounded or lens-shaped 5-10 m wide blocks of mylonitized Fe-Ti metagabbro embedded within a matrix dominantly composed of strongly foliated serpentinite. This zone of tectonic mixing (c. 50m thick) is sandwiched between the prasinite and the metagabbro from the Lago Superiore Unit. Lenses of prasinite (± calcschist) can also be observed at the vicinity of the contact with the prasinite. Although a significant amount of rotation, folding and flattening is observed, it is emphasized that lithologies observed within the shear zone are rigorously identical to the walls of the shear zone (except serpentinite). Angiboust et al. (2011) noted that serpentinite slivers systematically line the major eclogite-facies shear zone networks in the Lago Superiore Unit. These slivers could have been tectonically emplaced during shearing and displacement along the eclogite-facies shear zone. Ongoing deformation led to the dissemination of abraded shear zone wall fragments within mylonitized serpentinite. This outcrop also shows that the “block-in-matrix” structure (which is widespread in the Monviso Meta-ophiolite Complex) does not necessarily reflect a “mélange” dynamics but could rather testify of complex, localized abrasion processes occurring at depth along a crustal shear zone under eclogite-facies conditions.

STOP 1.7: Prasinite (Viso Mozzo Unit)

South of Lago Lausetto, along the path leading to Rifugio Giacoletti
(Coord: N 44° 41’ 22’’   E 7° 05’ 19’’, elev. 2400 m a.s.l.)

A narrow alluvial plane, approximately trending NW-SE, masks the trace of the eclogitic shear zone separating the Fe-Ti metagabbro mylonites (to the E) from foliated prasinite (to the W). Strongly mylonitized serpentinite slivers can be occasionally observed along the shear zone (for instance at the small pass south of Lago Lausetto or on the ridge north of Lago Superiore; Fig. 3).
No primary features, such as pillow structure or porphyritic texture, are preserved within the prasinite. However, on the ground of the homogeneous fine grain-size and rock bulk chemistry, and by comparison with similar pillowed rocks of the Monviso Meta-ophiolite Complex, the prasinite may be considered as the metamorphic product of original basaltic flows. Additionally, the prasinite lies in a structural position which is identical to the prasinites from Viso Mozzo slopes (in which rare relics of pillows can be observed).

Prasinite usually exhibits a mm-thick layering characterized by alternating darker and lighter layers, which consist of glaucophane ± Na-pyroxene and epidote + albite, respectively. The poorly developed foliation is defined by the preferred dimensional orientation of prismatic minerals. Homogeneous prasinite (Fig. 10e) is fine-grained and mainly composed of albite, clinozoisite/epidote, chlorite, amphibole, and relict accessory rutile, titanite, apatite and opaque ores. The amphibole is crossite, usually with a rim of a dark-green amphibole; crossite appears to be in stable association with albite. In the crossite-rich layers remnants of a Na-pyroxene (usually aegirine-augite) are frequently recognized. Locally, phengite and calcite also occur, suggesting its derivation from original tuffite.
The homogeneity of T_{max} estimates (*c.* 520-550°C; based of Raman Spectroscopy of Organic Matter and conventional thermobarometry) suggests that the observed difference in metamorphic grade between the two walls of the shear zone does not reflect differences in *P-T* conditions but rather variable degree of resistance to late retrograde blueschist- to greenschist-facies deformation (Angiboust et al., 2011, 2012a).

STOP 1.8: FeTi-oxide metagabbro boudins and eclogite-facies shear zones and veins (Lago Superiore Unit)

Western side of Lago Superiore
(Coord: N 44° 41' 44''   E 7° 05' 14'', elev. 2325 m a.s.l.)

At Lago Superiore (Fig. 11a), the layered eclogitic sequence (Stop 1.6) has been extensively studied. According to Philippot & Kienast (1989), the main structural feature of the area is a 10- to 30 m-thick major ductile shear zone composed of mylonitic rocks and bounded at the hanging wall by a thrust fault that separates the layered sequence from the overlying, mainly metabasaltic rocks of the Viso Mozzo Unit (Fig. 3, see also Angiboust et al., 2011). Within the sequence, small (10x10 cm) lenses of low-strain domains still preserve the igneous microstructure and are surrounded by large volumes of eclogite-facies mylonites. Cigar-shaped boudins of eclogite within the metagabbro outline a complex interference pattern of folds. The long axis of the boudins strikes N-S. A marked Na-pyroxene and garnet stretching mineral lineation L_1 is also oriented N-S (Fig. 3). Mylonites consist of porphyroclastic omphacite in a fine grained matrix of recrystallized omphacite, rutile and almandinic garnet (Philippot & Kienast, 1989).
Eclogite-facies veining (Fig. 11b-d) is an important feature of this area and has been used to characterize the fluid migration during subduction (Philippot, 1987, 1993; Philippot & Kienast, 1989; Philippot & Selverstone,

1991; Philippot & van Roermund, 1992; Nadeau et al., 1993; Spandler et al., 2011) and the nitrogen deep cycle (Busigny et al., 2011). Three types of eclogite-facies veins (Fig. 12) have been recognized in the eclogite boudins (Philippot, 1987):

(i) tension gashes trending sub-perpendicular to both the foliation plane S_1 and lineation L_1; (ii) dilatant fractures parallel to S_1; and (iii) complex sets of shear fractures that randomly crosscut both S_1 and L_1. Tension gashes and dilatant fractures are small (a few cm thick and 20-50 cm long) in comparison with dm- to m-scale shear fractures. Type (i) and (ii) veins contain Na-pyroxene ± rutile ± apatite; type (iii) veins contain a Na-pyroxene-garnet-rutile-apatite assemblage.

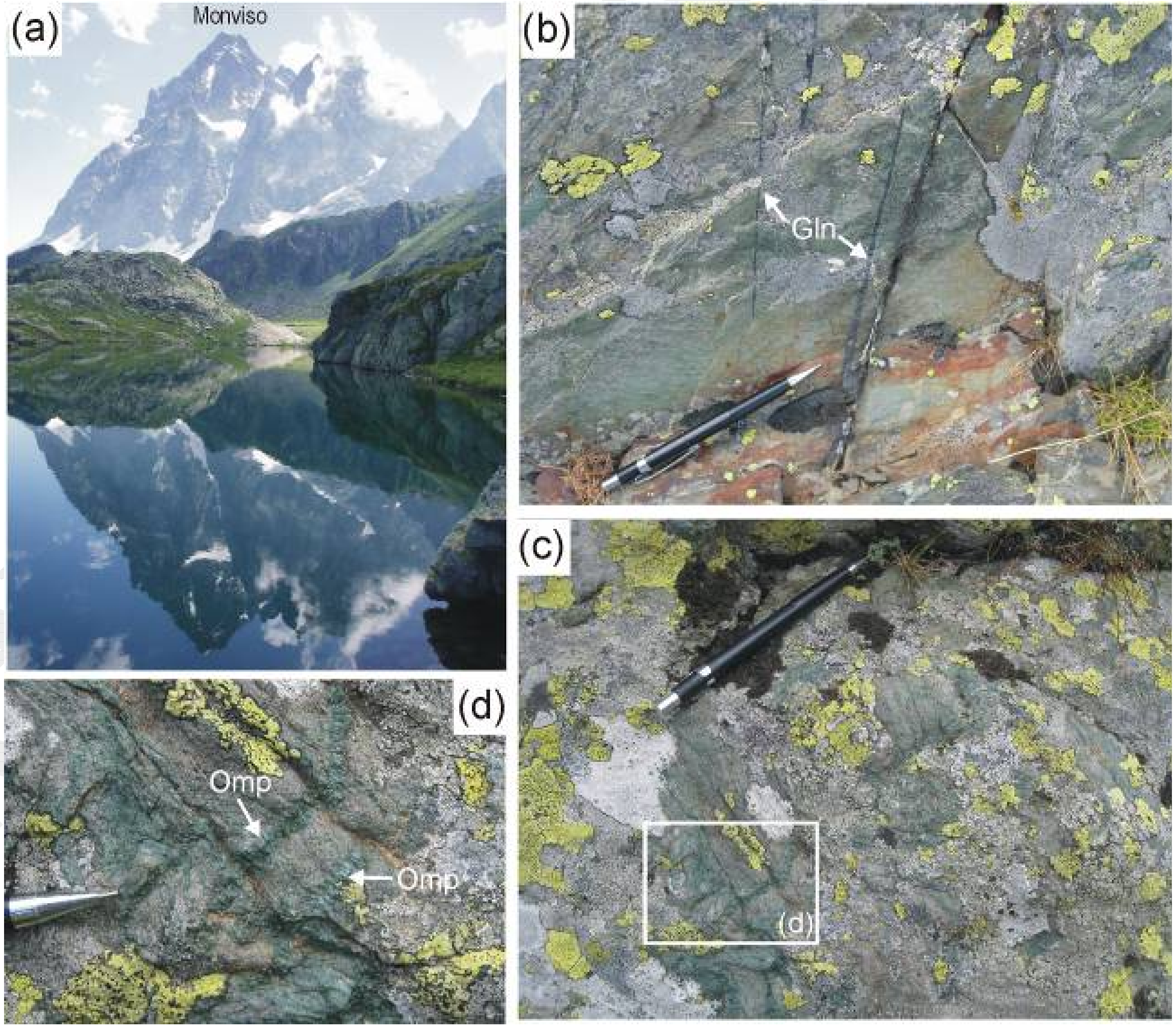


Fig. 11 – Stop 8: FeTi-oxide metagabbro boudins and eclogite-facies veins. (a) Outcrops at Lago Superiore; (b,c,d) Glaucophane- (b) and omphacite- (c,d) bearing veins cross-cutting the eclogitic FeTi-oxide metagabbro foliation.

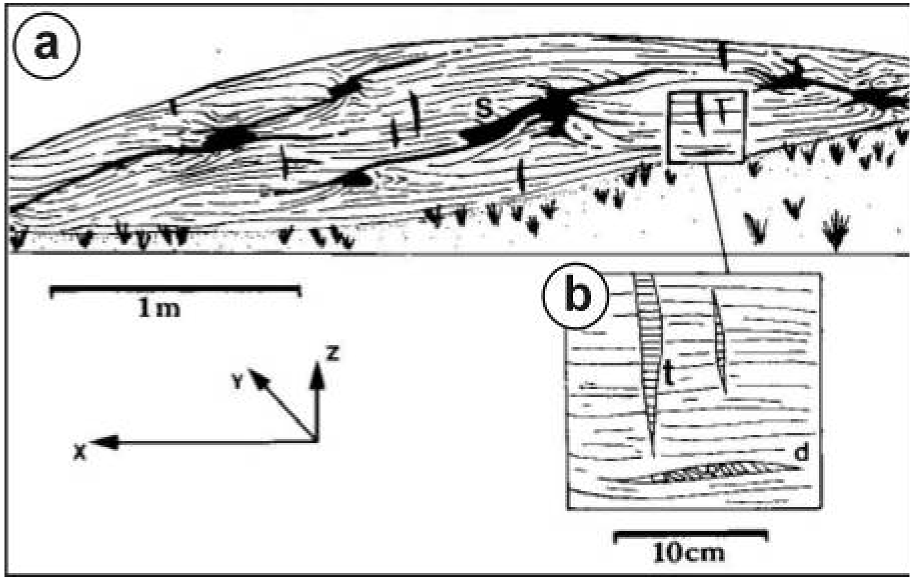


Fig. 12 - Geometry of the three vein types occurring in boudins of eclogitic metagabbro from Lago Superiore; **d**: dilatant fractures; **t**: tension gash; **s**: shear fracture (simplified from Philippot & Selverstone, 1991, their Fig. 2).

