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Training network on reactive geological systems from the mantle to the abyssal sub-seafloor

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Field Excursion Guidebook

2nd day (10/04/2016)

Western Liguria: pre-Alpine and Alpine, subduction history of the Erro-Tobbio peridotite Unit (Voltri Massif).

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1. *Introduction and aims*

The field trip is focussed on the main structural and petrologic features related to the mantle evolution during exhumation (from spinel-facies to shallow plag-facies lithospheric environments), and to the subsequent subduction-dehydration of serpentinized oceanic mantle that was originally part of the Mesozoic Ligurian Tethyan Ocean. The general geology and the exhumation processes of the high-pressure ophiolites of western Liguria will also be discussed.

The original oceanic lithosphere consisted of a peridotite basement locally intruded by gabbroic bodies and discontinuously covered by basalts and radiolarian cherts. Present knowledge indicates that the Ligurian Tethys formed by passive extension of the Adria-Europe lithosphere leading to sea-floor exposure of the subcontinental mantle. The ocean was later subducted during Alpine convergence between Europe and Adria plates.

The excursion to the Voltri Massif is focussed on the petrologic and tectonic aspects related to the pre-alpine exhumation to the sea-floor, and to subduction of ultramafic and mafic rocks. The ultramafic rocks diffusely preserve remnants of the pre-Alpine protoliths and evolution (STOP 1 and 2), related to lithosphere extension and low-grade alteration near the sea-floor. Emphasis will be given to subduction and eclogite-facies recrystallization of serpentinized peridotites, to the structures of fluid production and to the released fluids (STOP 3, 4).

The field trip will visit a classic locality for extension-related mantle evolution and serpentinite subduction: the Gorzente River, within the Erro-Tobbio peridotites of the Voltri Unit (Western Ligurian Alps). This locality displays the evolution from spinel facies peridotites to plagioclase-facies impregnation and tectonite to mylonite deformation (STOP 1, 2). The serpentinized peridotites also display the subduction structures and the veins formed during high-pressure dehydration (STOP 3, 4). On a larger scale, the serpentinized peridotites form a major body associated to '*country*' serpentinites that enclose metre to hundred metre-scale bodies of mafic rocks with eclogite- to blueschist-facies overprint. One of these minor bodies will be visited and the main petrologic and tectonic features related to subduction and burial of the whole area will be discussed (STOP 4).

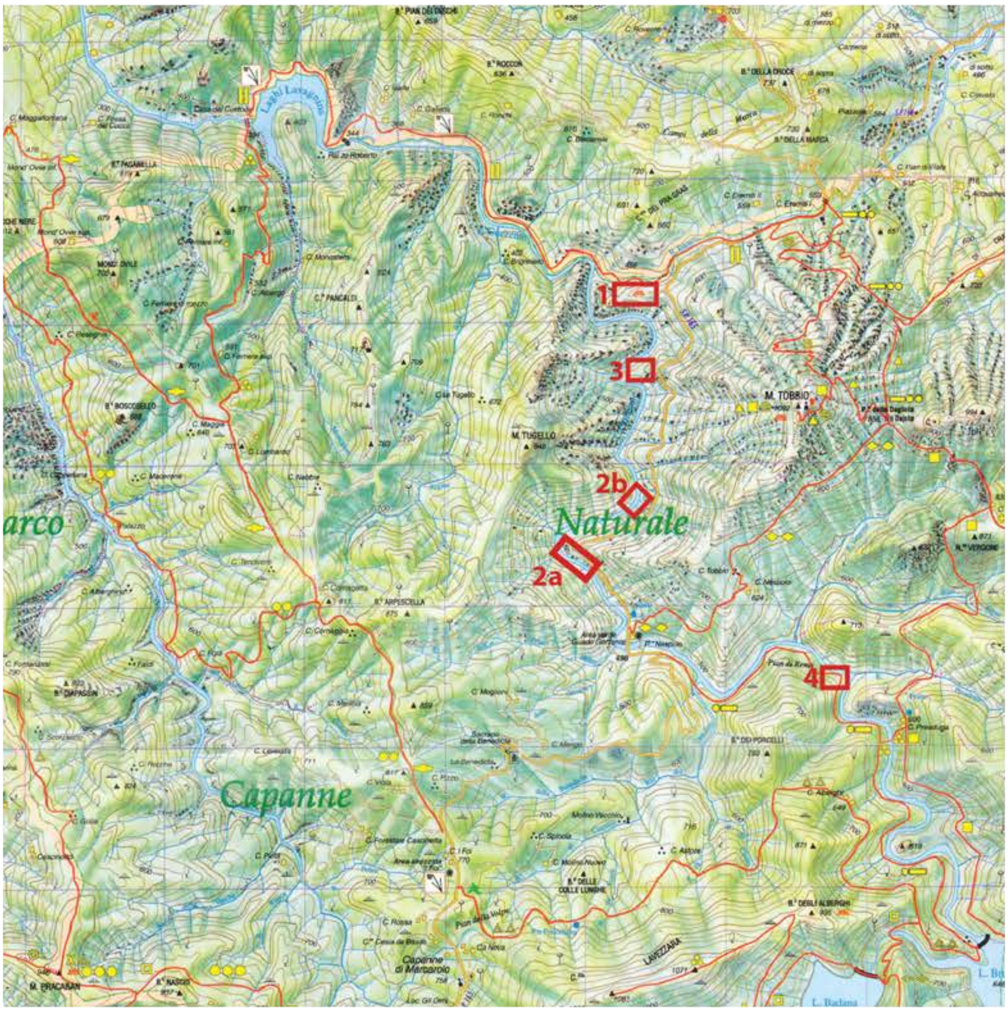


Figure 1: Map of the excursion, showing the location of the different STOPS.

2. *Regional setting*

Ophiolites exposed along the Western Alpine-Northern Apennine (and Corsica) orogenic system formed the oceanic lithosphere of the Ligurian Tethys that during Late Jurassic to Cretaceous divided the Europe and Adria continental plates.

Ophiolites show structural-petrographic features indicating that:

- Mantle rocks underwent a composite subsolidus evolution after depletion by partial melting and accretion to the subcontinental lithosphere;

- Episodes of reactive melt flow and melt impregnation affected the mantle rocks during their pre-oceanic evolution;

- Gabbroic rocks intruded the mantle peridotites;

- Peridotites and gabbros were exposed at the sea-floor prior to basaltic volcanism and deposition of radiolarian cherts.

Sheeted dyke complexes are lacking and co-magmatic relations do not exist between the gabbro bodies and the basaltic dykes and flows. Accordingly, a general consensus exists on the idea that the Jurassic Ligurian Tethys was floored by a peridotite-gabbro basement (Decandia and Elter, 1969; Piccardo, 1983; Lemoine *et al.*, 1987), subsequently covered by a discontinuous layer of younger pillow basalts and radiolarian cherts, i.e. the first oceanic sediments. The cherts are coeval to the basaltic lava flows and are younger than 160-150 Ma (De Wever and Caby, 1981; Marcucci and Passerini, 1991). Accordingly, the inception of the oceanic stage, following continental breakup, is not older than Late Jurassic.

3. *Palaeogeography of the Ligurian ophiolites*

The Ligurian Tethys developed in connection with pre-Jurassic rifting and Late Jurassic opening of the Northern Atlantic (Lemoine *et al.*, 1987; Dewey *et al.*, 1973; Figure 2). Reconstructions suggest that the basin was limited in size and did not reach the width of present-day oceans. Structural reconstructions of the oceanic lithosphere suggest formation environments comparable to present-day slow and ultra-slow spreading centers (Rampone and Piccardo, 2000). Age data indicate a narrow time span between end of divergence and onset of convergence. Oceanic accretion in the Ligurian Tethys began during Late Jurassic and continued for approximately 25 Ma (cf. Winterer and Bosellini, 1981). Plate convergence leading to subduction of the oceanic lithosphere was Early Cretaceous, about 25 Ma after cessation of the oceanic spreading (Hunziker, 1974). The subduction zone was south-west trending, with Europe underthrusting the Adria plate: subduction likely was intra-continental in the northern sector of the Western Alps and intra-oceanic in the Ligurian domain.

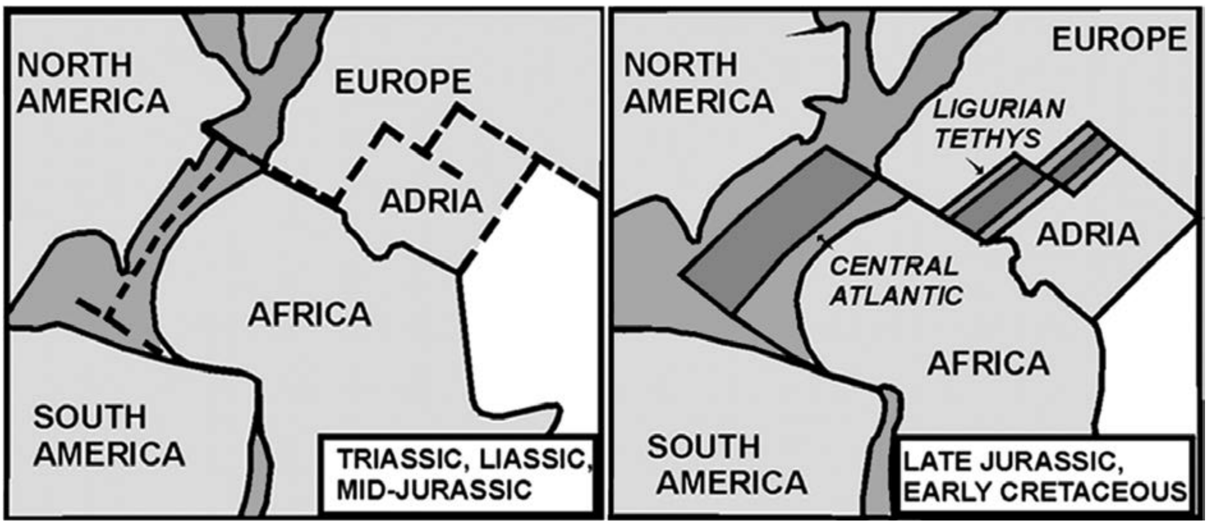


Figure 2 - Mesozoic evolution of Central Atlantic and Ligurian Tethys oceans, from rifting to ocean formation (redrawn and modified after Lemoine et al.*, 1987).*

The Ligurian Tethys closed during Early Tertiary, when fragments of the oceanic lithosphere emplaced as west-vergent thrust units in the Alps and east-vergent thrust units in the Apennine.

Depending on stratigraphy, structural and metamorphic characteristics, the various ophiolitic sequences of Liguria have been ascribed to different palaeogeographic domains within the Ligurian Tethys. The Erro-Tobbio peridotite unit (ET) has been inferred to be located close to the Adria continental margin (Scambelluri et al., 2012). The Northern Apennine ophiolites, which record low-grade orogenic metamorphism, were located east of the subduction zone, closer to the Adria margin (Figure 3) and in a setting comparable to ocean-continent transition zones and very slow spreading ridges.

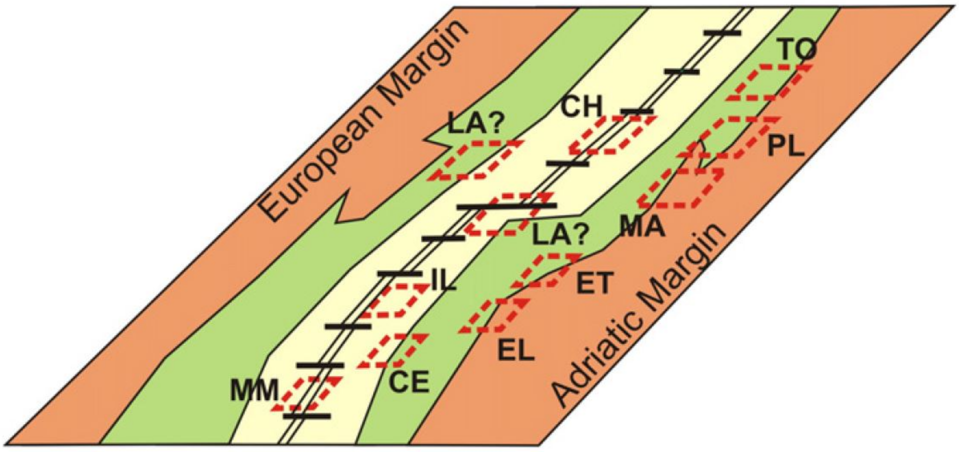


Figure 3 - Generalized paleogeographic restoration of Upper Jurassic Ligurian Tethys, with location of the different Alpine domains (Rampone & Hofmann, 2012).

4. The Voltri Massif

The Voltri Massif (Figure 4) crops at the southeastern end of the Western Alps and is part of the Internal Penninic Units of the Alps. Historically, several tectonic slices have been recognized in the Voltri Massif (e.g. Chiesa et al., 1975): these have been recently grouped in two major Units, Voltri and Palmaro-Caffarella (Capponi and Crispini, 2008), that underwent subduction metamorphism under eclogite- and blueschist-facies conditions, respectively. Both Units are affected by greenschist-facies overprint. The Voltri Massif is composed of meta-ophiolitic rocks, oceanic metasediments, minor platform metasediments of the European margin, slices of the subcontinental lithospheric mantle and a minor slice of continental crystalline rocks of European affinity (Valosio and Arenzano massifs; Chiesa et al., 1975; Messiga et al., 1992). Meta-ophiolites derive from the Ligurian oceanic lithosphere and correspond to serpentinites with metagabbros associated with metasediments (calcschists, minor mica- and quartz-schists) and metabasites. Mantle rocks (spinel to plagioclase lherzolites and harzburgites with minor pyroxenite and dunite) also occur and are widespread in the Erro-Tobbio peridotite body.