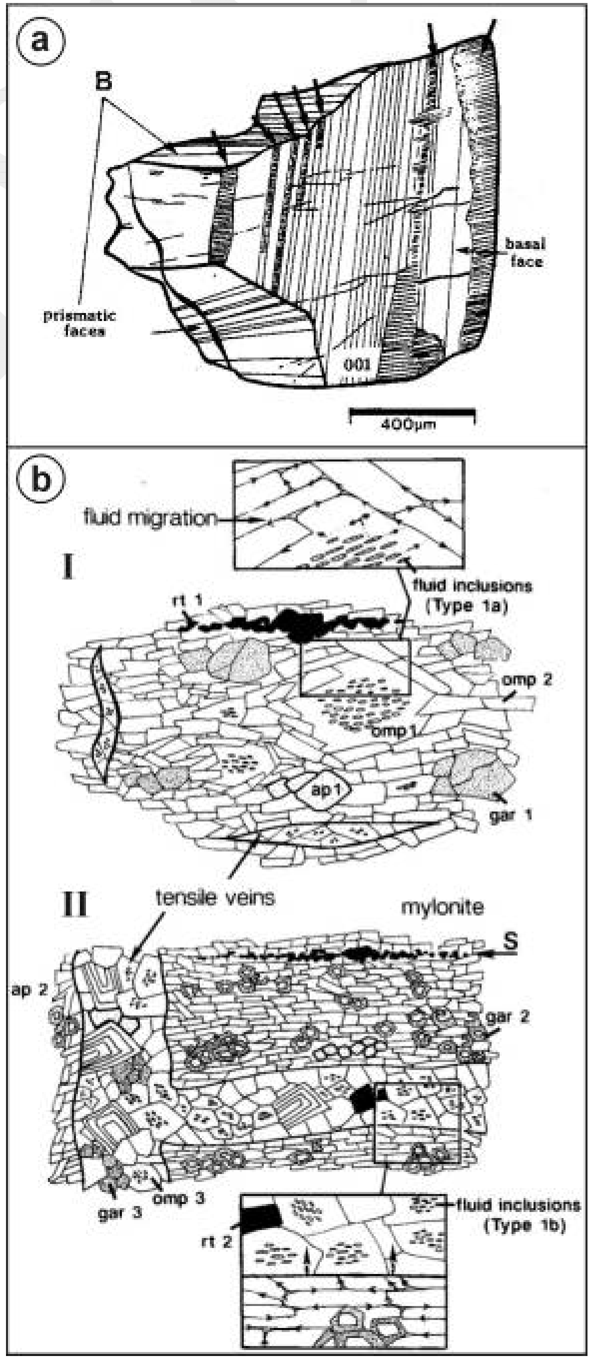


Fig. 13 - (a) Example of oscillatory-zoned pyroxene from the Lago Superiore dilatant veins and shear fractures. Arrows point to bands of fluid inclusions parallel to the growth zones; distributions of daughter minerals varies from one fluid inclusion band to another (simplified from Philippot & Selverstone, 1991, their Fig. 4). (b) Model of fluid migration involving localized fluid production and transport during ductile-brittle deformation of the Lago Superiore metagabbros. Sketches I and II correspond to two different increments of shear and fluid release in a moderately-strained and in a highly-strained mylonitic layer, respectively (Fig. 11 in Philippot & Selverstone, 1991). During the first increment of cracking (I), dynamic recrystallization of fluid inclusion-rich pyroxene clasts releases early inclusion fluids, which in turn promotes dislocation flow and subsequently wets the grain boundaries of the minerals (omp1: pyroxene clasts; omp2: recrystallized pyroxene grains; gar1, rt1 and ap1: fragmented garnet, rutile, and apatite). Sketch II shows a subsequent development of tensile fractures oriented parallel and normal to the foliation (S) (gar2: atoll-shaped garnets; omp3: vein pyroxene, locally with oscillatory growth zones; gar3, rt2, ap2: vein garnet, rutile and apatite). Arrows point to fluids wetting the grain boundaries of the minerals in the mylonitic rocks and show subsequent migration of the fluids within the veins.

These pyroxene grains are thought to have grown during successive crack-seal events. Boudins, folds and fractures are considered to have formed progressively during non-coaxial deformation (Philippot, 1987, 1988; Philippot & Kienast, 1989). Within all veins, many of the pyroxene grains show strong oscillatory and sector zoning (Fig. 13a), with jadeite-richer cores and diopside-richer rims (Philippot & Kienast, 1989). Remarkable Cr and Ni oscillatory enrichment in garnet, omphacite or rutile has been recently reported by Spandler et al. (2011).

In mylonites and associated veins, Philippot & Selverstone (1991) recognized five textural types of fluid inclusions that mainly occur in oscillatory- and sector-zoned omphacite (Fig. 13a). The inclusions turned out to contain aqueous brines and daughter crystals (halite, sylvite, calcite, dolomite, albite, anhydrite and/or gypsum, barite, baddeleyite, rutile, titanite, Fe-oxides, pyrite and monazite) that provide direct evidence for transport of HFSE, LILE, and LREE during subduction. Measured δD of fluid inclusions in the vein omphacite further supports that fluid flow was limited during eclogite-facies metamorphism; continuous fluid redistribution during brittle/plastic deformation modified the main features (salinity, composition and isotopic imprint) of the fluid phase on a cm-scale, but the system is thought to have been essentially closed (Philippot, 1993). Based on these data, Philippot & Selverstone (1991) proposed that the scale of fluid equilibration was small and that fluid, released in pulses, controlled the mineral deposition and growth (Fig. 13b). Spandler et al. (2011), even confirming that these veins largely formed from internally derived fluids, suggested that transient and episodic ingression via brittle fractures of external fluids (mainly derived from dehydration of serpentinites) is necessary to explain the local Cr, Ni and Mg enrichments.

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Day 2 - Metasomatism from & to ultramafics: the UHP continental Brossasco-Isasca Unit

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Geological setting of the southern Dora-Maira Massif

The Dora-Maira Massif (DMM) belongs to the "Internal Crystalline Massifs" of the Pennine Domain of western Alps (Fig. 1). The first modern geologic and petrographic study of the DMM was done by Vialon (1966) who subdivided the massif into three main lithotectonic units: the western and uppermost "Sampeyre-Dronero Unit", the "Polymetamorphic Unit", and the underlying "Pinerolo Unit". As to the southern DMM (Fig. 14), its rocks have become widely known after the discovery of coesite (Chopin, 1984) and other ultra-high pressure minerals and mineral assemblages, including pyrope, jadeite + kyanite, magnesio-staurolite, magnesio-dumortierite and the peculiar ellenbergerite-phosphoellenbegerite solid solution series (Chopin, 1987; Chopin et al., 1986, 1993, 1994, 1995). Such unusual minerals are associated to the only kyanite-bearing eclogites known in the western Alps (Kienast et al., 1991; Nowlan et al., 2000).

Detailed field and petrographic investigations (e.g. Henry, 1990; Chopin et al., 1991; Henry et al., 1990; Compagnoni & Hirajima, 1991; Michard et al., 1993, 1995; Sandrone et al., 1993; Turello, 1993; Compagnoni et al., 1995; Chopin & Schertl, 1999; Matsumoto & Hirajima, 2000; Groppo, 2002; Compagnoni et al., 2004) have shown that the southern DMM is a pile of imbricated thrust sheets, resulting from the Alpine tectonic juxtaposition and metamorphic reworking of slices of Variscan continental crust and its Triassic cover, locally separated by thin layers of calcschists and meta-ophiolites of the Piemonte Zone, these last derived from the Mesozoic Tethys ocean floor.

Most of these slices are composed of the same continental lithologic association, but are characterized by different Alpine climax overprints (e.g. Compagnoni & Rolfo, 2003, with refs.; Compagnoni et al., 2012). The following tectonometamorphic units have been distinguished in the southern DMM (Fig. 14), from the lower to the upper structural levels:

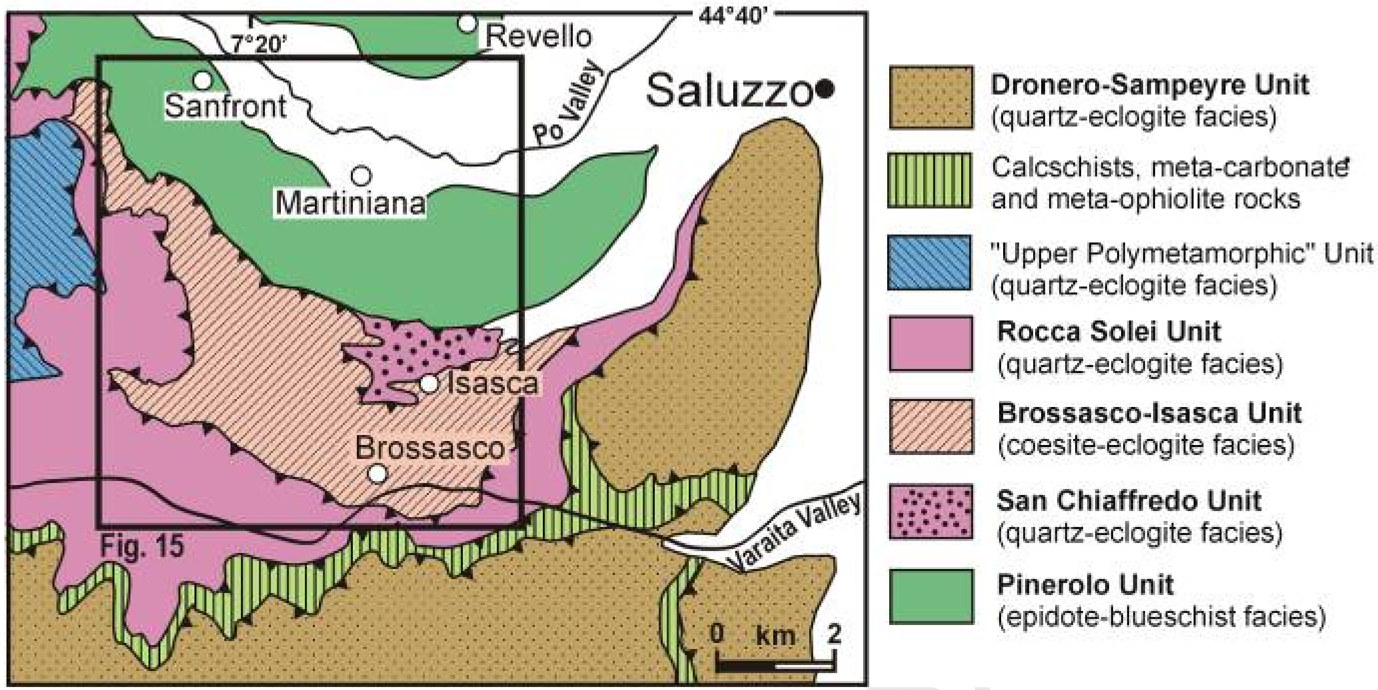


Fig. 14 – Tectonic sketch map of the southern Dora-Maira Massif (modified from Compagnoni et al. (2012) (see Fig. 1 for the location in the framework of the Western Alps). Tectonic units of the Massif are listed from the structurally lowermost Pinerolo Unit to the structurally highest Dronero-Sampeyre Unit.

(i) the <u>Pinerolo Unit</u> consists of graphitic micaschists locally garnet- and chloritoid- bearing, with intercalations of quartzite and minor paragneiss including lenses of metabasite. During the Alpine orogeny these rocks experienced peak metamorphism under epidote-blueschist-facies conditions (*T* ~ 400 °C, *P* ~ 0.8 GPa: Chopin et al., 1991; Avigad et al., 2003).

(ii) the <u>San Chiaffredo Unit</u> is a portion of pre-Alpine continental crust consisting of a Variscan amphibolite-facies metamorphic basement (now micaschist and eclogite) intruded by Permian granitoids (now fine-grained orthogneiss and augen-gneiss). During the Alpine orogeny this unit experienced peak metamorphism under quartz eclogite-facies conditions (*T* ~ 550 °C, *P* = 1.5 GPa).

(iii) the <u>Brossasco-Isasca Unit</u> (BIU) is a portion of pre-Alpine continental crust consisting of a Variscan amphibolite-facies metamorphic basement (now garnet-kyanite micaschist, impure marble and eclogite) intruded by Permian granitoids (now mainly augen-gneiss and minor metagranite and pyrope-bearing whiteschist). Peak metamorphic conditions in the coesite eclogite-facies within the diamond stability field (T ~ 730 °C, P = 4.0-4.3 GPa) were experienced during the Alpine orogeny.

(iv) the Rocca Solei Unit is a portion of pre-Alpine continental crust similar to the BIU and the San Chiaffredo Unit, consisting of a Variscan amphibolite-facies metamorphic basement (now garnet- and chloritoid-bearing micaschist and micaceous gneiss, marble and eclogite) intruded by Permian granitoids (now fine-grained orthogneiss and augen-gneiss with minor metagranitoid). During the Alpine orogeny this unit experienced peak metamorphism under quartz-eclogite facies conditions (T ~ 550 °C, P ~ 1.5 GPa: Chopin et al., 1991; Matsumoto & Hirajima, 2000).

All these tectonometamorphic units have been overprinted by a late Alpine greenschist - facies recrystallisation, which pervasively reworked and extensively obliterated the former HP or UHP metamorphic mineral assemblages.