To the east The Voltri Unit is separated from the Ligurian Apennines by several tectonic slices historically grouped into the Sestri-Voltaggio Zone (Figure 4), consisting of Triassic-Liassic limestones and dolostones associated with dismembered Mesozoic ophiolites. These rocks record lower-grade metamorphic conditions than Voltri and Palmaro-Caffarella Units, and range from low-T blueschist to pumpellyite-actinolite facies. Klippen of analogous blueschist-facies rocks overlie the eclogitic Voltri Massif to the W-SW. To the southwest the Voltri Massif overlies the Hercynian gneiss (granitoids with associated amphibolites) of the Savona continental basement, showing greenschist- to blueschist-facies Alpine metamorphism.

To the north the Voltri Massif is bounded by Tertiary sediments of the Piemontese Basin. The sedimentary succession unconformably overlies the high-pressure basement and begins with Late Eocene-Early Oligocene continental breccias (Costa Cravara Breccia), followed by the Lower Oligocene transitional to marine conglomerates of the Molare and Savignone formations (Gelati et al*.,* 1993). Breccias and conglomerates include clasts of high-pressure rocks, thus giving an upper limit of 35-38 Ma for the exhumation and erosion of the high-pressure Voltri Massif. Geochronological investigations of the high-pressure rocks yield ages of the eclogite-facies metamorphism ranging between 49 Ma (Federico et al., 2005; Figure 5) and about 33 Ma (Rubatto and Scambelluri, 2003).

Several studies (Chiesa et al., 1975; Messiga and Piccardo, 1974; Piccardo, 1977) have shown that in the Voltri Massif slices from different lithospheric levels (subcontinental mantle, Mesozoic oceanic lithosphere and sediments, continental margin basement and sediments) are tectonically coupled. Although much of the lithologic and tectonic associations in Voltri were achieved during subduction and early exhumation, the present-day geometry is largely dominated by collisional structures developed under greenschist-facies, represented by (i) late-stage (post-nappe) folding of high-pressure terrains and (ii) shallow thrusts that masked the subduction nappe setting.

The field trip is performed in the eastern sector of the Voltri Unit (Figure 4). In spite of the Alpine overprint, the ET unit preserves kilometer-scale volumes of unaltered peridotites that retain mantle textures and mineral assemblages, thus allowing the study of their pre-Alpine mantle evolution. It consists of partly serpentinized mantle peridotites (Erro-Tobbio) variably sheared along a steeply dipping regional schistosity trending North-South, serpentinites enclosing bodies of metagabbros and metasediments. The bodies are variable in size and record different peak conditions (from eclogite- to blueschists-facies) and different P-T paths (Malatesta et al., 2012a). Moreover, disrupted parts of the Palmaro-Caffarella Unit are sheared and boudinaged inside the Voltri Unit following the regional foliation. This tectonic setting can be interpreted assembled in a subduction domain at the interface between the converging plates (Malatesta et al., 2012b).

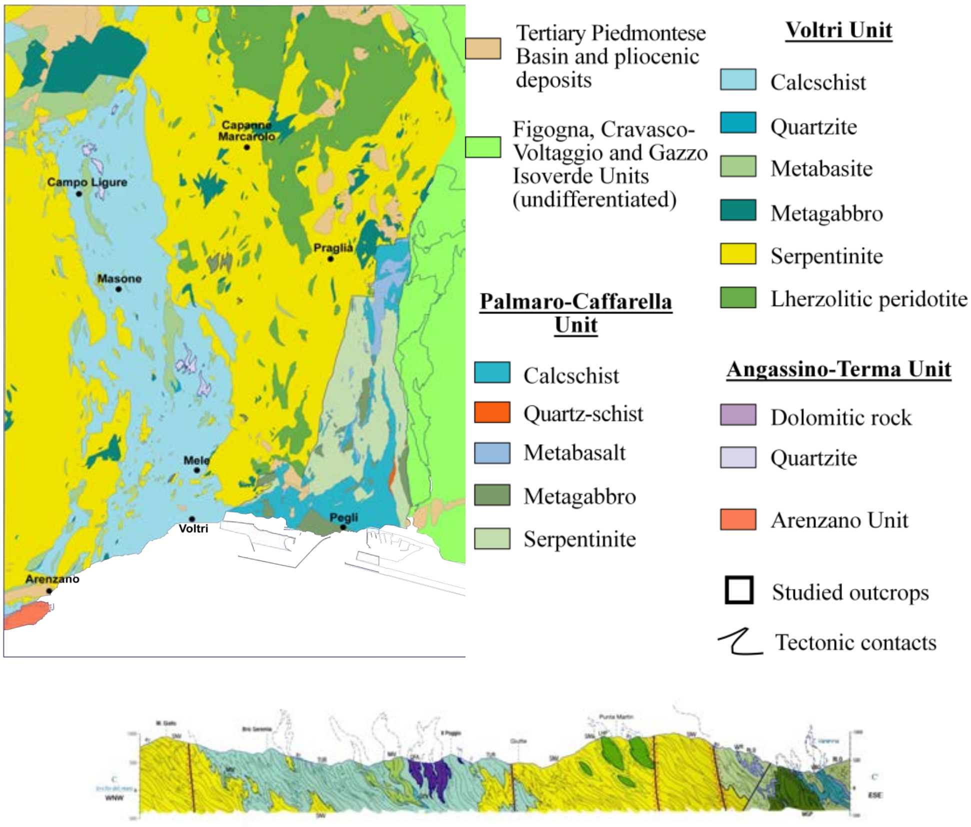


Figure 4 - Geological sketch map of the Voltri Massif showing location of the excursion Stops (after Capponi and Crispini, 2008)

5. The Voltri eclogites

Eclogites of the Voltri Massif have been described since long (Mottana and Bocchio, 1975; Ernst et al., 1982; Piccardo et al., 1979; Piccardo, 1984; Messiga and Scambelluri, 1991). These rocks will not be visited during the excursion. They outcrop in the central and western sector of the Massif (westernmost part of Figure 4 and outside the map) and mostly derive from precursor gabbros, mainly Fe-Ti-rich varieties and subordinated Mg-Al-rich gabbros. They are enclosed in high-pressure (olivine-bearing) serpentinites; intrusive contacts with the host ultramafites are locally preserved (Cimmino et al., 1979). Mafic intrusives correspond to crystal cumulates at different fractionation steps from a common primary tholeiitic N-MORB magma (Ernst et al., 1982; Piccardo, 1984). The most widespread eclogites display the classic garnet + Na-pyroxene + rutile assemblage (Ernst, 1976; Messiga, 1987; Messiga and Scambelluri, 1988; 1991; Messiga et al., 1995a) and derive from original Fe-Ti-oxide-bearing gabbros (the “superferrian eclogites” of Mottana and Bocchio, 1975). P-T conditions ranging between 13-18 kbar and 450-500 °C have been achieved for these rocks by different authors (Messiga and Scambelluri, 1991; Liou et al., 1998). The pre-subduction oceanic history of these rocks was locally accompanied by metasomatic exchange with the host ultramafic rocks which led to formation of metarodingites (garnet + diopside + zoisite/clinozoisite + chlorite + magnetite ± apatite) and of rocks containing Ti-clinohumite + diopside + magnetite + chlorite + apatite. The above assemblages are indicative of pre-subduction Ca- and Mg-metasomatism, respectively (Piccardo et al., 1980; Cimmino et al., 1981; Scambelluri and Rampone, 1999).

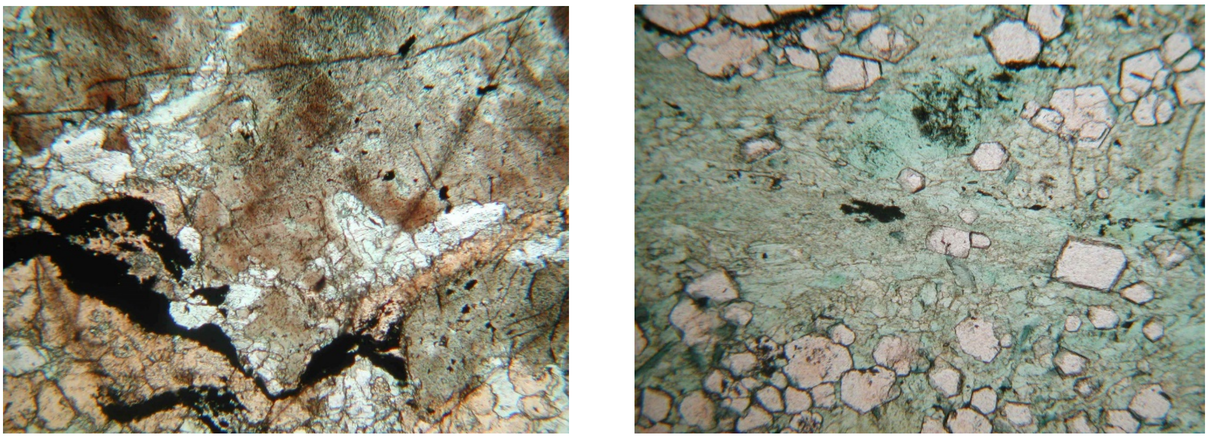


Figure 5 - Right: Voltri eclogite. Left Voltri meta-rodingite with pink garnet, white diopside after igneous brown augite. Also present is interstitial ilmenite. The rocks derive from precursor Ft-Ti-gabbros.

6. The Erro-Tobbio peridotites and serpentinites

6.1. The mantle history

Although the Erro-Tobbio peridotites have been involved in the Alpine orogenesis, the localization of alpine deformation along serpentinite mylonite shear zones has enabled preservation of km-scale volumes of less serpentinized peridotites that retain mantle textures and assemblages (Hoogerduijn Strating, 1991; Figure 6). This has allowed structural and petrological investigations on their upper mantle evolution that predated their shallow emplacement in the Ligurian Tethys (Bezzi and Piccardo, 1971; Ernst and Piccardo, 1979; Ottonello et al., 1979; Piccardo et al., 1990, 1992; Hoogerduijn Strating et al., 1990, 1993; Vissers et al., 1991; Scambelluri et al., 1995). Comprehensive recent reviews of the mantle history of these rocks are provided by Piccardo and Vissers (2007), Rampone and Borghini (2008).

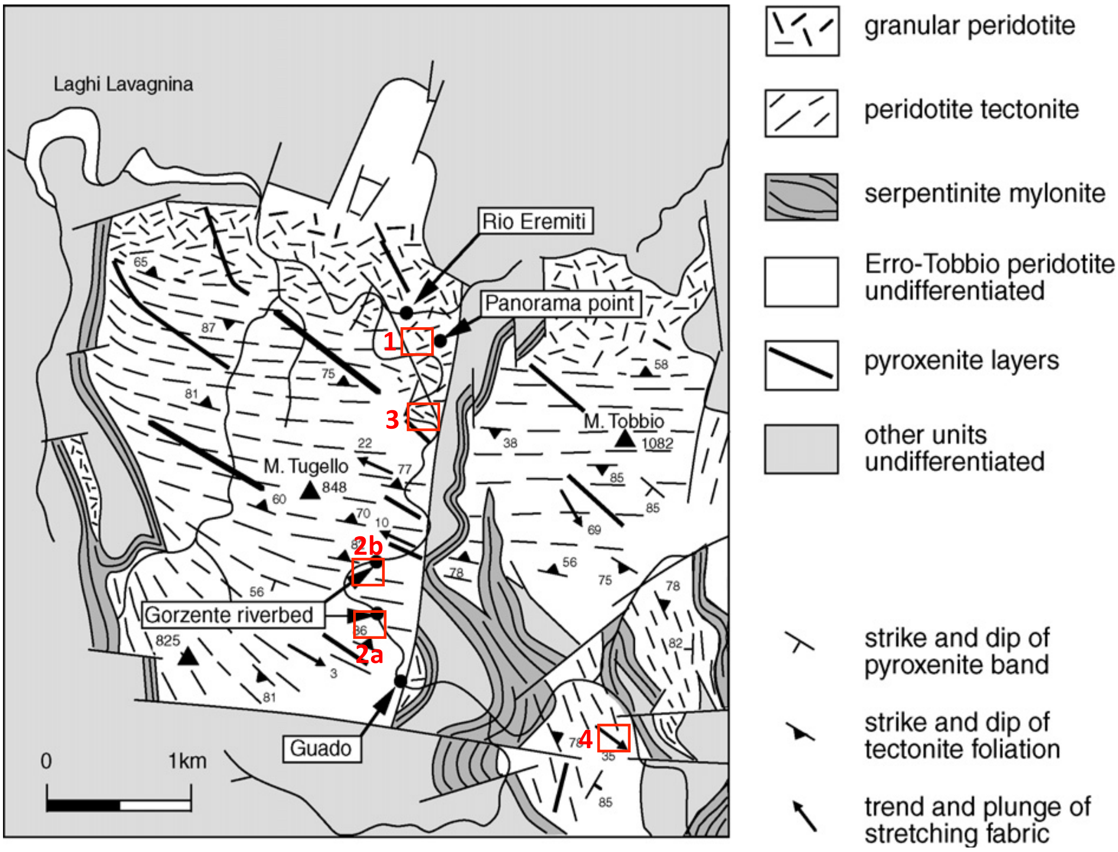


Figure 6 - Structural sketch map of the Erro-Tobbio peridotites (after Piccardo and Vissers, 2007) showing the different excursion localities.

The Erro-Tobbio mantle peridotites consist of partly serpentinized cpx-poor lherzolites and harzburgites, which commonly show spinel-bearing assemblage. They vary from granular types to peridotite mylonites. The structural and petrologic analysis of the peridotite domains preserving the pre-subduction history has shown that the oldest textures and assemblages are retained by granular spinel-facies peridotites. These rocks record the deep early intrusion of mafic melts that crystallized spinel-pyroxenites (Rampone and Borghini 2008). The spinel peridotites became overprinted by spinel-, to plagioclase-, to hornblende-bearing peridotite tectonites and mylonites forming composite km-scale shear zones (Vissers et al., 1991). These phase transition point to exhumation of the Erro-Tobbio peridotites to shallow levels in the extending lithosphere, as shown in Figure 7. Geothermometry (Hoogerduijn Strating et al., 1993) indicates progressive Temperature decrease from spinel-facies (1000-1100°C; about 20 Kbar) to plagioclase-facies (900-1000°C), to hornblende-facies (< 900°C).

It has been recently shown that the pre-alpine tectonic mantle evolution of the Erro-Tobbio peridotites was accompanied by multiple episodes of melt/peridotite interaction (occurred at progressively shallower depths), which originated a variety of rocks (reactive harzburgites, dunites, plagioclase-rich peridotites) and rock textures (Piccardo and Vissers (2007; Rampone and Borghini., 2008).

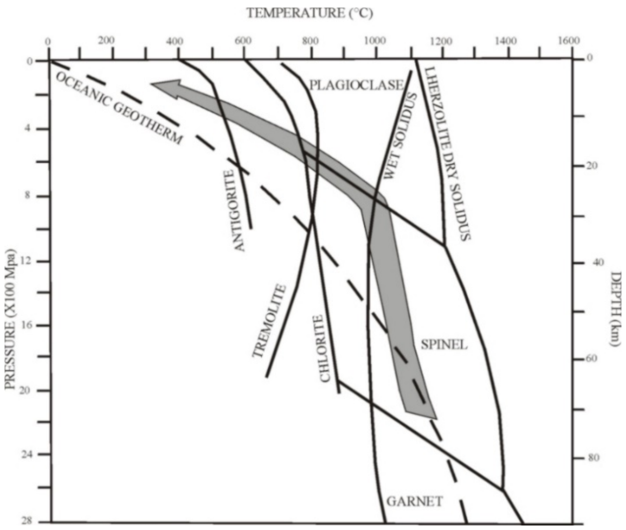


Figure 7 - Pressure – Temperature path showing the mantle evolution of the Erro-Tobbio peridotite (after Hoogerduijn Strating et al., 1993).

The **spinel granular peridotites** have protogranular to porphyroclastic textures and locally preserve opx + cpx + spinel rounded clusters, possibly after precursor high-P garnet. The spinel peridotites display olivine coronas showing lobate contacts and partly replacing exolved mantle orthopyroxene and clinopyroxene, indicating melt/rock reactions causing pyroxene dissolution and olivine crystallization. Widespread development of such reactions caused pyroxene dissolution and significant increase in modal olivine (up to volume 86%) and produced “reactive harzburgites” (Rampone et al., 2004, 2005a). The granular peridotites and the pyroxenites appear overprinted by subsequent spinel tectonites and mylonites distributed in kilometre-scale extensional shear zones developed during progressive exhumation (Hoogerduijng Strating et al., 1992; Piccardo & Vissers, 2007).

Bulk-rock and mineral compositions of the Erro-Tobbio spinel peridotites point to an overall depleted signature. The most important features are: i) the depletion in fusible components (i.e. low Ca, Al, Ti, LREE contents in both bulk rocks and constituent clinopyroxene) (Figures 8 and 9), ii) the Nd isotopic compositions of clinopyroxenes (Figure 10). Major (and trace) element compositions of minerals in all the spinel peridotites (both granular and tectonite types) are remarkably similar. In spite of this, the bulk-rock compositions define striking correlations, i.e. increasing FeO_{tot}, Ni, Co, and decreasing Al_2O_3, SiO_2, CaO, Sc, Cr, Yb_N, with increasing MgO, similar to those recently recognized in the abyssal peridotites (Niu et al., 1997; Asimow, 1999) (see Figure 8). The bulk rock chemical variations are coupled with systematic modal changes, namely progressive cpx, opx decrease and olivine increase, at increasing bulk MgO. These bulk-rock/mineral compositional contrasts suggest that the Erro-Tobbio spinel peridotites cannot be simply interpreted as mantle residues after variable partial melting degrees. Instead, they can be explained as resulting from combined histories of partial melting and subsequent melt migration, involving pyroxene dissolution and olivine precipitation reactions (Rampone et al., 2005).

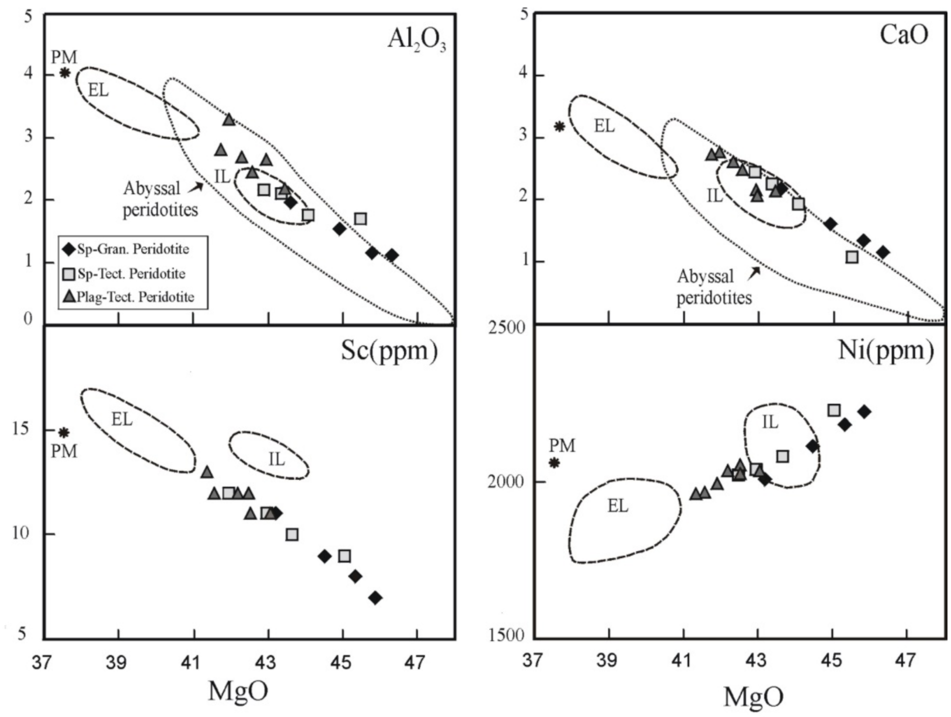


Figure 8 - Bulk rock variations of MgO (wt %) vs CaO, Al_2O_3, Sc (ppm) and Sc (ppm) in the Erro-Tobbio peridotites (Rampone et al, 2005). Abbreviations: Sp-Gran. = Granular spinel peridotites; Sp-Tect. = Spinel peridotite tectonites; Plg-Tect. = Plagioclase peridotite tectonites. Primordial mantle estimates are from Hofmann (1988). The representative compositions of abyssal peridotites are from Dick (1989).

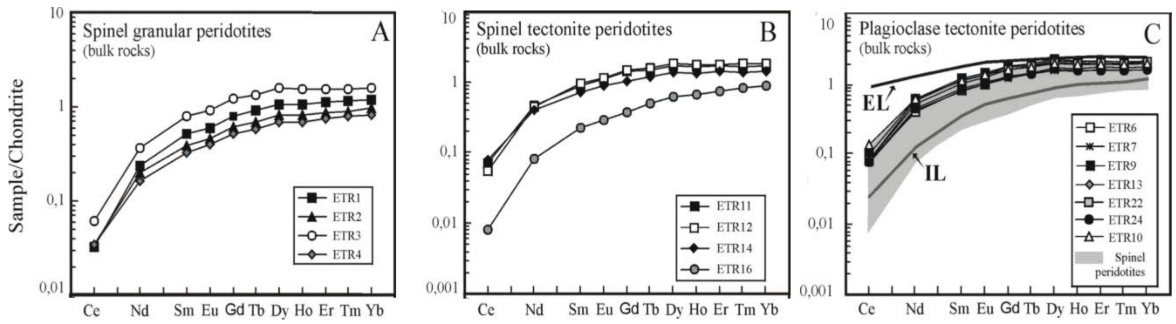


Figure 9 - REE patterns of the Erro-Tobbio peridotites (Rampone et al, 2004 2005).

In km-scale areas the spinel peridotites grade to **plagioclase peridotites** significantly enriched in plagioclase. The latter mostly occurs in veins and blebs interstitial to mantle olivine or crosscutting single olivine grains, suggesting that its crystallization was related to melt impregnation rather than subsolidus recrystallization (Borghini et al., 2007; Piccardo & Vissers, 2007; Rampone & Borghini,). Plagioclase enrichment occurs at the expense of both granular and tectonite spinel peridotite types. Peculiar microtextures like (i) crystallization of unstrained orthopyroxene (+ plagioclase) intergrowth partially replacing exolved and deformed mantle clinopyroxene, and (ii) occurrence of large undeformed poikilitic orthopyroxene replacing kinked mantle olivine, point that impregnation occurred after the development of the extension-related tectonite foliation. The bulk compositions of plagioclase peridotites plot at the "more fertile" end of all trends discussed above (higher Ca, Al, Sc, V, HREE, modal cpx contents) (see Figures 8 and 9). Thus, it could be argued that these peridotites were refertilized by impregnating melts crystallizing plagioclase. However, subsolidus plagioclase-facies recrystallization occurs in highly deformed plagioclase mylonites, likely after melt impregnation (Hoogerduijn Strating et al., 1993, Rampone et al., 2005). Sm-Nd isotope data, performed on two samples of plagioclase–bearing mylonites, have yielded two essentially parallel isochrons giving ages of 273 and 313 Ma (± 16 Ma) for the plagioclase-bearing recrystallization (Figure 10). These data could indicate that lithospheric extension and the mantle decompressional evolution were already active since late Carboniferous-Permian times (Romairone, 1999; Rampone et al., 2005).

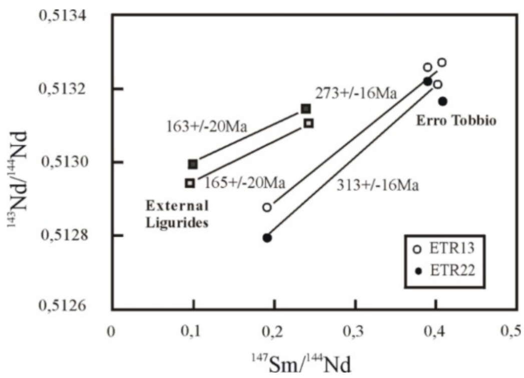


Figure 10 - $^{143}Nd/^{144}Nd$ vs $^{147}Sm/^{144}Nd$ diagram for clinopyroxene and plagioclase separates and whole rock from two samples (ETR13 and ETR22) of the Erro-Tobbio plagioclase tectonites (Rampone et al., 2005). Also shown are data for plag-cpx pairs from two External Liguride plagioclase peridotites (Rampone et al., 1995)

6.2 The intrusion of gabbroic bodies and dykes

The Erro-Tobbio peridotites are intruded by discrete bodies (up to 1 km wide) and dykes of gabbroic rocks (Figure 11), which are locally recrystallized to rodingitic and eclogitic assemblages, but frequently preserve magmatic textures and assemblages. The gabbroic bodies mostly consist of ultramafic cumulates and olivine gabbros; the dykes are constituted by troctolites, and olivine gabbros. Basaltic dykes are also found. The REE composition of magmatic clinopyroxene and whole rocks in all the above intrusives indicate a clear MORB affinity for the parental magmas. Sm–Nd geochronological data on MORB-type gabbros yield ages between 177 ± 7 Ma and 182 ± 19 Ma (initial εNd: 9–9.5), and a weighted mean age of 178 ± 5 Ma with initial εNd of 9.2 ± 0.4 (2σ) (Rampone et al., 2014).

Field evidence indicates that the subsolidus spinel- to plagioclase-facies transition, the melt percolation and entrapment, and the intrusions of MORB-type gabbroic bodies, occurred at deeper conditions, where mantle rheology still allowed plastic deformation and melt migration via porous flow. The subsequent dyke intrusion of MORB magmas, occurred at shallower levels, where the upwelling lithospheric mantle reached more rigid and fragile conditions because of the conductive heat loss.