The coesite-bearing Brossasco-Isasca Unit

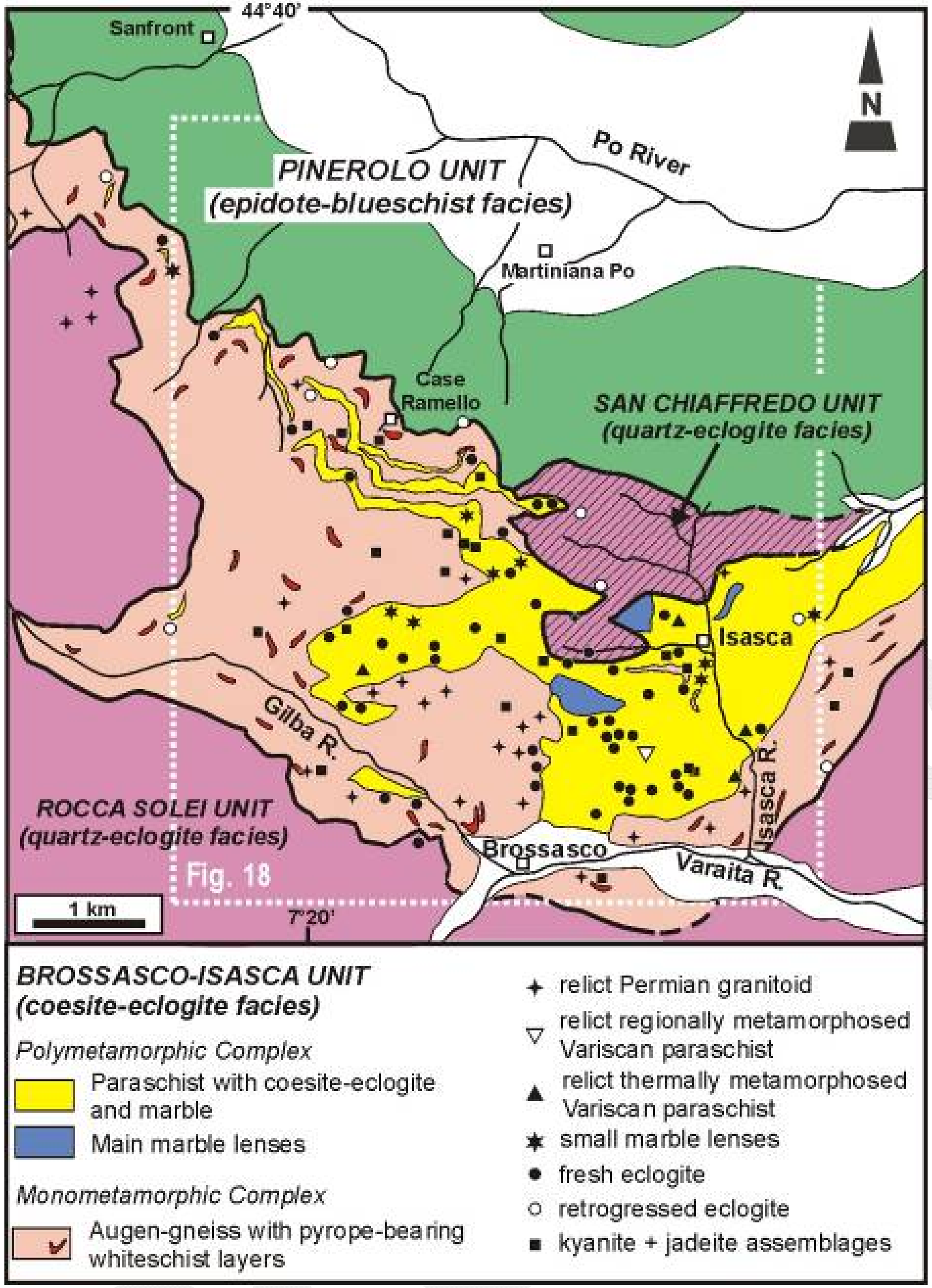
The coesite-bearing Brossasco-Isasca Unit is about 10×4×1 km in size and is tectonically sandwiched between the overlying Rocca Solei Unit and the underlying Pinerolo Unit, which are characterized by Alpine quartz eclogite-facies and epidote blueschist-facies conditions, respectively. Areal extent, internal setting and geologic interpretation of the BIU, mainly derived from field and laboratory studies of Compagnoni & Hirajima (1991), Turello (1993), Compagnoni et al. (1995), Groppo (2002), Compagnoni & Rolfo (2003), Compagnoni et al. (2004) and Compagnoni et al. (2012) are reported in Fig. 15. A new geologic map including the whole BIU has been recently published (Compagnoni et al., 2012).

Two lithostratigraphic complexes have been distinguished in the BIU (Compagnoni et al., 1995): a "Monometamorphic Complex" derived from the Alpine tectonic and metamorphic reworking of Permian granitoids, and a "Polymetamorphic Complex", derived from the Alpine reworking of a Variscan amphibolite-facies metamorphic basement. The two complexes roughly correspond to the "gneiss and metagranite" and to the "varied formation" of Chopin et al. (1991). Relics of pre-Alpine metamorphic assemblages and of pre-Alpine intrusive contacts (Fig. 15) located all along the boundary between the Monometamorphic and Polymetamorphic Complexes (Biino & Compagnoni, 1992; Compagnoni & Hirajima, 1991; Compagnoni et al., 1995), allowed inferring the relationships between the two complexes before the Alpine orogeny (Fig. 16).

The Monometamorphic Complex
It mainly consists of augen-gneiss grading to medium- to fine-grained orthogneiss, which locally preserves relics of the granitoid protolith. This gneissic complex includes layers of pyrope-bearing whiteschist from a few dm- to tens of m- thick (Fig. 15) that are considered to be metasomatic rocks derived from the granitoid protolith (Compagnoni & Hirajima, 2001; Ferrando et al., 2009; Ferrando, 2012); further details on the whiteschist are given in the description of Stop 2.3.
Rare layers of garnet-jadeite-kyanite-quartz granofels are also included within the orthogneiss (e.g. Compagnoni et al., 1995; Compagnoni & Rolfo, 2003). The UH*P* peak assemblage in this granofels consists of coesite, pyrope-rich almandine, jadeite (often replaced by fibrous albite + paragonite + hematite + magnetite), kyanite, very minor phengite and rare colourless glaucophane. Coesite is found as inclusion in garnet, jadeite and kyanite. This unusual lithotype has been considered by Schreyer et al. (1987) and Sharp et al. (1993) as the possible product of partial melting of the host whiteschist protolith, but convincing evidence for this interpretation is still lacking (see also Grevel et al., 2009).
The gradual transition from undeformed metagranitoid to foliated orthogneiss indicates that the latter derived from the Alpine tectonic and metamorphic reworking of the former (Biino & Compagnoni, 1992), i.e. the orthogneiss are monometamorphic rocks. A Permian age of ca.275 (U/Pb SHRIMP on zircons) is the most probable intrusion age for the granitoid protolith (Gebauer et al., 1997 with discussion on previous radiometric data).

The Polymetamorphic Complex
This complex includes paragneiss with eclogite and marble intercalations (e.g. Compagnoni et al., 1995, 2004), locally preserving relics of pre-Alpine mineralogies and structures (Fig. 15). A detailed description of different lithologies of the Polymetamorphic Complex is given in Compagnoni et al., 1995; here we focus on a few lithologies preserving the most interesting UHPM mineral assemblages and/or pre-Alpine relics.

Kyanite-almandine paraschist
The most typical lithology of the Polymetamorphic Complex is a brownish micaschist with porphyroblasts of kyanite and almandine, up to 2 cm across. Microscopically, the rock consists of phengite, garnet, kyanite, quartz, jadeite, abundant accessory rutile and aggregates of yellow-greenish tourmaline. A second generation of acicular kyanite, most likely developed during retrograde decompression, locally occurs at the rim of kyanite porphyroblasts. Jadeite is generally replaced by very fine-grained intergrowths of albite + paragonite,



almandine is partly altered to biotite and/or chlorite, and kyanite is partly replaced by paragonite and locally by an earlier margarite corona.

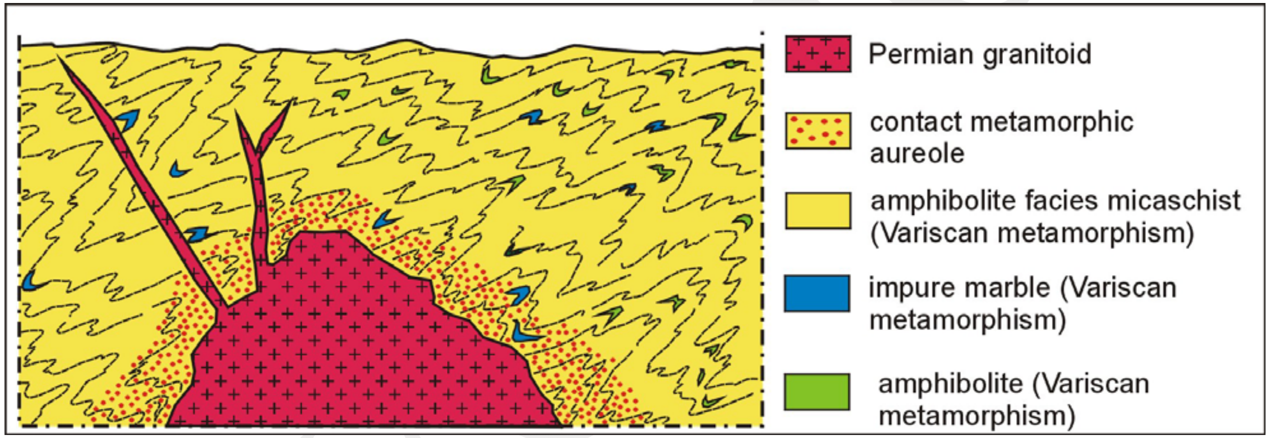
Porphyroblastic kyanite and almandine are usually crowded with mineral inclusions, which locally define a rough planar internal foliation. Almandine usually includes chloritoid, rutile, staurolite, kyanite, jadeite, quartz and coesite, chlorite and white micas. Kyanite inclusions may be found everywhere in the garnet.

On the contrary, staurolite, chloritoid and quartz occur only in the garnet core and mantle, whereas coesite is found only in the garnet rim. Distribution of mineral inclusions in garnet and its chemical zoning (increasing pyrope content from core to rim) point to a growth during a prograde *P-T* trajectory from the quartz to the coesite stability field.

Detailed investigations of the phengite (a zoned 3T-polytype: Ferraris et al., 2000) revealed the occurrence of talc + quartz nano-scale "exsolutions" that likely formed during exhumation in the quartz stability field, by a reaction which produced a less celadonite-rich phengite.

Fig. 15 - Simplified geological map of the central portion of the coesite-bearing "Brossasco-Isasca Unit" and relationships with the other adjoining units, labelled in the map (modified from Compagnoni & Rolfo, 2003 and Castelli et al., 2007, see Fig. 14 for location).

Fig. 16 - Pre-Alpine relationships among the BIU lithologies as inferred from the structural and mineralogical relics escaped to the Alpine polyphase pervasive deformation (modified from Compagnoni et al., 2012). Permian granitoids (protolith of the Monometamorphic Complex) were emplaced into a Variscan amphibolite-facies metamorphic basement consisting of micaschists, marbles and amphibolites (protolith of the Polymetamorphic Complex).



Eclogite

Within the basement paraschist, numberless small bodies of eclogites occur, which appear to derive from boudinage of original concordant layers. At the eclogite-paraschist contact, cm-thick discontinuous brownish-red garnetites frequently occur. The eclogite is very fine-grained and usually contains idioblastic garnet, lineated omphacite, phengite with well-developed preferred orientation, quartz (as polygonal granoblastic aggregates), local zoisite, accessory rutile and minor apatite, zircon, sulphides and local graphite.

Some light green quartz-rich eclogites also contain kyanite. Most eclogites exhibit a faint foliation defined by the alignment of the rutile grains. The occurrence of small coesite inclusions in the omphacite suggests that the granoblastic quartz aggregates derive from former coesite. In the retrograded eclogites, omphacite is typically replaced by a fine-grained symplectitic intergrowth of diopsidic clinopyroxene and albite ± green amphibole, and phengite is partially replaced by symplectitic aggregates of biotite + oligoclase (Groppo et al., 2007a). Often, the eclogites are cut by coarse-grained discordant veins consisting of idioblastic omphacite, minor interstitial quartz and accessory apatite. In other eclogites, discontinuous layers or pockets of coarse-grained quartz, omphacite and accessory apatite are common, which appear to have derived from folding and transposition of former discordant metamorphic veins.

Marble

The marble consists of both calcite- and dolomite-rich layers locally with thin discontinuous interlayers of micaschist and boudins of eclogite. The grain-size is usually coarse, but very fine-grained mylonitic types locally occur, reflecting different stages of the polyphase Alpine tectonometamorphic evolution (e.g. Compagnoni & Rolfo, 2003). At Costa Monforte (where the largest marble lens occurs: Fig. 15), the most common type is a pure phengite-bearing calcite marble, in which white mica is a chemically-zoned, highly-ordered 3T phengite (Ferraris et al., 2005). This marble is locally characterized by silicate-rich (garnet, clinopyroxene, phengite, rutile, epidote, and titanite) domains stretched parallel to the regional UHP foliation. Microstructural, petrological and crystallographic data show that the marbles underwent the same Alpine tectonometamorphic evolution of the country paragneiss (Ferraris et al., 2005; Castelli et al., 2007).

Relics of pre-Alpine paragneiss and paraschist

In the paraschists, pre-Alpine relics are generally rare. The most common mineralogical relict of the pre-Alpine history is a porphyroblastic garnet up to 2 cm in diameter, overgrown by a thin corona of small idioblastic Alpine garnets, these last including small needles of rutile with a "sagenitic" arrangement (Groppo et al., 2006; Spiess et al., 2007).

The best-preserved remnant of a pre-Alpine metamorphic rock, mostly characterised by pseudomorphous and coronitic reactions, has been recognized at Case Garneri, halfway between Brossasco and Isasca. Macroscopically, this exceptionally well preserved rock is very dark and includes irregular cm- to dm-sized light portions of metatects.

Under the microscope, the dark portions consist of relict garnet, nematoblasts of former sillimanite pseudomorphically replaced by kyanite aggregates, red-brown biotite partly replaced by phengite ± rutile and surrounded by new coronitic garnet, lens-like aggregates of patchy albite + zoisite ± kyanite with jadeite relics after former plagioclase, and polygonal granoblastic quartz aggregates. Such mineral relics and pseudomorphs are embedded in a dark grey matrix, mainly consisting of fine-grained garnet + kyanite ± phengite ± biotite, interpreted as derived from former contact metamorphic cordierite. Accessory minerals are very abundant: graphite, ilmenite completely replaced by rutile and surrounded by a garnet corona, apatite, zircon, monazite and tourmaline. In some SiO_2-undersaturated sites, pseudomorphs consisting of corundum ± rutile aggregates, surrounded by an inner kyanite and an outer garnet corona occur, which may have derived from former spinel. The rough pre-Alpine rock foliation is defined by the preferred orientation of the graphite flakes and sillimanite nematoblasts and by the preferred dimensional orientation of the former plagioclase blasts. Pre-Alpine garnet contains inclusions of plagioclase, quartz, biotite, rare sillimanite, and accessory apatite, tourmaline, and dusty graphite. This fine-grained graphite usually defines a closely spaced internal foliation

often discordant with respect to the external foliation. Pre-Alpine garnet, which usually exhibits a strongly corroded shape and appears extensively transformed into contact cordierite, is in turn replaced by a dark grey aggregate of garnet and minor kyanite (Table 1).
The light portions locally preserve an igneous microstructure and consist of plagioclase replaced by jadeite (usually albitized) and zoisite, K-feldspar partly replaced by phengite + quartz intergrowths, granoblastic polygonal quartz aggregates, cordierite completely replaced by garnet and kyanite, and minor biotite and muscovite (Table 1). Graphite is ubiquitous and locally occurs as large deformed flakes. Around the leucocratic portions, biotite aggregates and kyanite aggregates after a former prismatic contact andalusite locally occur. Coesite has not been found: however, the ubiquitous presence of granoblastic polygonal quartz aggregates, locally pseudomorphous after a pre-Alpine single quartz crystal, is considered to be the unambiguous evidence of the former occurrence of early-Alpine coesite.