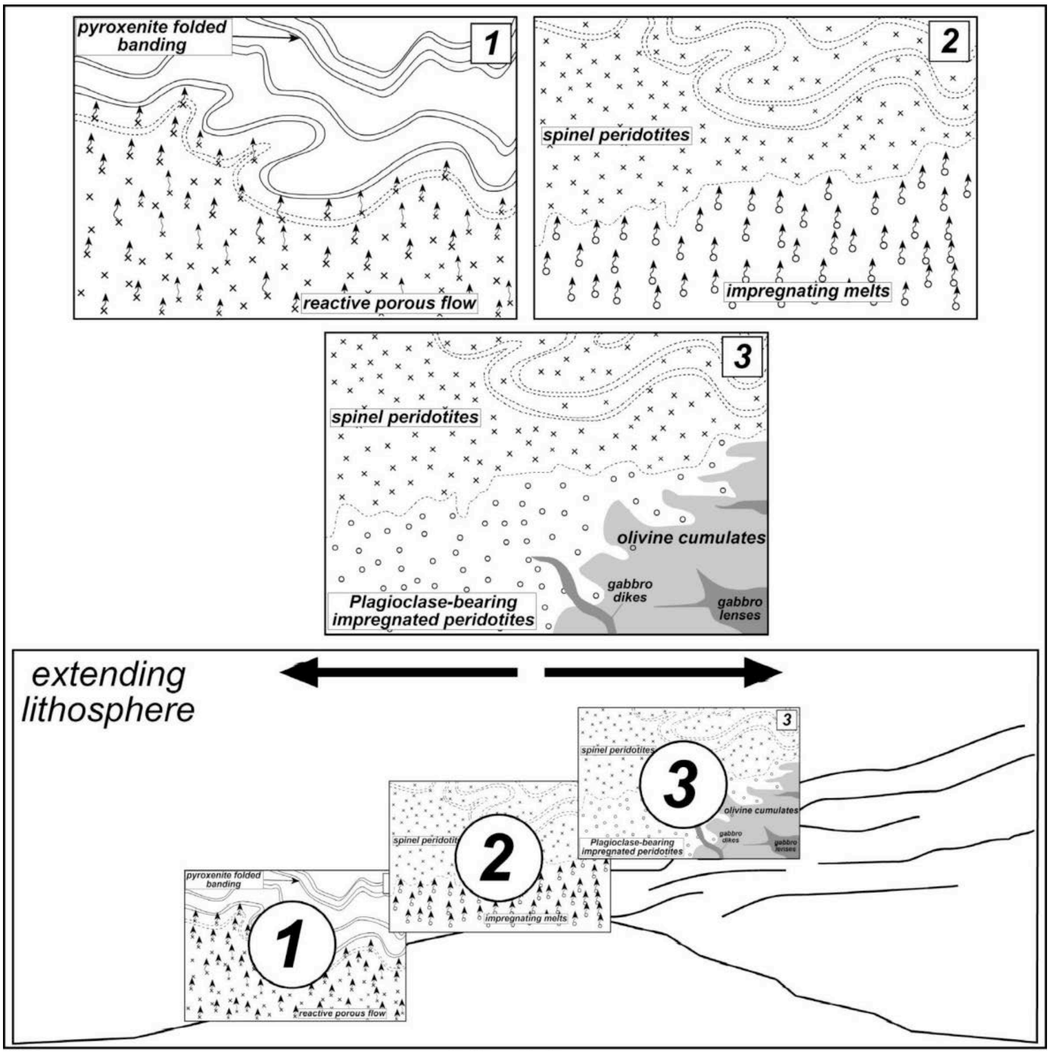
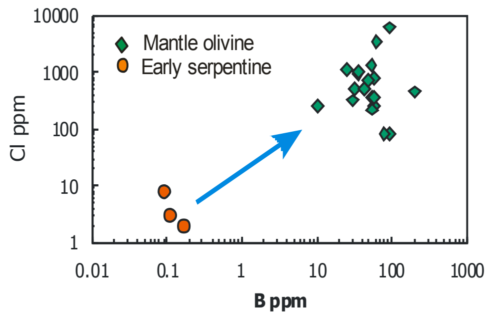


Figure 11: Scheme of the evolution of the different melt-rock interaction events observed in the Erro Tobbio ophiolitic unit (Rampone & Borghini, 2008)

6.3 The pre-subduction serpentinization

The shallow hydration of the mantle peridotites led to serpentinization and rodingitization of the mafic dykes (Cimmino et al., 1979; Piccardo et al., 1988). Hydrous assemblages including low-grade serpentine (mostly chrysotile and lizardite), chlorite, brucite and mixed-layer phyllosilicates statically replaced the mantle minerals and were in turn overgrown by subduction-related antigorite-bearing assemblages (Figure 12; Scambelluri et al., 1997). Chrysotile, lizardite and antigorite show variable Al_2O_3 (max 0.7 wt% in chrysotile; 2.5 to 7 wt% in lizardite; 0.5 to 3.2 wt% in antigorite) and variable B and Cl contents. Early (pre-subduction) chrysotile and lizardite can contain appreciable Cl and B (Figure 12). The mixed-layer phyllosilicates grew at the expense of mantle spinel, which can contain up to 5 wt% K_2O and 0.35 wt% Cl; they are overgrown by high-pressure chlorite lacking chlorine and alkalies.

The above features point to an early stage of peridotite interaction with sea-water-derived Cl-alkali-bearing solutions (Scambelluri et al., 1997; 2004).



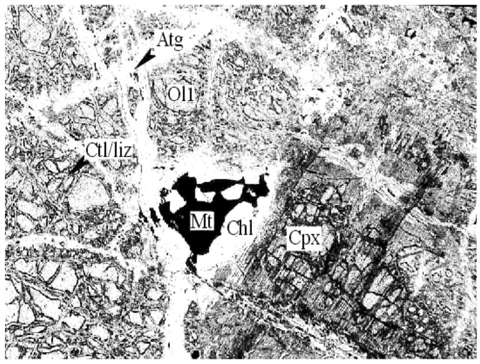


Figure 12 - B-Cl composition of mantle olivine and early chrysotile and lizardite from Erro-Tobbio. Right: serpentinized peridotite with relict mantle cpx and olivine replaced by early mesh chrysotile cut by antigorite veinlets. Mantle spinel is overgrown by low-grade magnetite and chlorite

The early serpentinites also show marked enrichment in bulk rock Sr to values much higher than in unaltered mantle peridotites and the oxygen isotope data available for the Erro-Tobbio early serpentintes are generally enriched in bulk ^{18}O (5.7 to 8.1‰). Altogether these data have suggested early low-temperature interaction with seawater-derived fluids. Recent work on the halogen composition of the Erro-Tobbio serpentinites indicate that alteration was not driven by simple seawater: the halogen study indicates input from seawater and sedimentary sources during hydration of either oceanic mantle at outer rise environments, or of fore-arc mantle affected by uprising slab fluids (John et al., 2011; Kendrick et al., 2011).

6.4 The alpine evolution

The Erro-Tobbio peridotites and the associated mafic rocks underwent subduction and eclogite-facies recrystallization (Piccardo et al., 1988; Hoogerduijn Strating et al., 1990; Scambelluri et al., 1991). Peak assemblages consist of omphacite + garnet + Mg-chloritoid + zoisite + talc + chlorite in metagabbros (Messiga et al., 1995), and olivine + titanian clinohumite + antigorite + diopside in peridotites. P-T estimates suggest that the eclogitic parageneses in metagabbros formed at 18-25 Kbar and 500-650^{o}C (Messiga et al., 1995). Structural studies (Scambelluri et al., 1991; 1995) have shown that eclogitization of metagabbros is coeval with formation of metamorphic olivine in the associated ultramafic rocks. During the Alpine cycle, deformation strongly affected the serpentinite mylonite domains: in domains outside the deformation horizons, subduction metamorphism developed statically and large patches of the pre-subduction features have been preserved.

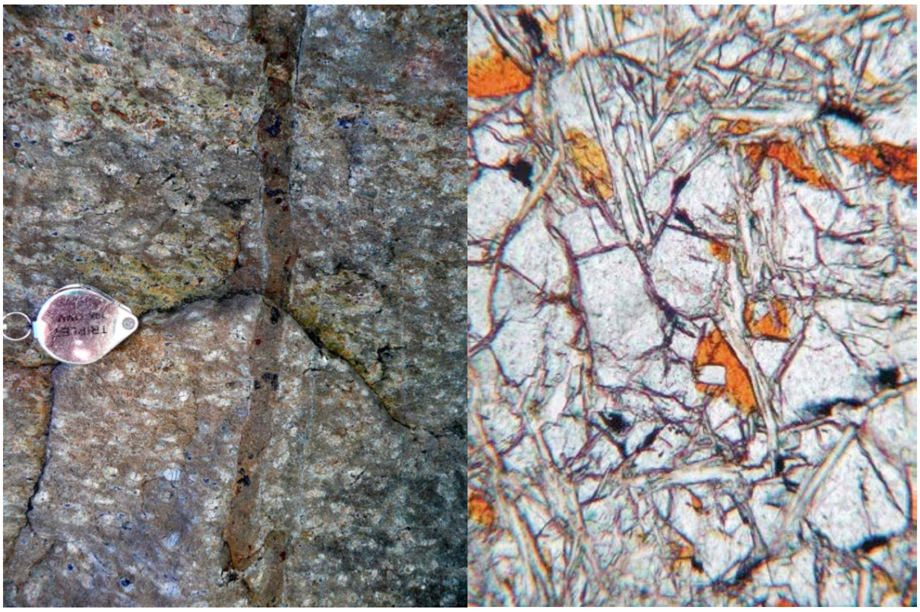


Figure 13 - Left: mesoscale view of undeformed serpentinized peridotite. The rock still preserves a mantle tectonite fabric (plunging from left to right in the picture) made of pyroxene porphyroclasts and spinel (black spots). The mantle fabric is statically overgrown by brown patches containing the high-pressure assemblage (olivine, antigorite, Ti-clinohumite; photomicrograph on the right) and is cut by a vertical vein containing the same assemblage + magnetite. Right: static high pressure assemblage with olivine (white), antigorite and Ti-Clinohumite (orange crystals).

In serpentinized peridotites that escaped alpine deformation, early serpentine cut by prograde antigorite veins can be observed. Antigorite is in turn cut by olivine + Ti-clinohumite + antigorite + diopside + chlorite + magnetite developed as static aggregates in the rocks and as vein assemblages (Figure 13).

Serpentinite mylonites record a polyphase Alpine dynamic recrystallization. Partially rodingitised mafic dikes and eclogitic metagabbros appear boudinaged in these domains indicating that deformation post-dated the magmatic crystallization and the subsequent Ca-metasomatism under low-grade conditions. This indicates that serpentine mylonites are related to Alpine deformation. On the basis of overprinting relationships (Figures 14 and 15) it has been established the following sequence of Alpine structures:

1 Antigorite mylonites: they display a slaty cleavage formed by antigorite and a stretching lineation marked by elongated magnetite. Associated shear bands are olivine-free and indicate top-to-the-NW sense of movement. The antigorite shear bands are only preserved in restricted domains.

2 En-echelon olivine veins, cutting antigorite mylonites. The orientation of veins, as well as locally developed mineral fibres in the veins (Figure 14a), point to a top-to-the-NW shear sense. The vein paragenesis consists of olivine + titanian clinohumite + diopside +chlorite + magnetite. Presence of olivine indicates the prograde character of the deformation with respect to olivine-free antigorite mylonites.

3. Multiple olivine shear bands: antigorite mylonites and en-echelon veins are overprinted by shear bands (Figure 14b) in which the olivine-bearing paragenesis is stable.

Orientation of the shear bands still indicates a top-to-the-NW shear sense suggesting that the structures formed during a continuous process. In zones with intense shearing a second set of olivine shear bands with a greater angle to the foliation developed. These multiple olivine shear bands provide evidence for continuous deformation (Platt and Vissers, 1980) in an overall top-to-the-NW sense of shear close to peak metamorphic conditions of about 600 °C and 2.0±2.5GPa (Figure 15).