Variscan Regional metamorphism	***Permian contact metamorphism***	***Alpine UHP metamorphism***	***Alpine retrogression***	***Type of replacement***
Grt			Bt + Chl	pseudomorphous
	[Crd]	Grt + Ky	Ser + Chl	pseudomorphous
Bt	Bt	Grt + Phg	Bt + Chl	coronitic
Kfs	Kfs	Kfs	microperthite	pseudomorphous
		(+ H_2O) → Phg ± Qtz		coronitic
[Pl]	[Pl]	(+ H_2O) → Jd + Qtz + Zo + Ky	Ab/Olg + Czo + *Wm*	pseudomorphous
[Sil]		Ky	Ser	pseudomorphous
	[And]	Ky	Ser	pseudomorphous
[Qtz]	[Qtz]	[Coe]	Qtz	pseudomorphous
Ms	Ms			
	[Spl?]	Grt + Crn	Ser	pseudomorphous
[Ilm]	[Ilm]	Rt + Grt	Ttn + Chl	coronitic
Mnz		Ep	Ep	
Ap, Gr, Zrn, Tur				

Table 1 – Metamorphic evolution of the paraschist from the Polymetamorphic Complex.

On the basis of mineral compositions and microstructure, the inferred polymetamorphic evolution includes: (i) a pre-Alpine, most likely Variscan, medium-P regional metamorphic event (in the sillimanite stability field), which reached the upper amphibolite-facies and developed the leucocratic metatects; (ii) a pre-Alpine low-*P* static recrystallization, most likely related to the thermal metamorphism connected to the intrusion of the Permian granitoids, which led to the widespread development of contact biotite, andalusite, cordierite; (iii) a peak Alpine UH*P* static recrystallization, characterised by pseudomorphous and coronitic reactions involving the growth of coesite, jadeite, garnet, zoisite, phengite, kyanite, and rutile; (iv) a late Alpine greenschist facies retrogression, responsible for the complete coesite inversion to polygonal granoblastic quartz and the replacement of jadeite by albite (Table 1).

Relics of pre-Alpine assemblages in marble

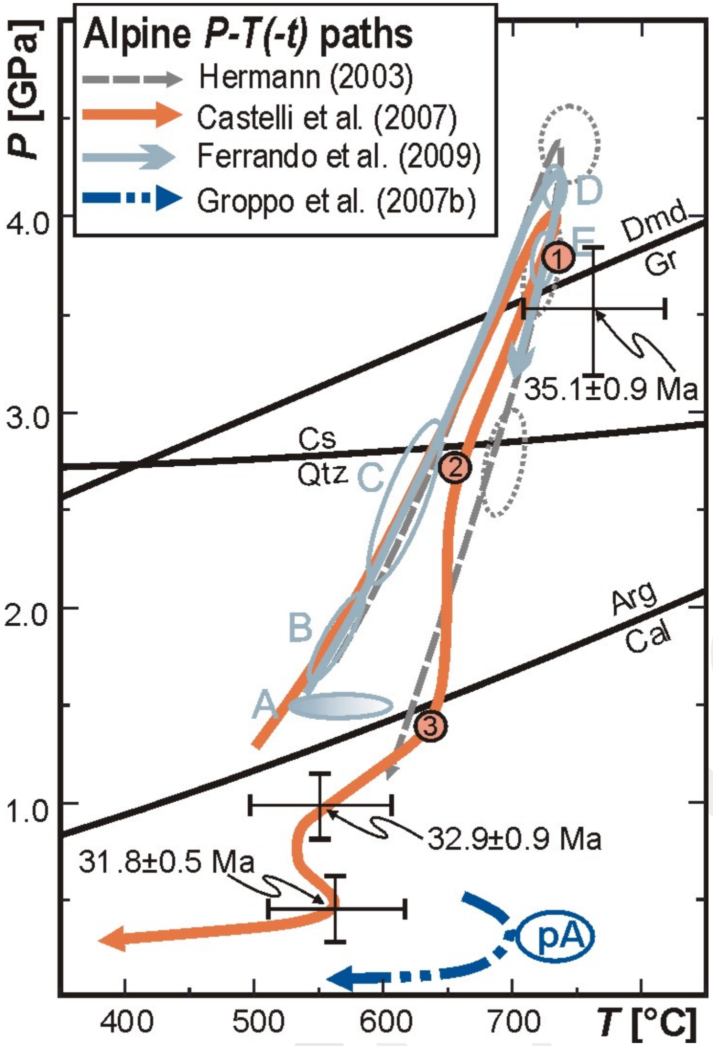
In the Costa Monforte marble lens, a layer of dolomite marble retains scanty silicate-rich domains broadly aligned parallel to the regional foliation that still preserves relics of a pre-Alpine paragenesis only partially overprinted by the Alpine metamorphic assemblages (Castelli et al., 2007). The fabric is characterized by irregular to lens-shaped aggregates (up to a few cm-thick) of spinel + ilmenite + dolomite + corundum + chlorite ± calcite, set in a matrix of fine- to medium-grained chlorite + calcite ± dolomite; minor högbomite $[(Mg,Fe^{2+})_3\ (Al,Ti)_8\ O_{10}\ (OH)]$, margarite and rutile also occur. The relict bluish-green spinel occurs as medium- to coarse-grained, fractured and corroded, porphyroclasts which locally include roundish crystals of ilmenite. Porphyroclastic dolomite does not display calcite exsolution, but is only partially replaced by calcite at the rim and along fractures.

Microstructural relationships and petrological data allow inferring a partially preserved pre-Alpine mineral assemblage consisting of spinel + ilmenite + dolomite + calcite (+ plagioclase, inferred on the basis of petrological modelling) (Groppo et al., 2007b). This high-*T* and low-*P* assemblage was re-equilibrated into a peak Alpine assemblage consisting of corundum + chlorite (clinochlore) + rutile + dolomite + aragonite (now completely inverted to calcite). During the UH*P* metamorphism, spinel was pseudomorphically replaced by fine-grained aggregates of bluish corundum + clinochlore; coarser-grained corundum locally developed at the rim of Crn + Chl pseudomorphs after spinel. Pre-Alpine ilmenite was replaced by aggregates of rutile. The inferred pre-Alpine plagioclase was likely replaced by UH*P* zoisite, but no evidence of this phase has been observed in the marble, possibly because it was consumed by lower-*P* reactions. These reactions, which developed at the expense of both the high-*T* pre-Alpine and the UH*P* Alpine assemblages, were responsible for the growth of calcite inverted from aragonite, högbomite and a new generation of ilmenite after spinel +

corundum + rutile, sheridanite overgrowth after clinochlore and radial aggregates of margarite around the fine-grained aggregates of corundum + clinochlore.

The *P-T* evolution of the Brossasco-Isasca Unit

The whole metamorphic evolution of the Brossasco-Isasca Unit is summarized in the *P-T* diagram of Fig. 17.



The oldest event, recognized in the paraschists of the Polymetamorphic Complex, is a pre-Alpine polyphase amphibolite-facies regional metamorphic recrystallization, most likely Variscan in age. The occurrence of metatects in paragneiss suggests that during the regional Variscan metamorphism, *T* higher than the wet granite *solidus* were reached (*T*~700°C at *P*~0.3 GPa: Compagnoni et al., 1995). This regional event is overprinted by a high-*T*, low-*P* contact recrystallization connected to the emplacement of the Permian granitoids (Compagnoni & Hirajima, 1991; Compagnoni et al., 1995; Compagnoni & Rolfo, 2003).

Fig. 17 - Synopsis of selected Alpine *P-T* estimates and inferred Alpine and pre-Alpine (blue curve: pA) *P-T* paths for the BIU Polymetamorphic Complex (modified from Castelli et al., 2007). The solid, orange curve is the Alpine *P-T* path proposed by Castelli et al. (2007) that integrates the peak and retrograde *P-T* estimates by Groppo et al. (2007b: circles labelled 1 to 3). Solid ellipses labelled A to E are the P-T conditions of prograde, peak and retrograde stages of the BIU, respectively, estimated by Ferrando et al. (2009), whereas dotted ellipses are those estimated by Hermann (2003). The error bars of ages refer to the geochronologically dated *P-T* estimates of Rubatto & Hermann (2001). The dotted-dashed blue path (pA) is the pre-Alpine *P-T* path of the BIU from Groppo et al. (2007a).

The *P-T* conditions of the pre-Alpine events are compatible with the relict mineral assemblage preserved in the spinel-ilmenite-bearing dolomite marble, for which a binary fluid at $X(CO_2) > 0.60$ has been modeled for high-*T* decarbonation reaction(s) along its prograde pre-Alpine path (Groppo et al., 2007b).

From whiteschists, three prograde Alpine stages have been inferred, corresponding to: ca. 1.6 GPa, ≤600°C; 1.7–2.1 GPa, 560–590 °C; and 2.1<*P*<2.8 GPa, 590<*T*<650°C, respectively (Ferrando et al., 2009). As to the UH*P* climax, while the inferred peak *T* (750 ± 30°C) little changed from the first estimate (Chopin, 1984), the peak pressure progressively increased from the original value of Chopin (1984: $P \geq 2.8$ GPa) to the 3.6 GPa of Nowlan et al. (2000) and to the almost 4.5 GPa determined by Hermann (2003) on the basis of experimental studies. Pressures of about 4.0-4.3 GPa within the diamond stability field have been confirmed by recent petrologic studies on both a garnet-clinopyroxene-phengite marbles (Castelli et al., 2007) and whiteschists (Ferrando et al., 2009).

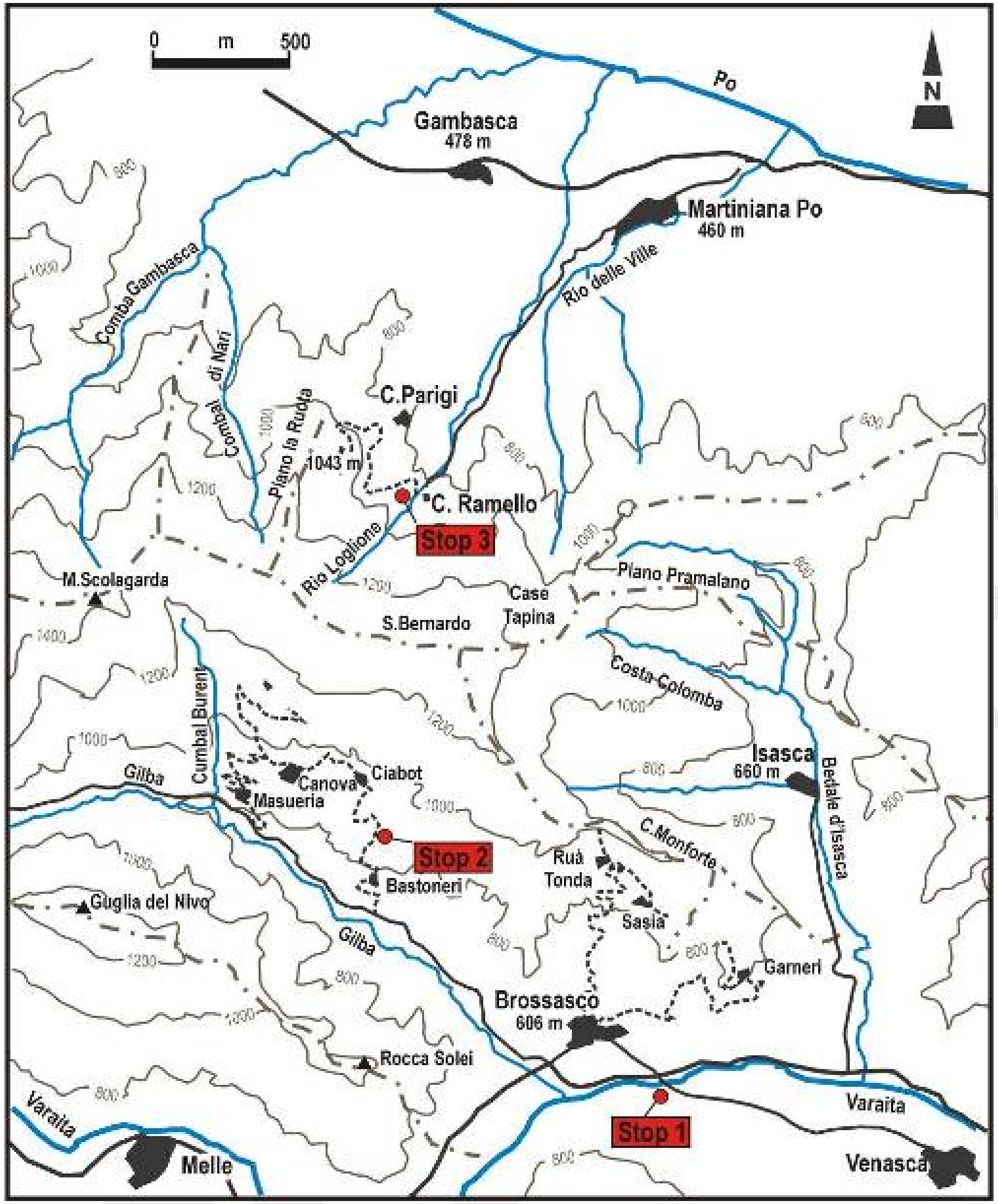
The absence of micro-diamond in the metapelites has been explained by Hermann (2003) by the low temperature of peak metamorphism, the absence of a free fluid phase and the short residence time in the diamond-facies conditions, resulting in a sluggish kinetics of the transition of graphite to diamond. In the marbles, Castelli et al. (2007) show that the lack of micro-diamond is likely related to the very-low $X(CO_2)$ values of the coexisting fluid phase at UH*P* conditions, which favoured relatively high $f(O_2)$ conditions.
The evolution following the *P*-climax is marked by a significant decompression coupled with a moderate cooling (Chopin et al., 1991; Schertl et al., 1991; Compagnoni et al., 1995; Nowlan et al., 2000). Two significant *P-T* points along the retrograde path (points 2 and 3 in Fig. 17) have been estimated by thermodynamic modelling from different bulk compositions at the coesite/quartz and aragonite/calcite transition, respectively (Groppo et al., 2006; 2007a; Castelli et al., 2007). The whole Alpine evolution is, therefore, defined by a clockwise *P-T* path, characterized by an extremely close loop (Fig. 17). A moderate *T* increase is shown at low-pressure, which likely corresponds to the second greenschist-facies metamorphic event (Compagnoni et al., 1995; Borghi et al., 1996).