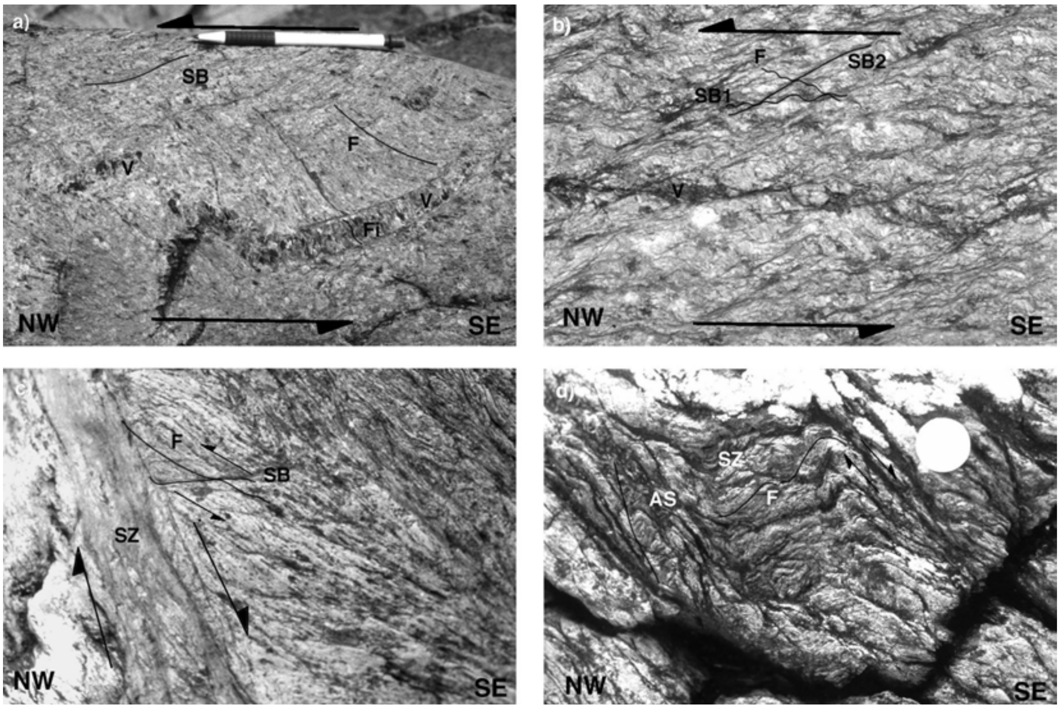


Figure 14 - (a) En-echelon olivine veins (V) with fibers (Fi) and olivine bearing shear bands (SB) crosscut the antigorite mylonite foliation (F) indicating a top-to-NW sense of movement (length of photo 30 cm). (b) Olivine shear bands (SB1) is cut by a second set (SB2). The multiple shear bands indicate continuous deformation during subduction. An olivine vein (V) is boudinaged by the olivine bearing shear bands (length of photo 15 cm). (c) A cm-wide top-SE shear zone (SZ) crosscuts the prograde structures. The orientation of the top-SE shear zone as well as the rotation of the foliation (F) into the shear zone indicate a top-to-the-SE sense of movement. The earlier formed olivine shear bands (SB), indicating a top-to-the-NW movement, are preserved in areas not affected by the S±C fabric (length of photo 10 cm). (d) The late stage NW-vergent folding (axial surface AS) overprints the S±C (SC) fabrics with still visible top-to-the-SE sense of shear and all previously formed structures (length of photo 25 cm).

Formation of peak metamorphic olivine is caused by reaction antigorite + brucite = olivine + fluid that brings to stability of olivine + antigorite + diopside + Ti-clinohumite + chlorite and to metamorphic fluid release. The most obvious evidence of fluid release is the development of the widespread olivine-bearing vein systems (Scambelluri et al., 1991; 1995).

Major implication of antigorite stability in HP ultramafic rocks is that they represent the most effective carriers of water into deep levels of subduction zones and that they maintain extremely low densities at mantle depths (Scambelluri et al., 1995; Hermann et al., 2001).

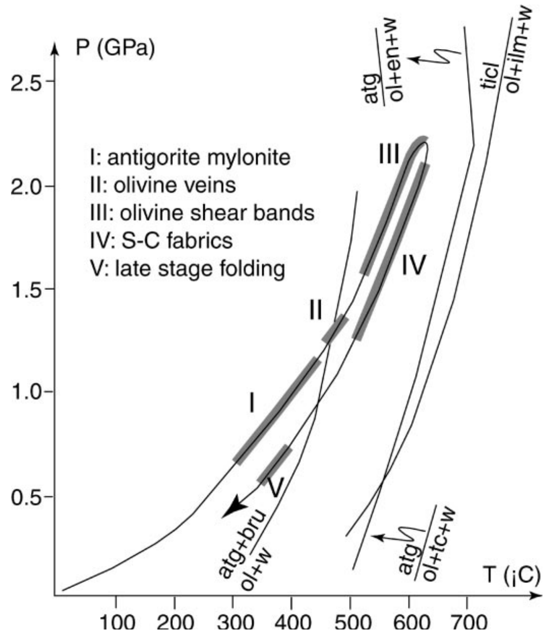


Figure 15. Pressure – temperature - deformation diagram for the Erro-Tobbio unit. Reactions (after Hermann et al., 2000)

6.5 Fluid phase and element mobility during high-pressure dehydration

The shallow hydration of the Erro-Tobbio peridotite has relevant consequences on deep water transport during subsequent subduction and on element cycling into the mantle by release of deep de-serpentinization fluids. As shown in Figure 15, the Erro-Tobbio subduction path crosses a first olivine-in dehydration reaction that produces the fluid phase drained by the vein systems shown in Figures 13 and 14.

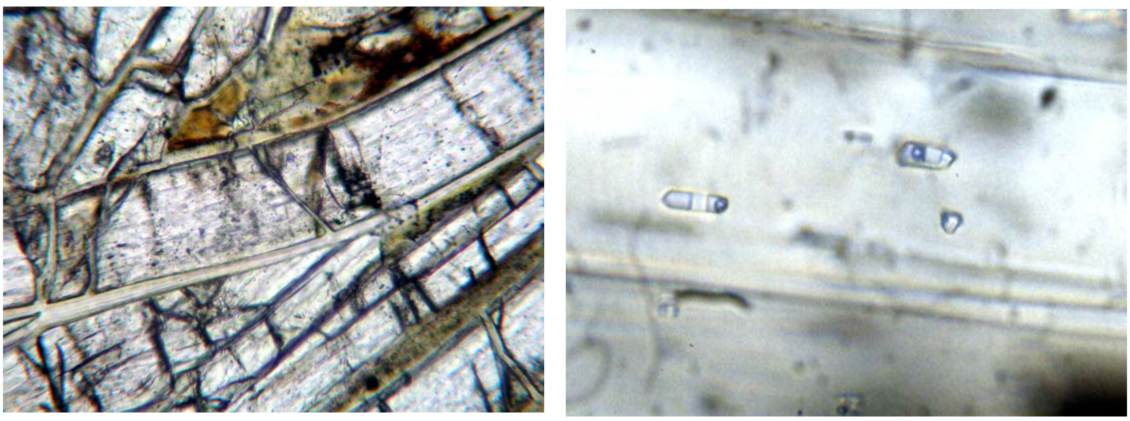


Figure 16 - Left. Deformed vein of diopside. The inclusions follow the deformed cleavages. Right: primary aqueous inclusion with salt daughter crystals following the diopside cleavage.

Remnants of this fluid are inclusions in vein forming minerals, like diopside and olivine (Figure 16). Inclusions in diopside line along the cleavage: in vein samples deformed by the olivine shear bands of stage III (Figure 15) the inclusions follow the deformation patterns of diopside. This implies that they pre-date the high-pressure deformation and are primary (Scambelluri et al., 1997). These inclusions are aqueous brines that frequently contain a salt daughter crystal (Figure 16) with salinity of 40-45 wt% NaCl equivalents.

Uptake of chlorine and alkalis in early serpentine and other alteration phases during shallow peridotite hydration and finding of salt component in the subduction fluid provides strong evidence for deep halogen recycling in high-pressure fluids (Scambelluri et al., 1997).

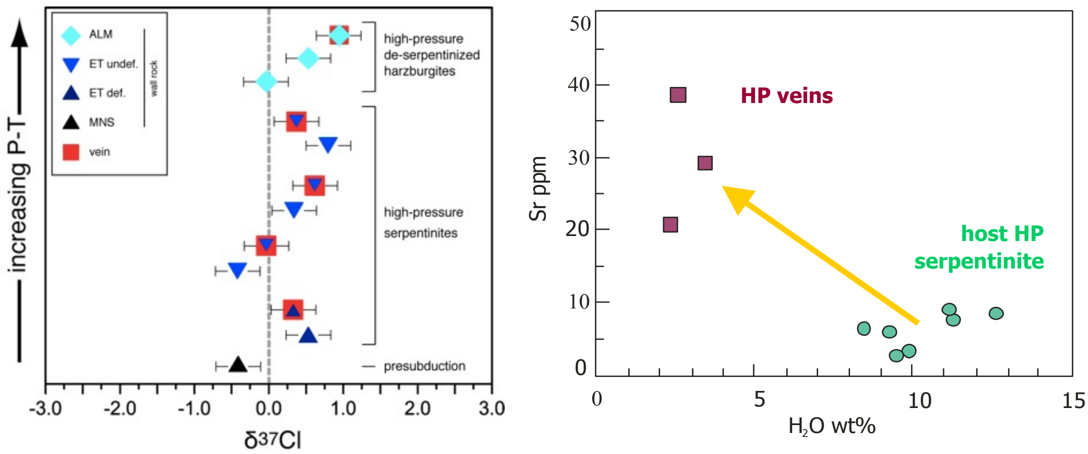


Figure 17 - Right: Cl isotope composition of the Erro-Tobbio serpentintes (John et al., 2011). Left: Sr content of high pressure Erro-Tobbio serpentintes and of associated veins

This is also confirmed by recent halogen analysis in these rocks (John et al., 2011), also showing that the overall dehydration process does not produce significant Cl isotope fractionation (Figure 17). Same as Cl, Sr is recycled from rocks into the high-pressure fluid (Figure 17; Scambelluri et al. 2001) as witnessed by appreciable Sr contents in veins (Figure 17, 18). The similarity of trace element compositions of high pressure ultramafites and veins (Figure 18) indicates internal fluid and element recycling into vein systems at the time of dehydration. The recognition that pre-subduction water, chlorine, alkalis and strontium were carried by the vein fluid, indicate closed system behaviour during eclogitization and internal cycling at 80 km depth of exogenic components.

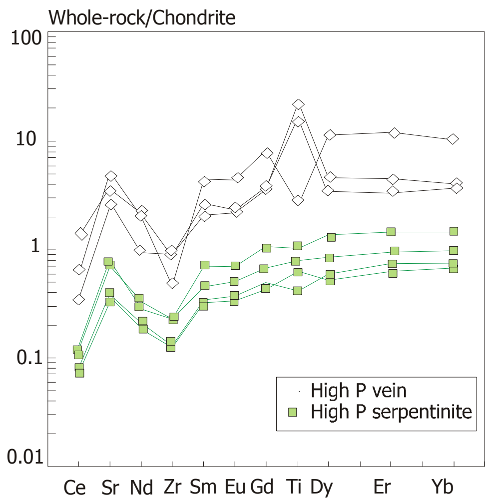


Figure 18 - REE and Sr composition of the Erro-Tobbio HP serpentintes and veins (Scambelluri et al., 2001).

7. Recent data and research address on the Erro-Tobbio subduction context

The Erro Tobbio peridotite has long been considered as an exhumed slice of subducted oceanic lithosphere. Overall, the mantle peridotite geochemistry, the exhumation history to shallow oceanic environments and the gabbro affinity, call for an oceanic pre-subduction context, a slow spreading ridge or an ocean-continent transition environment (Rampone and Borghini, 2008; Piccardo and Vissers, 2007). Early serpentinization has been associated to interaction with seawater in oceanic environments (Frueh Green et al., 2001). The subduction story clearly shows a prograde evolution along a cold subduction geotherm accompanied by element loss at the time of dehydration (olivine formation) and veining (Scambelluri et al. 1995, 2001, 2004). This has been interpreted as the story of a downgoing serpentinized oceanic slab.

However, recent geochemical work (John et al., 2011; Kendick et al., 2011) indicates that early Erro-Tobbio serpentinization was driven by fluids that exchanged with sediments and/or with sedimentary pore fluids. The suggested formation environments are either the outer rise, where fluids penetrate downwards along megacracks in the lithosphere (e.g. Ranero et al., 2003), or the fore-arc mantle, where sediment compaction and shallow dehydration release slab fluids hydrating the overlying mantle. These works thus shift the serpentinzation environment from internal oceanic settings to domains close to the subduction zone.

Other perspectives arise from the application of B isotope systematics to these rocks (Tonarini and Scambelluri, 2012). B isotopes fractionate during subduction dehydration and/or phase changes leading to water loss, such as transition from chrysotile to antigorite (e.g. Kodolanji and Pettke, 2011). ^{11}B is released to fluids and the rock residues become increasingly depleted in ^{11}B assuming negative δ^{11}B values. One can thus expect progressive ^{11}B depletion of the Erro-Tobbio serpentinites during burial from low-grade chrysotile- to antigorite-bearing serpentintes. The data produced (Tonarini and Scambelluri, 2012) indicate that the δ^{11}B of these rocks is heavy (16‰ to + 24‰ δ^{11}B) and no fractionation occurs during transition from early low-grade to high-pressure serpentintes. Rather, it appears that the antigorite mylonite are richer in water and Sr and display high d^{11}B compared with the pre-subduction varieties. This suggest hydration, rather than fluid loss, during transition from early chrysotile to antigorite assemblages at increasing T and P. In general, $^{87}Sr/^{86}Sr$ ranges from 0.7044 to 0.7065, i.e. lower than oceanic serpentinites formed from seawater. These data suggest two implications:

First, the analyzed HP serpentintes are ^{11}B reservoirs for subduction. They maintain high ^{11}B down to a first fluid release veining event, thus the $\delta^{11}B$ of olivine veins that fingerprint the composition of released fluids, justify high $\delta^{11}B$ measured in many arc lavas. Fluids released by dehydration of such HP serpentinites by full antigorite breakdown should have up to + 20‰ $\delta^{11}B$. The comparable ^{11}B fingerprint of serpentintes, their fluids and arc lavas provides a strong link between serpentinite subduction and arc magmatism.

Second, the combination of δD, $\delta^{11}B$, $^{87}Sr/^{86}Sr$ of the HP serpentinite apparently favour their location above the subducting slab. This implies that serpentinites were formed by low-T fluids likely arising from a subducting lower plate and that during their subduction history they were (or became) part of an environment located above the slab.

The application of B isotope systematics may thus open new perspectives on the tectonic location of high-pressure rocks in orogenic belts and to identify the provenance of the various tectonic slices (whether a slab, or a mantle wedge) which are part of plate interface environments.

Field itinerary

It has been attempted in this excursion to present the various structural and petrological aspects of the ET peridotite in their relative order of development with time. All stops have a general theme which is considered to be most clearly illustrated in those localities. All stops are indicated on the map of the region.

STOP 1: Theme: Granular peridotites and low strain tectonites

Locality: Along the road from Voltaggio to Capanne di Marcarolo, northwest of the Mt. Tobbio, where the road bends from EW to NS.

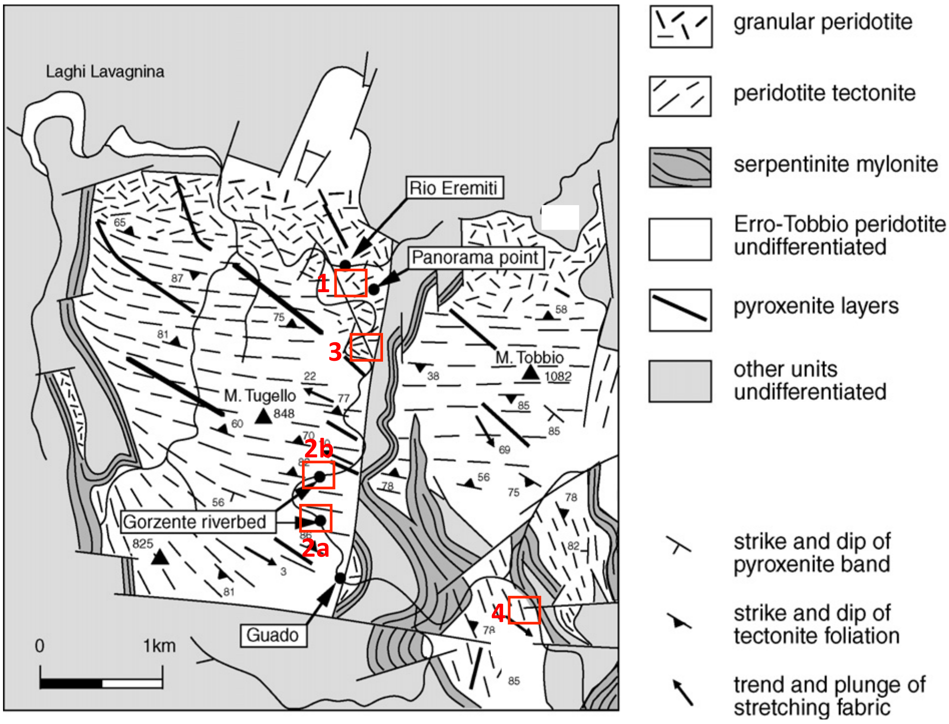


Figure 19: Structural map of the Mount Tobbio area, with indicated STOPS.

The structural map of Figure 19 shows the local geology in detail. Within this area, a transition has been mapped between granular peridotites and peridotite tectonites. This transition is remarkably gradual, from virtually undeformed granular peridotite in the North, through low- and intermediate-strain tectonites to distinctly high-strain peridotite tectonites to the South. The outcrops visited on this stop are within the northern part of this transition, and show almost undeformed granular as well as relatively weakly deformed tectonite structures. The area is further dissected by brittle dextral and sinistral oblique-slip faults affecting the primary contact between the ET and Beigua units. One of these faults runs approximately parallel to the general trends of the road and river. Within the northern part of the map area an EW trending region occurs in which no foliations are developed. The peridotites in this part of the area have a coarse-grained (0.5-1cm) granular texture, and all mineral phases have a similar grain size. In outcrops of such virtually undeformed granular peridotites, orthopyroxene, spinel and clinopyroxene often occur in characteristic clusters (Figure 20) as described by Mercier & Nicolas (1975) and Nicolas (1986a,b). These clusters are thought to originate from precursor garnets decomposed during depressurization and recrystallization in the spinel-lherzolite stability field (Green & Burnley, 1988). Unfortunately such clusters are not very clear at the outcrop scale. The microstructure of the granular peridotites is illustrated in Figure 20 and 21a. The olivine and pyroxene grains show minor internal deformation characterized by very open kinks and most grains show a weak undulatory extinction. The microstructure of weakly to moderately foliated, low-strain tectonites is dominated by large olivine grains showing well-developed deformation sub-structures of sharply defined, high angle kink-like subgrains with undulatory extinction (Figure 21b). The progressive development of the peridotite microstructure with increasing tectonite strain is discussed below (Stop 2).

Figure 20: Left: Outcrop of granular peridotites showing no clear foliation and the coarse-grained granular texture. Right: Pyroxenite layering within the granular peridotite

Plagioclase-bearing assemblages have not been found in the peridotites of this area. Instead, the spinel is commonly replaced by magnetite surrounded by a white weathering Mg-chlorite halo. In fresh peridotites all contacts between greenish-brown spinel and clinopyroxene are sharp without any reaction product.

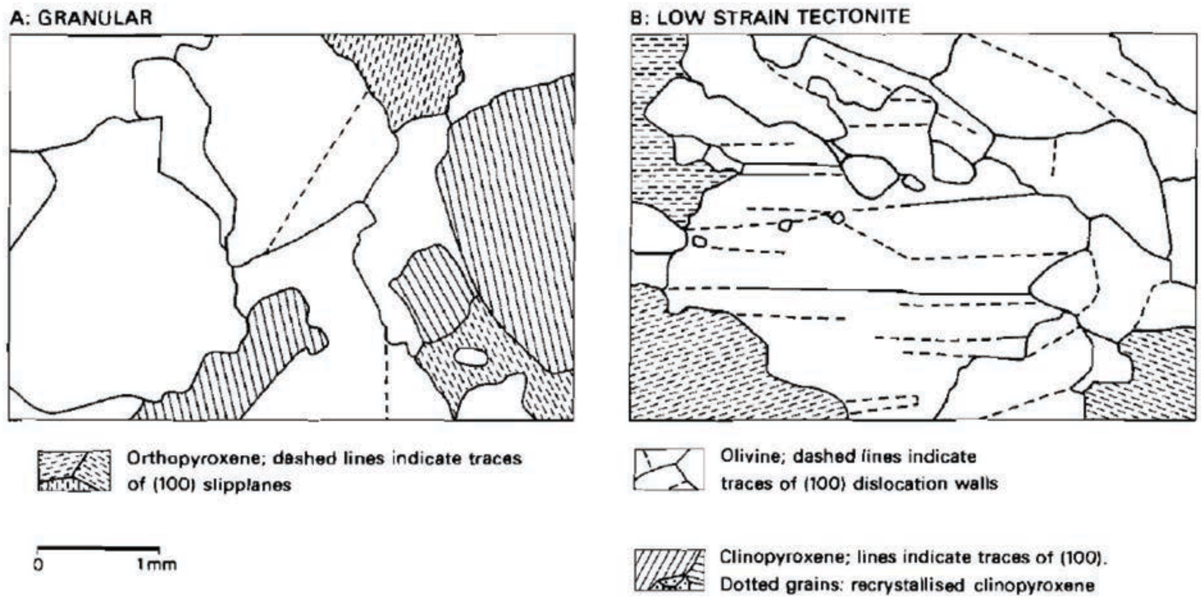


Figure 21: Line drawings from thin sections showing the microstructure of (A) granular peridotites, and (B) low-strain peridotite tectonites (Hoogerduijn Strating, 1991).

STOP 2: theme: Peridotite tectonites, pyroxenite and dunite layers.

Locality: park cars near the guado (where the road to Capanne di Marcarolo crosses the Torrente Gorzente). Follow the track on the right hand side just before (North) the guado. This track leads to locality 2 after about 200-300m. Via track, slightly above the riverbed, a further 250m leads to locality 2b.

In this area, the ET peridotites have a well-developed tectonite foliation (S_t) defined by the shape preferred orientation of deformed pyroxene, spinel and olivine grains. Diffuse withish aggregates (mostly constituted by chlorite), distributed along the foliation are likely relics of previous plagioclase. The tectonite deformation therefore occurred at the expences of previously impregnated plag-rich peridotites. In this part of the area (Figure 19) this foliation has a relatively constant orientation on a kilometre-scale, and strikes between E-W and SE-NW with steep to moderate dips. Stretching lineations in the plane of the tectonite foliation are common but their orientations may vary, In the Mt Tobbio area, for example this lineation is subhorizontal in the River Gorzente section but subvertical on Mt. Tobbio's East side. The Northern Gorzente section shows a gradual transition from granular peridotites in the N near stop 1 to peridotite tectonites further south. Apart from a clearly increasing intensity of the tectonite foliation across this transition, an increasing strain is strongly suggested by a gradually decreasing angle between the early pyroxenite layers and the tectonite foliation. In low-strain tectonites with a moderately developed foliation, the primary layering is oblique to S_t (whereas, for example in intensely foliated tectonites from the southernmost part of the Gorzente section this primary layering is virtually parallel to S_t. From inspection of the map it is evident that the peridotite tectonites occur within kilometre-scale zones of localized deformation, and the gradual nature of the transition in relatively undisturbed parts of the section indicates that this transition between granular

and tectonite peridotites does not result from juxtaposition along a later fault or serpentinite mylonite zone.
During our walk along the river bed we will concentrate on two outcrops in particular. Outcrop 2a shows nice example of the tectonite foliation on polished surfaces in the riverbed. The tectonites contain porphyroclast systems with a tendency to asymmetric sigma-type structures. Outcrop 2b shows the only evidence obtained so far for pre-tectonite folding (Figure 23). The following section aims to describe the main microstructural and petrologic aspects of the peridotite tectonites without particular references to the selected outcrops.

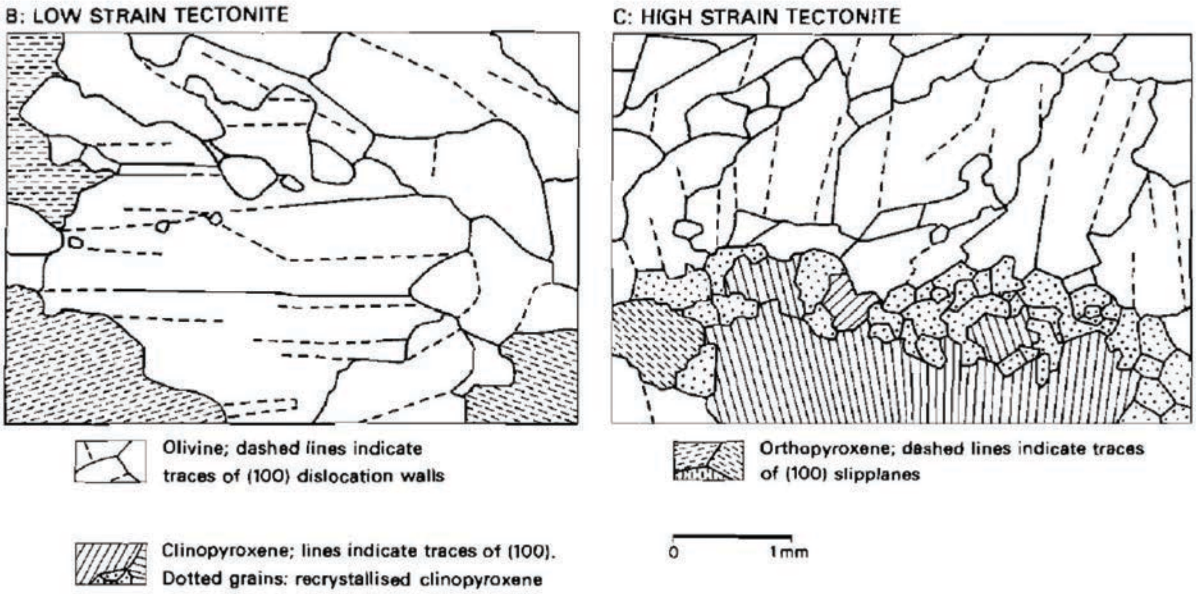


Figure 22: Line drawings from thin sections showing the microstructure of (A) low-strain and (B) high-strain peridotite tectonites (Hoogerduijn Strating, 1991).

It has already been emphasized at STOP 1 that the microstructure of the low-strain tectonites is dominated by large olivine grains with well-developed deformation substructures of sharply defined kink-like subgrains and undulatory extinction (Figure 21a). Large olivine grains have not been preserved in tectonites with a more intensely developed foliation. Instead, the microstructure is characterized by smaller (0.5-1.5mm) elongate to tabular shaped grained (Figure 22b) with weakly curved to straight boundaries and moderate development of sub-boundaries and undulatory extinction. This microstructural transition, allied to the transition from granular peridotites to peridotite tectonites, is interpreted here as the result of grain size reduction during dynamic recrystallization.