The age of the Alpine UH*P* metamorphism

The UH*P* metamorphism of the BIU was long believed to be Cretaceous in age (e.g. Moniè & Chopin, 1991), as carefully discussed in the review by Di Vincenzo et al. (2006). However, the most recent geochronological data support a different scenario. The first results suggesting a Tertiary age for the UH*P* metamorphism have been obtained by Tilton et al. (1989, 1991) who proposed that the growth of the pyropes from the UH*P* whiteschists occurred at 38–40 Ma. This interpretation was based on: conventional U–Pb zircon data, yielding in a concordia diagram a lower intercept age of 38.0 ± 2.8 (2σ) Ma; Sm–Nd garnet results from seven samples, yielding an errorchron with an apparent age of 38.3 ± 4.5 (2σ) Ma; and U–Pb data on pyrope, ellenbergerite, monazite and white micas, giving apparent U–Pb and Th–Pb ages of ~31 to ~35 Ma, respectively. Gebauer et al. (1997) further supported these results constraining the age of the metamorphic zircon domains at 35.4 ± 1.0 Ma. This age was also confirmed by Duchêne et al. (1997), who reported a Lu–Hf age of 32.8 ± 1.2 Ma, obtained on a garnet–whole-rock pair from a pyrope whiteschist.

The *P*–*T*–*t* path of the BIU was later refined by Rubatto & Hermann (2001) using SHRIMP *in situ* dating of single growth zones of titanite grains from two calc-silicate nodules in marbles. Titanite domains considered to have developed during the UH*P* stage yielded a mean age of 35.1 ± 0.9 Ma, those developed during exhumation at ~35 km a mean age of 32.9 ± 0.9 Ma, and those crystallized at ~17 km a mean age of 31.8 ± 0.5 Ma. Combining these radiometric data with a zircon fission track age of about 30 Ma (Gebauer et al., 1997), and making reasonable assumptions about the conversion of pressure to depth, maximum exhumation rates of 3.4 cm/year have been estimated, thus allowing the reconstruction of one of the best-documented exhumation paths for UHP terrains. These rates imply that exhumation acted at plate tectonic speeds similar to subduction (Rubatto & Hermann, 2001), and was significantly faster than erosion, thus suggesting that exhumation is driven by a combination of tectonic processes involving buoyancy and normal faulting.

DAY 2 - Itinerary



During the excursion we will focus on three stops that are located on the right side of the southernmost Val Varaita (Stop 2.1), on the left side of Vallone di Gilba (Stop 2.2), and on the right side of Valle Po (Stop 2.3), respectively (Fig. 18).

Fig. 18 - Location of the geologic Stops in the Brossasco-Isasca Unit, between Valle Po and Val Varaita (see Fig. 15 for location).

Route: From Revello by minivan to Brossasco, Val Varaita

Stop 2.1: Metagranite recrystallized under UH*P* conditions with greenschist-facies shear zone
Brossasco, hillrock near the bridge on Torrente Varaita (*Val Varaita*)
(Coord: N 44° 34' 00''   E 7° 22' 05'', elev. 590 m a.s.l.)

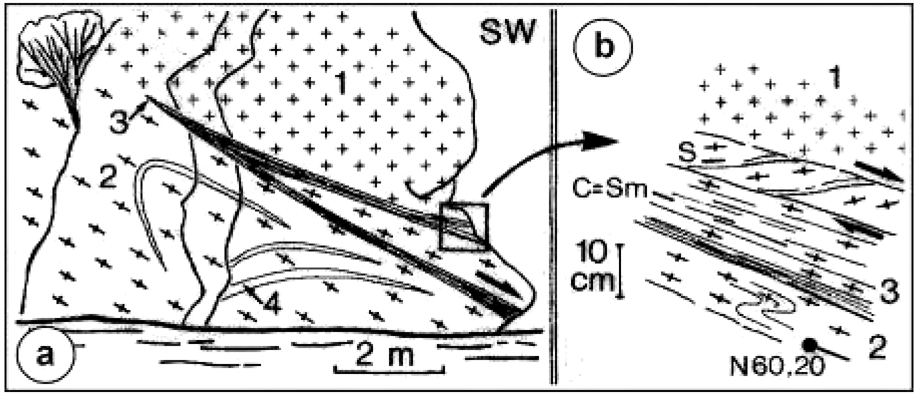


Fig. 19 – Stop 1: Sketch of the metagranite outcrop (a) and a detail (b) at the northern vertical cliff of the hillrock near the bridge on Torrente Varaita, Brossasco (from Michard et al., 1995), characterized by an heterogeneous deformation. (1) undeformed porphyritic metagranite; (2) mylonitic orthogneiss with passively folded quartz veins (4) and minor isoclinal folds (detail); (3) ultramylonite (top-to-the-SW S/C structures in the transitional zone 1-3); Sm: incipient greenschist-facies foliation.

The metagranite is exposed along the cut of a country road running westwards parallel to the river (Torrente Varaita) (Fig. 19).
The best preserved metagranite is porphyritic (Fig. 20a) and locally contains aplitic and pegmatitic dykes, cognate microgranular enclaves and centimetric xenoliths of biotite-rich schists. The massive metagranite is dissected by a dm-thick shear zone, which shows a well-developed mylonitic foliation and a significant grain-size reduction, accompanied by a general pervasive greenschist-facies recrystallization (Fig. 19). The transition from the massive metagranite to the surrounding augengneiss is locally gradual and occurs over a distance from a few centimetres to several metres by progressive development of a metamorphic foliation.
The metagranite is surrounded on all sides by the augengneiss of the Monometamorphic Complex, which is characterized by a greenschist-facies metamorphic overprint. This suggests either that most of the augengneiss deformation was acquired during the late-Alpine event or that an earlier, most likely early-Alpine, eclogitic foliation was penetratively reworked during the greenschist event. In the surrounding augengneiss the only evidence for the former UH*P* metamorphism is the local occurrence of a high-celadonite core in the phengite porphyroclasts, wrapped around by the main Alpine greenschist facies foliation, and the probable original occurrence of the assemblage grossular-rich garnet + rutile (now titanite: Chopin et al., 1991).

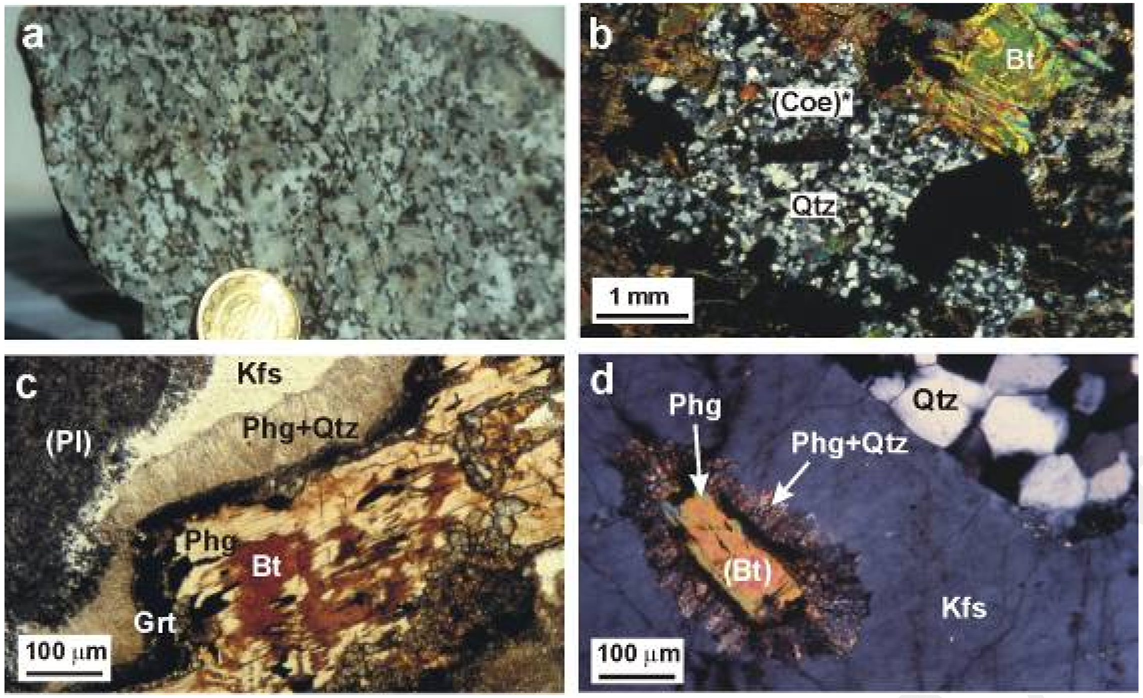
Under the microscope, the metagranite exhibits an igneous hypidiomorphic microstructure (Fig. 20b), only partially obliterated by the coronitic and pseudomorphous reactions of the Alpine metamorphic recrystallizations. The original igneous assemblage consisted of quartz, plagioclase, K-feldspar, biotite and accessory apatite, tourmaline and zircon. Garnet xenocrysts are locally observed. Except for K-feldspar, the igneous minerals were completely replaced by Alpine metamorphic phases (Table 2).

Igneous minerals	***Alpine UHP peak metamorphism***	***Alpine retrogression***	***Type of replacement***
Qtz	[Coe]	polygonal Qtz aggr.	pseudomorphous
Kfs	Kfs	microperthite	pseudomorphous
	$(+H_2O) \rightarrow$ high-Cel dactylitic Ms + [Coe]	Qtz	coronitic
[Pl]	Jd + (Coe) + Zo + Ky + Kfs	Ab + Czo	pseudomorphous
Bt	Mg-rich Bt + Rt + Ti-rich Phg	Bt + Ttn	pseudomorphous
Bt / Qtz	Phg (after Bt)		coronitic
	poorly-zoned Grt corona	Bt	
Bt / Kfs	Phg (after Bt)	Bt	coronitic
	continuous Grt corona		
	Grt + dactylitic Qtz corona		
	Phg + dactylitic Qtz corona (after Kfs)		
Bt / Pl	asymmetrically zoned Grt corona	Bt	coronitic
	Grt + Jd corona	Bt + Ab	
[Ti-(Fe)-rich phase]	Rt	Ilm / Ttn	pseudomorphous
Ap, Zrn, Tur			
Grt (xenocryst)			

Table 2 – Metamorphic evolution of the metagranite from the Monometamorphic Complex.

The *igneous biotite* is locally preserved as a textural relic, but its Mg-rich chemical composition indicates a metamorphic equilibration (Biino & Compagnoni, 1992; Bruno et al., 2001). Usually, biotite is pseudomorphically replaced by a high-celadonite Ti-rich phengite + rutile and surrounded by a continuous garnet corona (Fig. 20c,d). At the contact with the original igneous plagioclase site, the coronitic garnet is asymmetrically zoned with pyrope- and almandine-rich composition on the biotite side and grossular-rich composition on the plagioclase side.

The *igneous plagioclase* is pseudomorphically replaced at the UHP metamorphic peak by jadeite (up to $Jd_{90}CaEsk_{10}$) (Bruno et al., 2002) + zoisite + coesite (now quartz) + kyanite + K-feldspar (Fig. 20c). Because of the



general pervasive greenschist-facies overprint, jadeite is mostly retrogressed to albite + white mica. Fresh jadeite is usually preserved only in the pseudomorphs after igneous plagioclase, armoured within the igneous K-feldspar. Within such pseudomorphs, a small grossular-rich idioblastic garnet is developed, instead of zoisite, close to the biotite site.