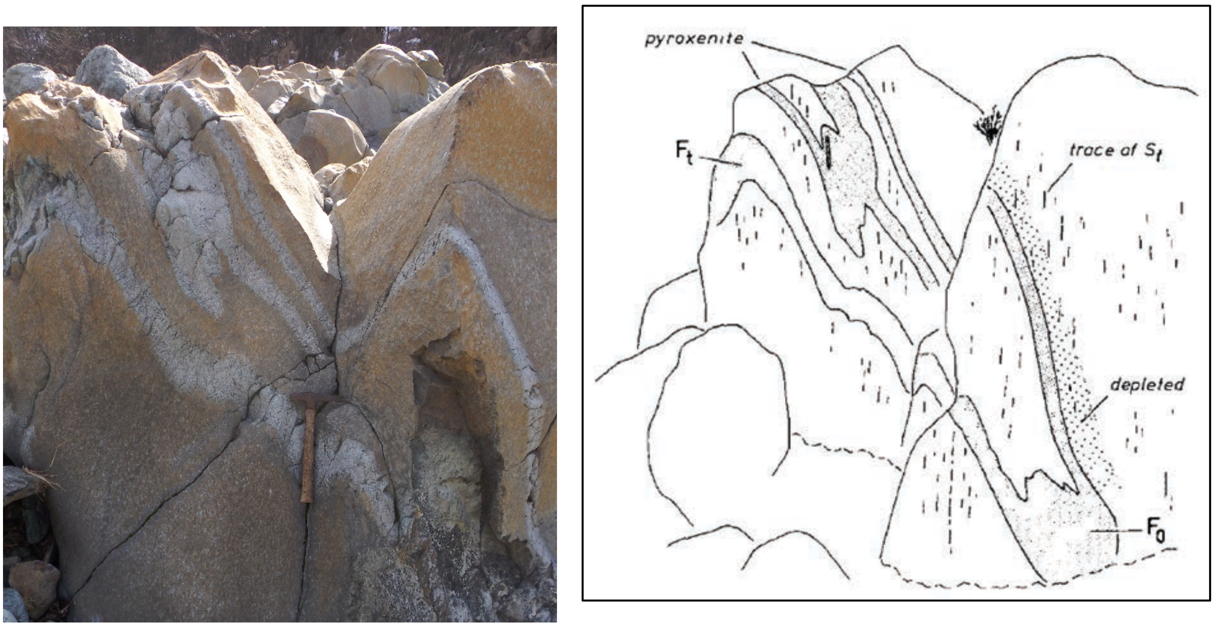


Figure 23: Sketch after field photograph of type III fold interference in peridotite tectonites. Note that tectonite foliation is axial planar to the upright (youngest) generation of folds (Hoogerduijn Strating et al, 1992)

In some tectonites a second foliation oblique to S_t is defined by the shape preferred orientation of elongate new olivine grains. The microstructure of these rocks is similar to that developed in certain types of S-C quartz mylonites (Lister & Snoke, 1984), and we envisage that it can potentially be used as a shear sense indicator. A fabric study in one such S-C type tectonite is illustrated in Figure 24, and shows a marked point concentration of [a] axes at angles of 15-20° to the pyroxene foliation and partial girdles of the [b] and [c] axes normal to the [a] maximum. The sense of shear indicated by the fabric is the same as that indicated by the oblique olivine grain shape foliation. These fabrics indicate that the dominant slip system allowing deformation in the tectonites was the high temperature [a](0kl) olivine system.

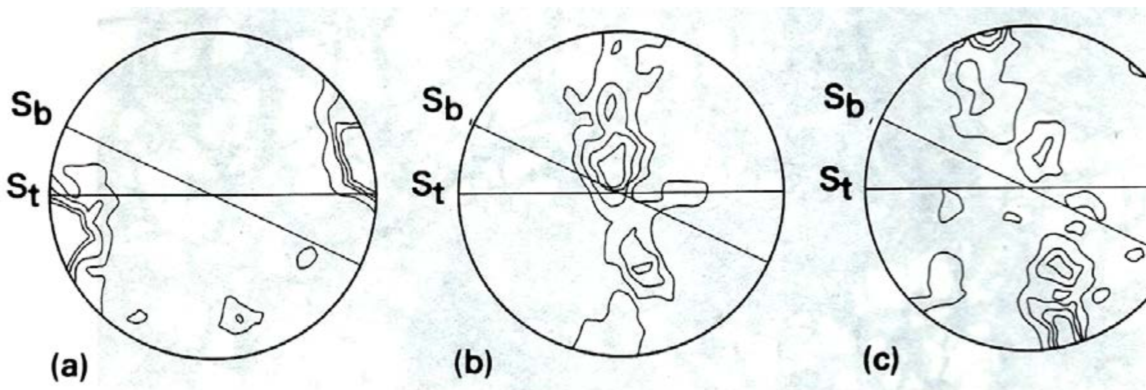


Figure24: Lattice preferred orientation patterns in peridotite tectonite. Note distinct maximum of [a] axes and girdle distribution of [b] and [c] axes.

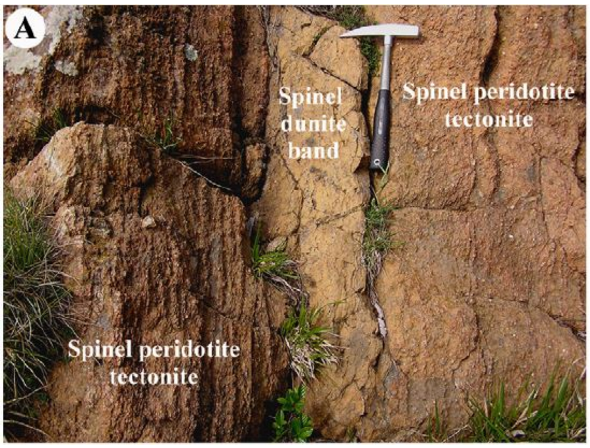
In summary the tectonite microstructures in the ET peridotites range from porphyroclastic to tabular (Mercier & Nicolas, 1975, Harte, 1977). These microstructures are similar to those developed in the Lanzo massif (Nicolas & Poirier, 1976) and some basalt xenoliths (Avé Lallement, 1985; Mercier & Nicolas, 1975). The elongate new olivine grains in the highly strained tectonites have similar sizes as the kink-like sub-grains developed at lower strains, suggesting that recrystallization involved an important component of subgrain rotation (Poirier & Nicolas, 1975). The mineral assemblage in the early tectonites is the same as that in the granular peridotites (i.e. spinel facies assemblage), indicating that the tectonite deformation of the ET lherzolite started in the spinel-lherzolite facies.

The pyroxenite layers represent deep-seated intrusions preceding the lithospheric exhumation history of the ET mantle. They display variably fractionated REE spectra. Unusual trace element signature (high Sc,V contents and low MREE/HREE ratios) in clinopyroxenes from one pyroxenite layer is witness of a precursor garnet-bearing magmatic assemblage (Rampone and Borghini 2008).

Both protogranular-porphyroclastic spinel lherzolites and spinel peridotite tectonite-mylonites are cut by elongate bands and bodies of spinel dunites. These dunites occur as metre- to decametre-wide discordant dunite masses crosscutting the foliation of previous rock types, and as elongated, dm- to metre-wide bands in sheared peridotites (outcrop of Stop 2a), with the bands being concordant with the main foliation of the surrounding peridotite tectonites and mylonites (Figure 25).

In other outcrops, the spinel foliation and pyroxenite banding in the host peridotite crosscut the contact between the peridotite and dunite and continue within the dunite as elongated Sp trains. This clearly suggests a replacive origin of the spinel dunites (see also Boudier and Nicolas, 1972; Boudier, 1978), i.e. these dunite channels have been formed by the focused and reactive migration of pyroxene-undersaturated melts across pre-existing compositional or structural discontinuities (Kelemen et al, 1995).

Concordant bands of dunite occurring in sheared peridotites indicate that replacive dunite bands followed pre-existing mylonitic bands This suggests that, within the shear zones, focused melt migration was enhanced by the deformation fabrics (defined by thin bands of strongly reduced grain size, and by the preferred shape orientation of the constituent grains and grain aggregates) of the percolated peridotite mylonites; accordingly, focused melt percolation was in such cases controlled by the pre-existing deformational structure in the mylonites (Piccardo and Vissers, 2007).



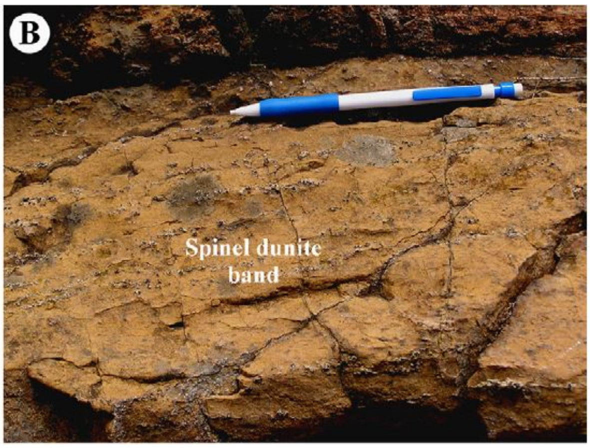


Figure 25: Spinel dunites and crosscutting relationships with surrounding peridotites

The pyroxenite layers as well as the enclosing peridotites have been slightly to severely recrystallized during both shallow depth (pre-alpine) and alpine evolution. As a rule Mg-chlorite, serpentine minerals and oxides replace previous orthopyroxenes and spinels, while mantle clinopyroxene relics are still preserved although frequently overgrown by diopside and/or tremolite. Later crosscutting veins and fractures showing Ol + Ti-Cl + Mag + Di + Mg-Cht associations reflect the evolution through the Alpine climax conditions.

STOP 3 - ***Gorzente River South, road West of Mt. Tobbio (20 minute walk along a foot path down the Gorzente River).***

Static recrystallization zone of high-pressure Alpine paragenesis on peridotites and rodingitized gabbros.

This outcrop shows the hundred metre-scale exposure of alpine metaperidotite preserving mantle granular textures and mineralogical relics of the mantle protoliths. Peridotite is intruded by mafic dykes. The Alpine recrystallization of peridotites occurred in absence of deformation and was accompanied by brittle deformation originating vein systems. The mafic dikes are rodingitized and display a pervasive metamorphic foliation. The geometric relationship between veins and foliations in the two rock types suggest that during subduction and HP recrystallization, plastic deformation of the mafic dikes was coeval with brittle deformation of the enclosing peridotite.

In thin section, the peridotite mantle minerals are partly overgrown by neoblastic Alpine assemblages. Spinel shows thin coronas of Mg-chlorite; clinopyroxene is replaced at the rim and cleavages by fine-grained diopside, olivine, Ti-clinohumite and subordinated antigorite; coarse mantle olivine is overgrown by radial antigorite in textural equilibrium with fine-grained olivine and Ti-clinohumite neoblasts. The coarse mantle olivine relics locally still record a sequence of overprinting serpentine-bearing assemblages: an old one made of chrysotile + magnetite + brucite in mesh-textures; the mesh serpentine is cut by veins and microfractures with antigorite that are finally cut by fractures containing olivine.

This sequence of assemblages indicates increase in metamorphic grade and prograde metamorphism from shallow partial serpentinization that culminates in peak HP formation of olivine + antigorite + Ti-clinohumite + diopside + chlorite. The same peak assemblage develops in the vein systems diffusely cutting the peridotite and representing the most evident feature of the Alpine peak metamorphism.

The rodingitized mafic dykes likely derive from pristine Mg-Al and Fe-Ti-gabbros. They show clinopyroxene porphyroclasts in a foliated diopside + chlorite + clinozoisite + magnetite matrix. The margins in touch with the peridotites are marked by foliated walls with Mg-chlorite and epidote. The above assemblage typically forms at HP conditions during recrystallization of mafic rocks which underwent pre-subduction rodingitization (Ca-enrichment stage).

Comparable prograde evolution is observed in a serpentinite mylonite shear zone within the peridotite on the eastern side of the river. In this case olivine + Ti-clinohumite + antigorite shear bands overprint a previous antigorite (olivine-free) shear foliation and point to extreme channelling of plastic deformation along this mylonite horizon during the prograde alpine subduction history. This mechanism of deformation partitioning allowed preservation of large domains where the mantle structures and assemblages have survived due to static and incomplete recrystallization to alpine metamorphic assemblages.

Concerning the P-T conditions of peak HP recrystallization of the Erro-Tobbio peridotite, information on the confining temperature can be deduced by the stability of antigorite in equilibrium with olivine and Ti-clinohumite, indicating 600-650°C at 2-2.5GPa. More reliable P for the HP climax event have been obtained in mafic systems: ET eclogitic metagabbro, which escaped rodingitization and recristallyzed to omphacite + garnet + chloritoid + talc during subduction, give pressure-temperature estimates of 2.0-2.5 GPa and 500-650°C (Messiga et al., 1995).

Static Serpentinite dehydration networks and fluid channelling structures

These features can be observed in a small part of the Erro Tobbio metaserpetinite (ET-MS): here a massive serpentinite (virtually undeformed during subduction) deriving from almost full serpentinzation of the granular peridotite observed in the first part of the outcrop shows a network of olivine veins.