Relics of *igneous K-feldspar*, up to 3 cm long, are preserved. They include pseudomorphs after plagioclase, biotite and quartz (Fig. 20d). Along deformation bands or around mineral inclusions, K-feldspar recrystallizes to polygonal granoblastic aggregates.

Fig. 20 – Stop 1: Metagranite recrystallized under UH*P* conditions. See also Table 2. (a) Hand specimen of metagranite: the only evidence of the UH*P* metamorphic overprint is a very rough foliation and the sugary appearance of the original igneous quartz, now consisting of a granoblastic polygonal aggregate of metamorphic quartz, derived from inversion from former coesite. The coin is 20 mm across. (b-d) Representative microstructures of the metagranites. (b) Fine-grained granoblastic aggregate of quartz, statically derived from inversion of coesite which replaced the igneous quartz during UH*P* metamorphism. Sample DM 496. XPL. (c) Igneous biotite replaced by a Mg-richer biotite + phengite. The biotite/phengite domain is surrounded by a first continuous garnet corona and a second corona against K-feldspar, which consists of a very fine-grained dactylitic intergrowth of phengite and quartz. Sample DM1086. PPL. (d) Igneous biotite included in K-feldspar, pseudomorphically replaced by a single phengite crystal and surrounded by a thin garnet corona and a dactylitic phengite + quartz intergrowth. In the upper right, a granoblastic polygonal quartz aggregate from former peak coesite, developed after primary euhedral igneous quartz. Sample DM 60. XPL. Modified from Compagnoni et al. (2004).

Around the biotite included in K-feldspar, a second reaction rim surrounds the garnet corona, which consists of an intergrowth of vermicular (dactylitic) quartz and highly celadonitic phengite growing from the garnet corona towards K-feldspar (Fig. 20c,d).

The *igneous quartz* is always replaced by a polygonal granoblastic aggregate (Fig. 20b,d), interpreted as derived from coesite inversion (Biino & Compagnoni, 1992; Compagnoni et al., 1995).
The polyphase metamorphic evolution recognized in the metagranite is summarized in Table 2 (Biino & Compagnoni, 1992; Compagnoni et al., 1995; Compagnoni & Rolfo, 2003).

Route: By minivan from Brossasco to Vallone di Gilba uphill to Case Bastoneri. Then 20 minute walk to the outcrop

Stop 2.2: Metagranites with xenoliths of paraschists and metabasics: relict intrusive contact between Permian granitoids and the country metamorphic basement.

North of Case Bastoneri, left side of Vallone di Gilba
(Coord: N 44° 35' 01'' E 7° 20' 27'', elev. 885 m a.s.l.)

A poorly deformed pre-Alpine intrusive contact between the Variscan basement (now the Polymetamorphic Complex) and the Permian granitoids (now the Monometamorphic Complex) is preserved. The blocks consist of a porphyritic metagranite with cm-sized K-feldspars, which includes numberless xenoliths of banded metamorphic rocks (Fig. 21). The xenoliths are composed of both metapelites and metabasics, which locally preserve a pre-intrusion folded foliation. Where the xenoliths are finely dismembered, fabric, mineralogy and mode of the igneous rock were modified: this suggests that a significant assimilation of the country metamorphic rocks by the intruding magma did occur.

Metagranitoids

The massive metagranite consisted of euhedral plagioclase, biotite, quartz, poikilitic K-feldspar phenocrysts (Fig. 21b), late interstitial quartz, and accessory apatite, zircon and rare tourmaline. Close to the biotite-rich metapelite xenoliths (see below), a rim enriched in euhedral plagioclase and minor quartz, usually occurs. The igneous minerals exhibit the same pseudomorphous and coronitic microstructures (Fig. 22a) described for the Brossasco metagranite (see Stop 2.1).

Xenoliths

The xenoliths display a variety of UH*P* metamorphic fabrics, grain-sizes and mineral assemblages. However, in spite of the extensive Alpine recrystallization, in many samples relict minerals and ghost coronitic and pseudomorphous microstructures still occur, which are useful to infer the pre-Alpine mineralogy and evolution.

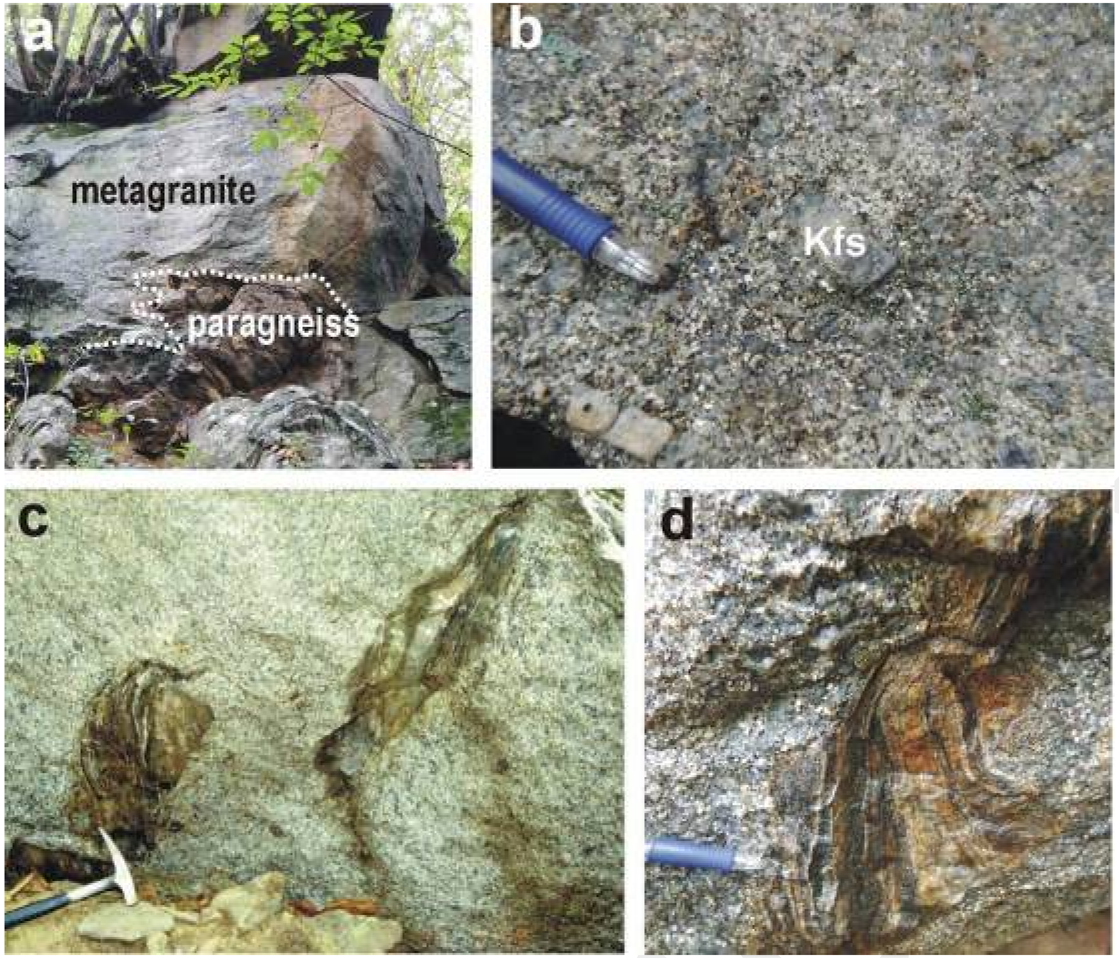
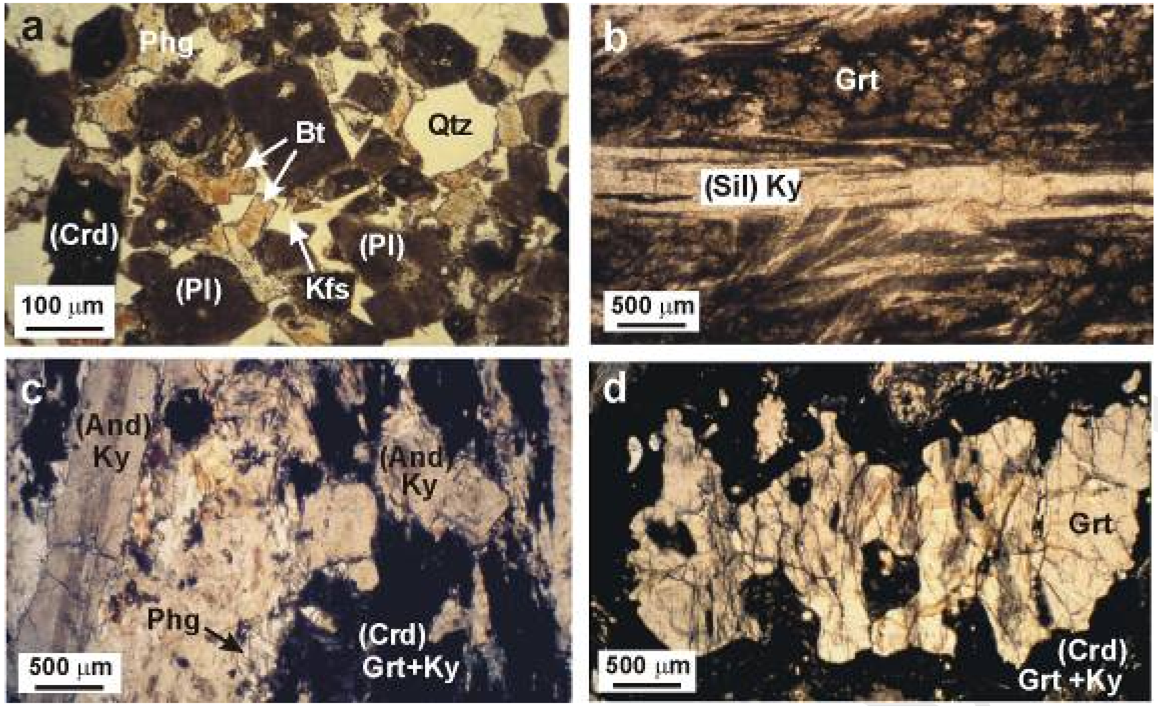


Fig. 21 – Stop 2: Metagranites with xenoliths of paraschists. (a) Relict intrusive contact between Permian granitoids and the country metamorphic basement (paragneiss), preserved in a block near Case Bastoneri. (b) Detail of the porphyritic structure of the metagranite. (c) Contact metamorphosed xenoliths of the country Variscan paraschists included in the late-Variscan porphyritic metagranite. Note that the xenoliths preserve a clear Variscan metamorphic foliation. (d) Detail of a xenolith of a banded metapelite.

The inferred evolution is similar to that recognized all over the relict lithologies of the Polymetamorphic Complex (Compagnoni & Hirajima, 1991; Compagnoni et al., 1995; Compagnoni & Rolfo, 2003).
Most xenoliths are *banded metapelites* (Fig. 21c d) which consist of layers rich in fine-grained randomly oriented phengite alternating with layers including:

- kyanite aggregates pseudomorphous after prismatic sillimanite (Fig. 22b);
- coronitic garnet with local "sagenitic" rutile inclusions after biotite;
- pre-Alpine porphyroblastic garnet surrounded by thin coronas of Alpine garnet;
- rutile aggregates rimmed with a garnet corona, most likely after primary ilmenite;
- jadeite-rich pyroxene (or its albite-bearing retrogressive products) with small zoisite crystals after pre-Alpine plagioclase;
- polygonal granoblastic quartz aggregates after coesite;

– aggregates of small garnets with minor phengite, most likely after pre-Alpine, contact metamorphic, cordierite (Fig. 22c,d).



In the samples more pervasively recrystallized during the Alpine peak metamorphic event, a grain-size coarsening of the UH*P* minerals occurs, which progressively obliterates the pre-Alpine textures: the rock is, therefore, converted to a phengite-garnet granofels with accessory rutile and local Na-pyroxene, quartz and zoisite.

Minor xenoliths are *metabasics*, entirely converted to quartz- and phengite- bearing eclogites with abundant accessory rutile. In the eclogite, the only recognized pre-Alpine relics are large titanite grains, partly replaced by a rutile ± omphacite intergrowth. In the garnet, the common occurrence of "sagenitic" rutile suggests its derivation from pre-Alpine biotite.

Fig. 22 – Stop 2: Representative microstructures of the metagranite and of the contact metamorphosed Variscan amphibolite-facies paraschist (see Tables 1, 2). (a) Metagranitoid collected near to the contact with the country metamorphic basement ("granitoid marginal facies"). Euhedral crystals of former igneous plagioclase (now a fine-grained aggregate of jadeite+zoisite±kyanite), cordierite (now a fine-grained aggregate of kyanite+garnet±biotite), and biotite (partly replaced by phengite) are set in a matrix of granoblastic polygonal quartz derived from inversion from coesite. Sample DM 1116. PPL. (b) Contact metamorphosed Variscan amphibolite-facies paraschist. Bundles of regional Variscan sillimanite, pseudomorphically replaced by a UH*P* kyanite aggregate, are embedded in a dark matrix derived from late-Variscan contact metamorphic cordierite, now mainly consisting of fine-grained UH*P* garnet aggregates. Sample DM493. PPL. (c) Relict Variscan paraschist showing kyanite pseudomorphs after contact metamorphic andalusite, and garnet+kyanite+white mica pseudomorphs after contact metamorphic cordierite. Sample DM1143. PPL. (d) Metapelite showing a corroded Variscan garnet partly replaced by late-Variscan contact metamorphic cordierite, now replaced by a fine-grained aggregate of Alpine garnet + kyanite. Sample DM495. PPL. Modified from Compagnoni et al. (2004).