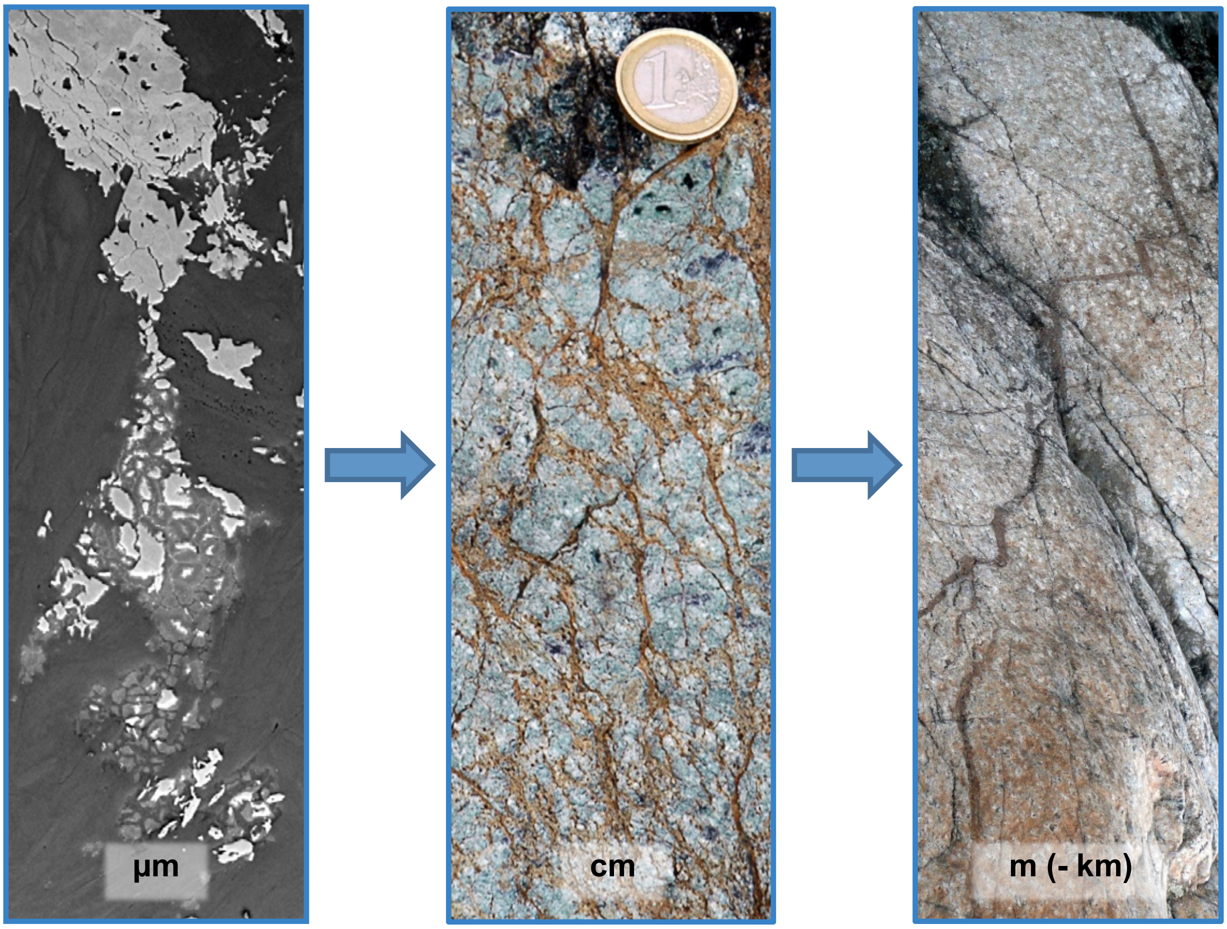


Figure 28 - Different scale view of olivine + magnetite + Ti-clinohumite veins cutting almost undeformed antigorite serpentinite. The µm-size veins (bright in the BSE image) feed the cm-size veins (reddish), which in turn develop into the m-size vein network that likely represents the structures of fluid release. The medium-bright greyish material at the tip of the µm vein consists of a disordered serpentine phase, which is so far only known as a metastable dehydration product of antigorite during serpentinite dehydration experiments.

Same as in other parts of this outcrop the serpentinite ET-MS underwent high-pressure partial dehydration of antigorite + brucite to produce olivine ± titanian-clinohumite + fluid in the presence of stable antigorite (Scambelluri et al. 1991) via a reaction such as:

1 Antigorite + 20 Brucite = 34 Olivine + 51 H_2O.

As this reaction progresses, antigorite dehydration results in a positive total volume change ((ΔV_{rxn} = +5.1%) and a net decrease in the solid volume (ΔV_s = -41.9%) (Healy et al. 2009), leading to an irreversible porosity increase and the generation of a fluid pore pressure excess (ΔV_f = +47.0%), i.e. fluid overpressure that eventual leads to fluid flow (Darcy flow).

Prograde olivine-bearing vein networks occur in this outcrop as undeformed veins within the bulk serpentinite (Figures 28 and 29). A striking characteristic of the ET-MS is the strong channelization and coalescence of veins across all scales (µm to m) to form a dehydration vein network (Figure 28). Vein coalescence occurs via side branching, where small channels join larger ones at angles between 80 to 100°, and infrequently two veins join in pairs another vein at an acute angle. The vein network exhibits topological properties that are intermediate between river and hierarchical vein networks, suggesting high transport efficiency (Kobchenko et al. 2013).

The process of vein network formation (Figures 28, 29) starts in fluid source regions located at the vein tips (Figure 28 left panel). These microsites show the common occurrence of multiple, micro- to nanoporous patches ahead of the vein tip (medium grey material in Figure 28, left panel), consisting of a polycrystalline aggregate of hydrous phyllosilicate grains (± secondary olivine). Isolated patches range in size from few µm to several tens of µm. Individual patches join up via µm-sized seams or overlap to form larger, interconnected areas (Figure 28). Incorporation of broken-up bulk antigorite debris into the patches and microcrack initiation from patches into the matrix antigorite imply that patches form as a result of reaction-front propagation into antigorite during its breakdown. Placing the DI-PP (disordered phyllosilicate phase; Pluemper et al., 2016) occurrence as an intermediate reaction step between antigorite breakdown and olivine nucleation, is confirmed by the presence of DI-PP as inclusions within the secondary olivine. Olivine grains are always found in connection with the vein and never isolated in the antigorite matrix. In the absence of DI-PP the antigorite matrix is free of porosity. Interconnected veins develop rapidly away from the fluid source regions (Figure 29). Together with the presence of nanocrystalline olivine grains within the DI-PP it is evident that olivine crystals grow on the expense of antigorite, where fluid is transported away from the dehydration front through the porous DI-PP network. Thus the veins grow like roots into the dehydrating serpentinite, while the vein material reflects the reaction product produced by this dehydration.

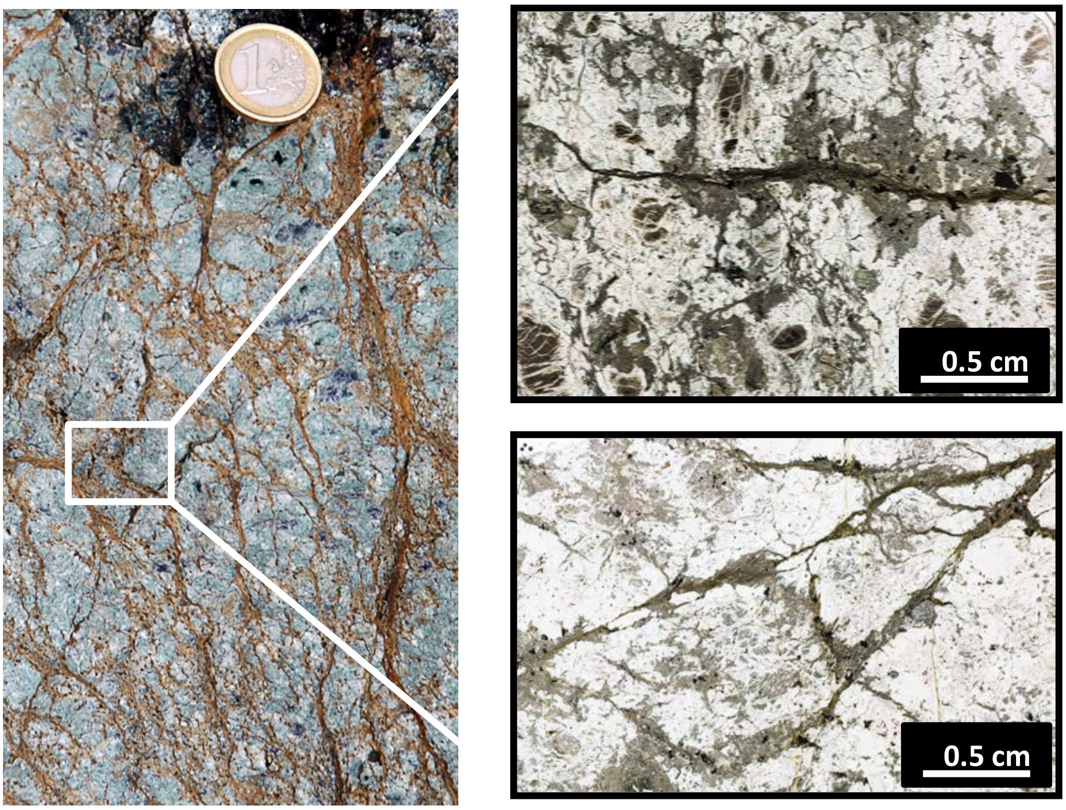


Figure 29 - Olivine + magnetite + Ti-clinohumite veins at various scales, showing that the small veins terminate in larger veins developing a vein network.

STOP 4 – ***Gorzente River North, east of the guado (30 minute walk following the path along river).***

The Alpine subduction evolution: high-pressure serpentinite mylonites**.**

These outcrops display mylonitic antigorite-foliations, shear bands and olivine-veins. The serpentine shear zones envelop undeformed and less serpentinized metre- to kilometre-scale mantle peridotite bodies.

The first outcrop is represented by a hundred metre-scale peridotite body preserving mantle tectonite structures and a spinel to plagioclase paragenesis. Pyroxenite bands are parallel to the mantle foliation. This latter is marked by the orientation of clinopyroxene and orthopyroxene porphyroclasts and by small spinel crystals. Mantle olivine forms the matrix of these rocks: it is diffusely overgrown by chrysotile and lizardite. Plagioclase is overgrown by chlorite micro-aggregates. These represent the mantle plus low-grade alteration textures.

Acicular antigorite associated with metamorphic olivine and Ti-clinohumite develop inside this mantle peridotite body as the result of static break-down of mantle minerals and chrysotile, during the Alpine peak metamorphism. In thin section, relics of mantle olivine show development of antigorite + magnetite after precursor mesh-textures chrysotile veins, followed by crystallization of neoblastic Alpine olivine (Figure 26 left). This indicates that early in the story the peridotite underwent static low-temperature serpentinization followed antigorite + magnetite formation and to peak olivine + Ti-clinohumite + antigorite assemblage, which mark the temperature increase during Alpine subduction.

Figure 26 - Left: mantle olivine (large crystal) overgrown by early mesh serpentine (chrysotile/lizardite). These are cut by veins with antigorite, in turn cut by a veinlet of new olivine (ol2). Right: eclogitic metagabbro.

The undeformed peridotite body is bounded by serpentinite mylonite zones and there is a gradual transition from partially serpentinized peridotite tectonites to increasingly strained serpentinized peridotites and finally to serpentinite mylonites. The mylonitic foliation is defined by alternating antigorite + magnetite and chlorite + magnetite bands. Along the foliation relics of clinopyroxene porphyroclasts are replaced by neoblastic Alpine diopside, stable with the mylonitic assemblage. Progressive deformation is accompanied by preferential growth of a very fine-grained assemblage of olivine + Ti-clinohumite + magnetite + antigorite $\pm$ chlorite along shear bands cutting the mylonitic fabric. Formation of the olivine paragenesis is indicative of the Alpine peak temperature in presence of deformation.

Metre-scale boulders of metagabbro represent eclogitized varieties of the MORB-type intrusions emplacing in the Erro-Tobbio peridotites at the end of their mantle exhumation history. This gabbro is undeformed and preserves primary igneous ipidiomorphic texture, with interstitial pyroxene and idiomorphic plagioclase (Figure 26, right). In these rocks the igneous clinopyroxene is replaced by green Cr-omphacite, plagioclase is pseudomorphosed by fine-grained aggregates of jadeitic pyroxene, zoisite and garnet, igneous olivine is overgrown by aggregates of tremolite-talc at the core, chloritoid $\pm$ garnet at the rim. Detailed descriptions are in Messiga et al. (1995). Presence of these rocks (cropping out several hundred meters from this stop) indicates that the peak metamorphism in serpentinites and gabbros occurred at eclogite-facies conditions, estimated around 25 Kbar and about 550-600 °C (Messiga et al., 1995).

After this outcrop we walk for about 20 minutes along the riverbed, until a second outcrop showing olivine + Ti-clinohumite veins cutting the antigorite serpentintes. Detailed mapping and structural descriptions of these rocks are reported by Scambelluri et al. (1995) and Hermann et al. (2000), and are presented in section 6.4 and in Figure 13 of this guidebook. Hereafter some pictures of serpentinites and veins (Figure 27).

Figure 27 - Olivine + magnetite + Ti-clinohumite veins cutting antigorite serpentinites

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