Route: By minivan from Vallone di Gilba in Val Varaita to Martiniana Po in Valle Po and uphill to Case Ramello

Stop 2.3: Typical whiteschist with pyrope megablasts

Case Ramello, Martiniana (Valle Po)
(Coord: N 44° 36' 54'' E 7° 22' 00'', elev. 860 m a.s.l.)

This is the type locality where Chopin (1984) first described the occurrence of regional metamorphic coesite in continental crust rocks. The BIU whiteschist occurs in the orthogneiss of the Monometamorphic Complex as layers or boudins from a few centimetres to ca. 20 m in thickness and from a few metres to hundreds of metres in length. The contact between the whiteschist and the hosting orthogneiss is marked by a few decimetres to several metres of a phengite-rich, banded gneiss locally preserving relics of the igneous K-feldspar phenocrysts.

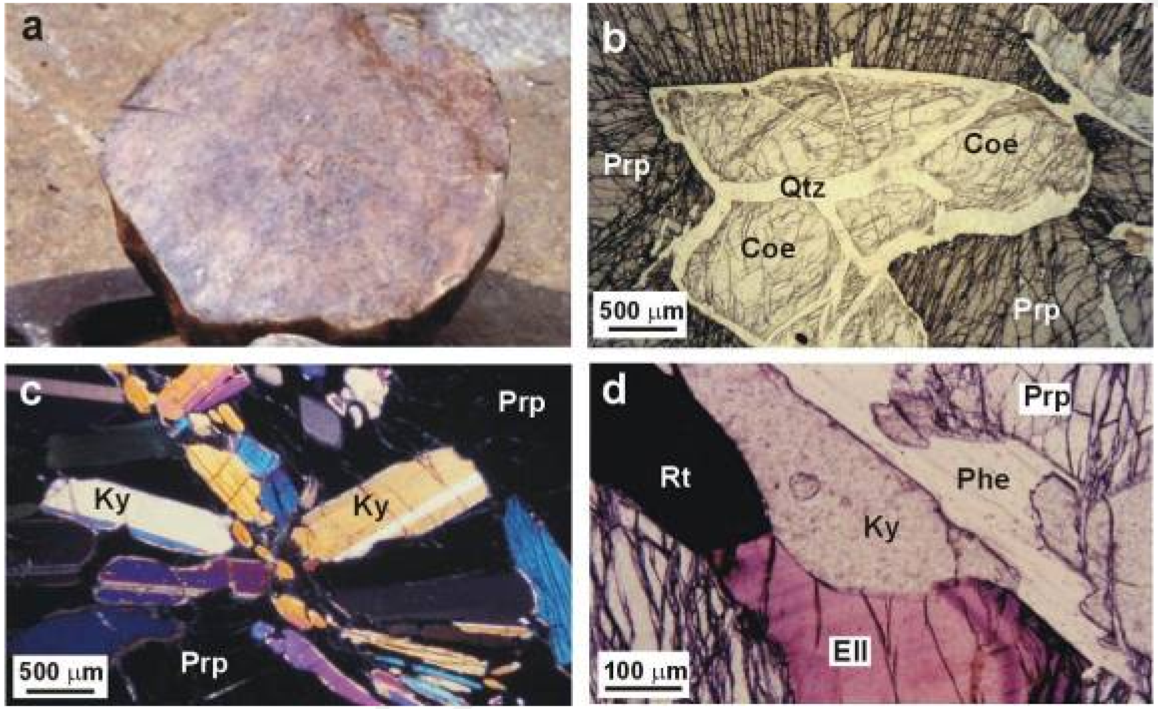


Fig. 23 – Stop 3: Whiteschists with pyrope megablasts. (a) Pyrope megablast, ca. 20 cm across. Note that pyrope is crowded of inclusions of kyanite, rutile, etc. (b-d) Representative microstructures of the pyrope-bearing whiteschists. (b) Relics of coesite included in a small pyrope porphyroblast and partially converted to quartz along rims and fractures. Radial fractures, consequent to volume increase during the coesite to quartz inversion, are developed in the hosting pyrope. Sample DM507. PPL. (c) Radial kyanite aggregate in pyrope. Sample DM450. XPL. (d) Zoned ellenbergerite in association with kyanite, phengite, and rutile armoured in a pyrope megablast. Sample DM450. PPL. Modified from Compagnoni et al. (2004).

Whiteschist consists of pyropes, usually exhibiting a trapezohedron habit, included in a foliated matrix of quartz inverted from coesite, high-celadonite phengite (polytype 3T, that is more stable than monoclinic polytype under *HP*-*UHP* conditions; Amisano Canesi et al., 1994; Ferraris et al., 2000, 2005), kyanite, talc and accessory rutile, zircon and monazite. On the basis of the modal amount of minerals, two kinds of whiteschist can be recognized: (1) Prp-rich whiteschist mainly consisting of pyrope megablasts (from 10 to 20 cm in diameter; Fig. 23a and Fig. 24) or porphyroblasts (from 2 to 10 cm in diameter; Fig. 24) with minor amount of kyanite, talc, phengite and quartz; (2) Qtz-rich whiteschist that consists of quartz, kyanite, phengite, pyrope neoblasts (<2 cm in diameter; Fig. 24) and talc (Ferrando et al., 2009).

Up to more than 50 per cent in volume of the pyrope blasts are commonly crowded with mineral inclusions (Table 3), some of which are not found in the matrix (Chopin, 1984; Schertl et al., 1991; Hermann, 2003).

Ubiquitous primary inclusions	
kyanite >> rutile > zircon, monazite, phengite, talc, clinochlore, paragonite, glaucophane phlogopite (usually retrogressed to K-deficient phlogopite or vermiculite), coesite	
Less common primary inclusions	***Probable retrograde minerals***
apatite	corundum
arsenopyrite	enstatite
dravitic tourmaline	gedrite
jadeite	sapphirine
magnesiochloritoid	
wagnerite $Mg_2PO_4(OH,F)$	
Data from Chopin (1984), Schreyer (1985), Chopin (1986), Chopin & Klaska (1986), Chopin et al. (1986, 1991, 1994), Schertl et al. (1991), Compagnoni et al. (1995), Simon et al (1997), Simon & Chopin (2001).	

New minerals	
Bearthite Chopin et al. (1993)	$Ca_2Al[PO_4]_2(OH)$
Ellenbergerite Chopin et al. (1986)	$Mg_6(Mg,Ti)_2Al_6[(OH)_{10}\|Si_8O_{28}]$
Phosphoellenbergerite Brunet et al. (1998)	$H_2(Mg,Fe,\square)_{14}[(OH)_6\|CO_3\|((P,As,S)O_4)_7]$
Magnesiodumortierite Chopin et al. (1995), Ferraris et al. (1995)	$(Mg,Ti)(Al,Mg)_2Al_4[BO_3\|(OH)_3\|O\|(SiO_4)_3]$
Magnesiostaurolite Chopin et al. (2003)	$(Mg,Li)_3Al_{18}[(OH)_4\|O_{12}\|(SiO_4)_8]$

Table 3 – Review of mineral inclusions in pyrope blasts from BIU whiteschists.

Near Case Ramello, four generations of pyrope are recognisable by microstructural relationships, chemical composition and mineral inclusions (Ferrando et al., 200). Prp 0 ($Prp_{25-37}Alm_{58-70}$) occurs as rare irregular and corroded relics at the core of porphyroblasts (Fig. 24) and includes staurolite, chloritoid, kyanite, chlorite and paragonite (Compagnoni & Hirajima, 2001). Prograde Prp I ($Prp_{69-81}Alm_{16-26}$) constitutes the core of megablasts and porphyroblasts (Fig. 24) and includes kyanite (Fig. 19c) (locally oriented), talc, phlogopite (usually retrogressed to vermiculite), ellenbergerite (Fig. 23d) (Chopin, 1986; Chopin et al., 1986; Chopin & Langer, 1988), Mg-chlorite, Mg-dumortierite, rutile, zircon, apatite and monazite, but never phengite. Coesite/quartz in Prp I are rarely observed, and only within porphyroblasts.

Peak Prp II ($Prp_{82-98}Alm_{1-16}$) constitutes the inner rim of megablasts and porphyroblasts and the core of neoblasts (Fig. 24), and it contains inclusions of coesite (Fig. 23b), kyanite, phengite, rutile, zircon, monazite and rare talc, but never Mg-chlorite and phlogopite. Early retrograde Prp III ($Prp_{86-67}Alm_{4-11}$) constitutes the outer rim (Fig. 24) and usually it is pseudomorphically replaced by phengite, biotite, talc, kyanite and chlorite. It includes phengite, kyanite, rutile and zircon.

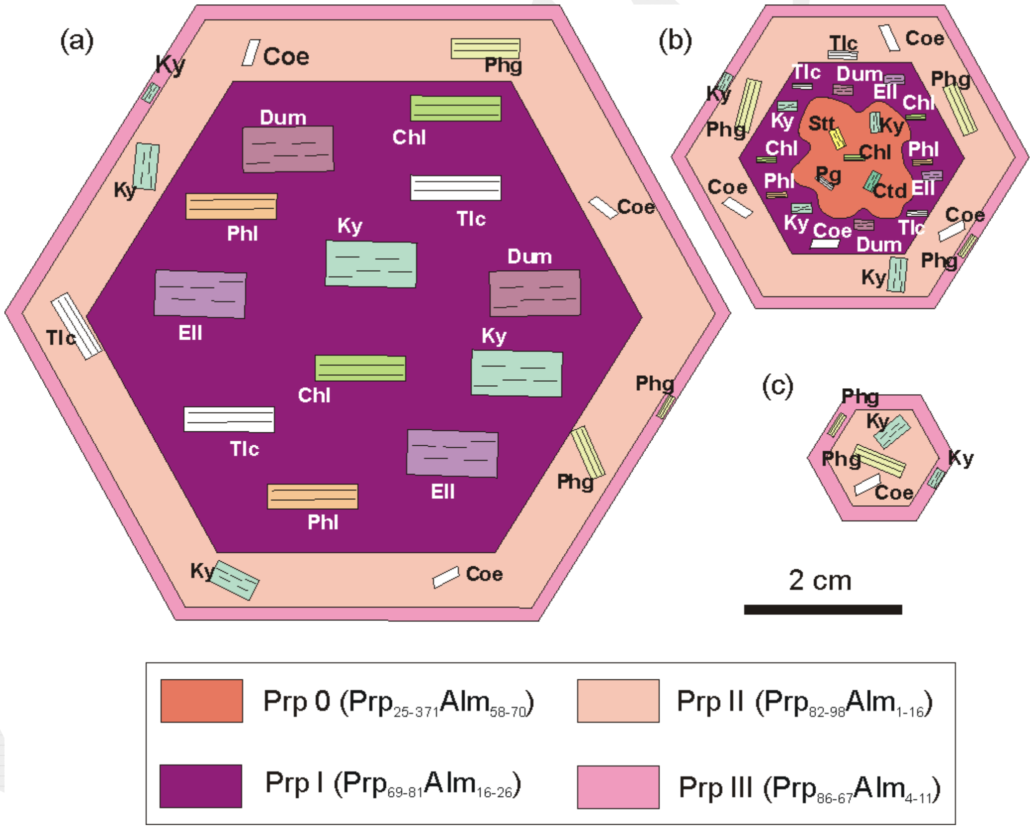


Fig. 24 - Simplified sketch (modified from Ferrando et al., 2009) showing the four generations of pyrope recognized in the UHP whiteschists near Case Ramello, their chemical zoning and mineral inclusions: (a) megablast, (b) porphyroblast, (c) neoblast. Mineral abbreviations after Fettes & Desmons (2007).

Similar pyrope-bearing whiteschist and/or quartzite have been reported from several other localities within the BIU (e.g. in the Gilba Valley). In addition to the prograde inclusions described for the Case Ramello locality, an almost pure magnesiochloritoid (up to 97 mole % end-member) was reported as prograde inclusion within pyrope by Simon et al. (1997). Rare corundum, enstatite, sapphirine, and gedrite, have been also described by Simon & Chopin (2001) to fill pyrope cracks most likely developed during early decompression.

Origin of the whiteschist: metasomatism from ultramafics

The chemical composition of whiteschist is unusual because characterized by an extremely high Mg/Fe ratio, high SiO_2 and the virtual absence of Ca and Na.

Chopin (1984) considered the BIU whiteschists similar to "the highly magnesian, metaevaporitic whiteschists" described and reviewed by Schreyer (1977) and, then, its origin should have been sedimentary and the probable protolith should have been "a mixture of quartz, Mg-chlorite and illite". Nevertheless, a metasomatic origin by fluid infiltration along ductile shear zones within the (meta)granitoid has been demostrated by many authors (e.g., Sharp et al., 1993; Gebauer et al., 1997; Compagnoni & Hirajima, 2001; Schertl & Schreyer, 2008; Ferrando et al., 2009; Grevel et al., 2009).

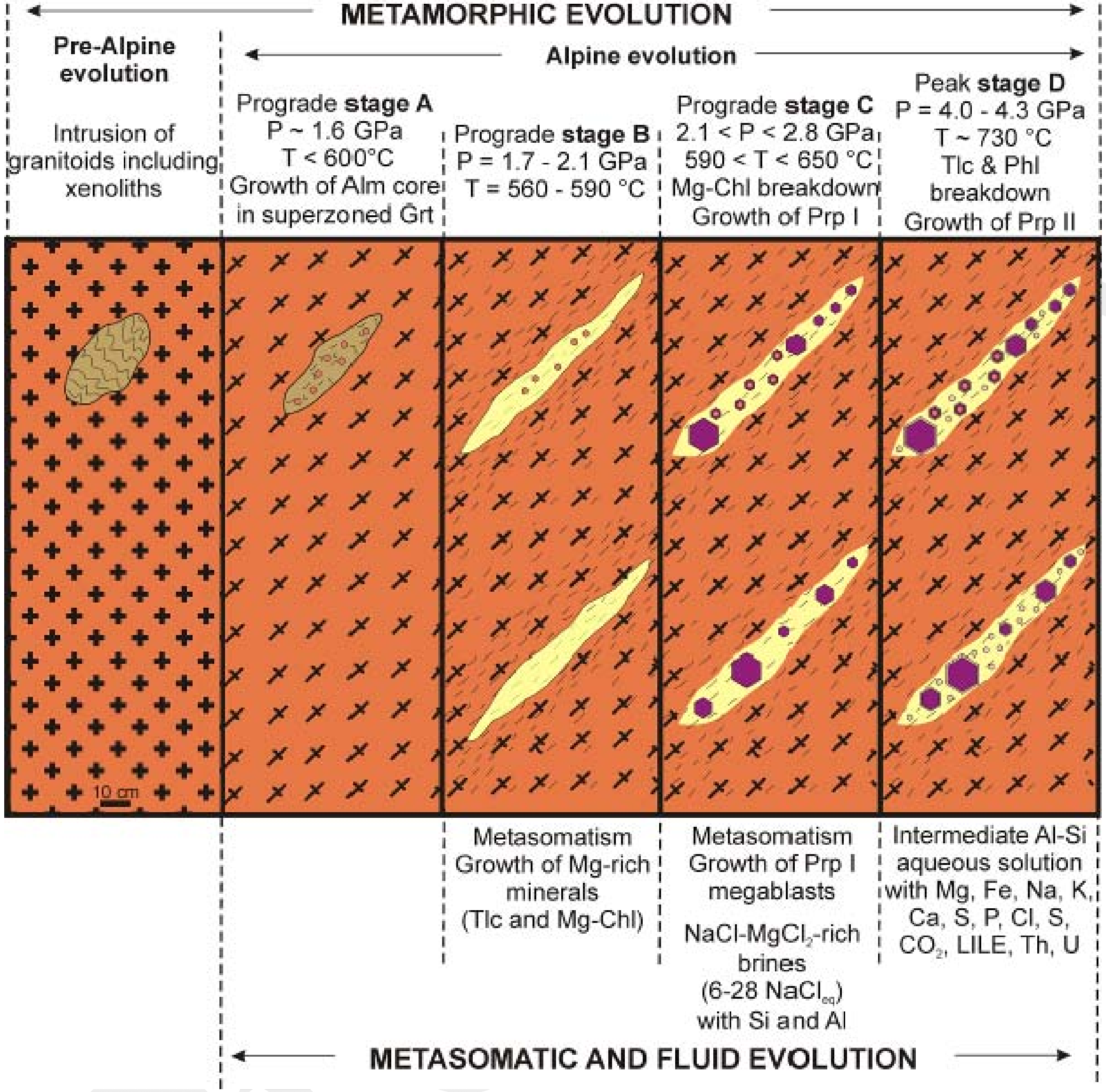


Fig. 25 - Origin and evolution of the BIU whiteschists up to UH*P* peak (modified from Ferrando et al., 2009). The different generations of garnet are shown (cf. Fig. 24). An original post-Variscan, most likely Permian, granitoid, including a paraschist xenolith of the country Variscan metamorphic basement, undergoes an Alpine prograde H*P* to UH*P* metamorphism. At stage A, an almandine-rich garnet grows in the xenolith. At stages B and C, the infiltration along shear zones of high-salinity, high-Mg fluids, generated from dehydration of antigorite in subducting serpentinites, metasomatised the orthogneiss and the xenoliths to generate the whiteschist. At the same time, metamorphic dehydration reactions – involving talc, muscovite and Mg-chlorite – also occurred within the whiteschists. At stage D, the peak mineral assemblage that formed in the presence of an alumino-silicate aqueous solution enriched in LILE, Th and U, likely originated by dehydration reactions involving talc and phlogopite.

Moreover, the chemical composition of the BIU whiteschist and its field relationships with the host orthogneiss are similar to those of other Mg-metasomatic rocks occurring in the Alps (see Ferrando, 2012 for a review). For example, the "silvery micaschists" occurring in the orthogneiss of the Gran Paradiso massifs (Chopin, 1981; Chopin & Moniè, 1984) are considered by Compagnoni & Lombardo (1974) as the H*P* metasomatic product of sheared granites because of their close field association and systematic transition to the surrounding orthogneiss.
From protolith, the debate moved to timing of metasomatism and origin of the metasomatic fluid. Sharp et al. (1993) and Sharp & Barnes (2004) interpreted the whiteschists as metasomatic rocks that originated by the influx of fluids from the subducting oceanic lithosphere. Gebauer et al. (1997) proposed a low *P*–*T* metasomatism occurred during the Permo-Triassic rifting of the Piemonte-Liguria Basin. The discovery in the BIU whiteschists of rare superzoned garnets with almandine-rich core—interpreted as relic of a former pelitic xenolith—and pyrope-rich rim allows Compagnoni & Hirajima (2001) to propose that the metasomatic process occurred in the presence of a hydrous fluid phase during prograde Alpine metamorphism, close to the UH*P* climax.
In a recent petrological, geochemical and fluid inclusion study, Ferrando et al. (2009) constrained a model for the metasomatic formation of the whiteschists and their subsequent UH*P* history (Fig. 25). An original post-Variscan granitoid, including xenoliths of the country Variscan metamorphic basement, undergoes an Alpine prograde metamorphism (P=1.6 GPa and T<600 °C) with the growth of almandine-rich garnet (Prp 0) in the xenoliths. A first prograde metasomatic event occurred at 1.7–2.1 GPa and 560–590°C, due to the influx of external fluids originated from the antigorite breakdown in subducting oceanic serpentinites: this event promoted the increase in Mg and the decrease of alkalies and Ca in the orthogneiss toward a whiteschist composition. During the following prograde evolution (2.1<P<2.8 GPa; 590<T<650°C), the metasomatic fluid influx coupled with internal dehydration reactions involving the Mg-chlorite breakdown promoted the coarse-grained growth of a first generation of pyrope (Prp I) in the presence of a $MgCl_2$-brine. Near the UH*P* metamorphic peak (4.0–4.3 GPa and 730°C), a fine-grained second pyrope generation (Prp II) grew in the presence of a alumino-silicate aqueous solution, mostly generated by internal dehydration reactions involving phlogopite and talc. The contribution of metasomatic external brines at the metamorphic climax appears negligible. In an early decompression stage (3.6–3.9 GPa and 720 °C), a third generation of pyrope (Prp III) grew in a close system.

HP-UHP evolution of the fluid phase in the whiteschist: chemical transfer in continental subduction zones

Few studies on DM whiteschists were devoted to characterize the fluid phase at HP-UHP conditions. Composition of HP fluid has been obtained from primary fluid inclusions within prograde kyanite in Prp I (Philippot et al., 1995; Ferrando et al., 2009). This fluid corresponds to a NaCl-$MgCl_2$-rich brine whose salinity varies from 3 to 28 wt% of $NaCl_{eq}$. The presence of Si and Al as additional dissolved cations, testifyied by paragonite as daughter mineral in the fluid inclusions (Fig. 26a), reduces below 0.8 the a_{H2O} in the fluid (Ferrando et al., 2009). Experiments of Hermann (2003) and Hermann et al. (2006) confirm the capability of this fluid to dissolve moderate amounts of solids. Stable isotope data (Sharp et al., 1993; Sharp & Barnes, 2004) indicate an esternal origin for this fluid that, then, is considered to be the metasomatic fluid originated by antigorite dehydration (Sharp & Barnes, 2004; Ferrando et al., 2009).

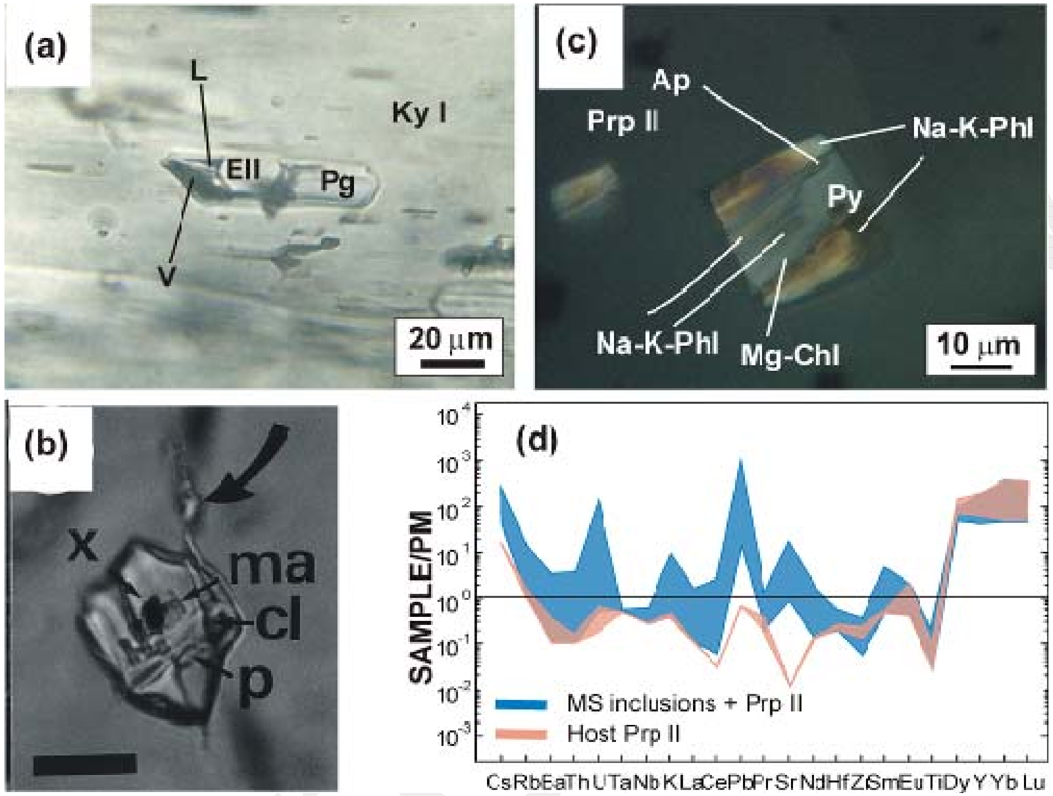


Fig. 26 – Fluid evolution in whiteschists from H*P* to UH*P* conditions (modified from Philippot et al., 1995 and Ferrando et al., 2009). (a-c) Photomicrographs of fluid and multiphase solid inclusions. (a) Primary three-phase (S + L + V) inclusion in prograde Ky (PPL). (b) Decrepitated (see arrow) fluid-free inclusion in pyrope megablast (PPL). ma = magnesite, p = Mg-phosphate, cl = a chloride and x = undetermined opaque mineral. (c) Multiphase solid (MS) inclusion in Prp II containing the typical mineral assemblage Mg-chlorite, Na-K phlogopite, Zn-rich pyrite, and Cl-rich apatite (XPL). (d) Trace-element concentrations in MS inclusions + Prp II normalized to the primordial mantle (McDonough & Sun, 1995), compared with those in Prp II. The fluid within MS inclusion is enriched in LILE (Pb, Cs, Sr, Rb, K, LREE, Ba), Th, and U.

Philippot et al. (1995) report the presence in pyrope megablasts of complex inclusions consisting of magnesite, Mg-phosphate, chloride, talc or Mg-chlorite, and an opaque mineral – representative of a brine composed of H_2O, CO_2, Cl, P, Mg, Na or K, Si and Al (Fig. 26b). In peak Prp II, Ferrando et al. (2009) describe primary multiphase solid inclusions consisting of Mg-chlorite, Na-K phlogopite, Cl-rich apatite, Zn-rich pyrite (Fig. 26c). These inclusions represent remnants of an alumino-silicate aqueous solution, containing Mg, Fe, alkalies, Ca and subordinate P, Cl, S, CO_3, LILE (Pb, Cs, Sr, Rb, K, LREE, Ba), U and Th (Fig. 26d). The composition of this fluid is characterized by lower H_2O and Cl contents, higher Si and Al contents, and by the presence of Ca with respect the fluid reported by Philippot et al. (1995). These differences probably indicate a high variability in the fluid chemistry at UHP conditions that also condition the a_{H2O} that can reach very low values (ca. 0.4). An internal origin for this fluid has been demonstrated by petrological and geochemical data (e.g. Sharp et al., 1993; Philippot et al., 1995; Chopin & Schertl, 1999; Ferrando et al., 2009).
So, fluid inclusion and stable isotope studies delineate a progressive and continuous transition from external-derived prograde Cl-dominated aqueous fluids to internal-derived peak alumino-silicate aqueous solution. At increasing pressure, the aqueous fluid becomes able to dissolve high amounts of cations (particularly Si and Al) involved in metamorphic reactions, while the Cl content progressively decreases. These fluids, showing LILE enrichment and HFSE depletion (Fig. 26d), could represent important metasomatic agents for the release of incompatible trace elements into the supra-subduction mantle wedge.

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