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Shear zone structures and microstructures in mantle peridotites from the Voltri massif, Ligurian Alps, N.W. Italy

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Abstract

In the Erro-Tobbio peridotite three generations of shear zone structures are developed in granular peridotite: peridotite tectonites, peridotite mylonites and serpentinite mylonites. Peridotite tectonites occur in a kilometre-scale shear zone developed under estimated conditions of 1100 to 1220° C and 16 ± 6 kbar, and flow stresses between 4 and 11 MPa. The deformation in this shear zone involved dislocation creep and concurrent recrystallization of olivine by a combination of subgrain rotation and extensive grain boundary migration. With decreasing pressure and temperature and increasing flow stress, spinel peridotite mylonites (T = 800–925° C, P = 6–8 kbar, σ = 93–153 MPa) and chlorite peridotite mylonites (T = 550–800° C, P = 4–6 kbar, σ = 331–786 MPa) developed in up to hundred metre scale shear zones transecting the peridotite tectonites. The mylonitic microstructures suggest that dynamic recrystallization of olivine occurred by a mechanism involving grain boundary bulging to nucleate new strain free grains. In addition, 'fluidal' microstructures dominated by pyroxene clasts with tails of fine-grained amphiboles suggest that superplastic mechanisms may have been operative in these mylonites. The sequence of tectonite shear zones followed by peridotite mylonites is inferred to result from the progressive ascent of asthenospheric peridotites during opening of the Mesozoic Piemonte-Ligurian ocean.

Serpentinite mylonites, developed at temperatures of 300–550° C and pressures of about 4 kbar, are observed in shear zones mostly associated with ductile overthrusting and imbrication. These structures are clearly related to nappe emplacement during Alpine collision.

Introduction

Many aspects of tectonics in the lithosphere are controlled by the rheological behaviour of the peridotitic upper mantle (e.g. Goetze & Evans 1979; Kirby 1985). This observation has motivated research on the rheology of peridotites, involving both experimental deformation studies (e.g. Avé Lallemant & Carter 1970; Carter & Tsenn 1987; Chopra & Paterson 1981, 1984; Kirby 1985; Kirby & Kronenberg 1987) and studies of naturally deformed mantle peridotites (Avé Lallemant 1967, 1985; Boland et al. 1971; Boullier & Nicolas 1975; Den Tex 1969; Green & Radcliffe 1972; Gueguen 1977; Mercier & Nicolas 1975; Nicolas 1986a, b; Nicolas et al. 1971).

Two groups of upper mantle rocks are accessible for direct structural and petrofabric studies: (1)

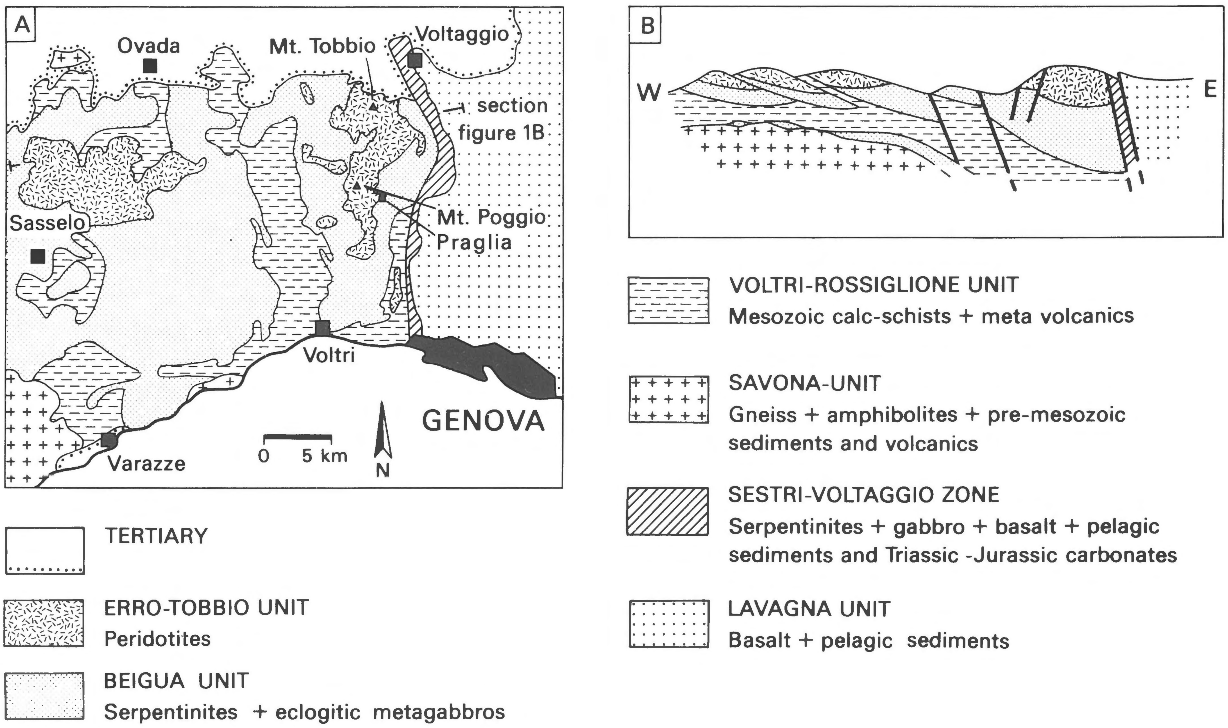

Fig. 1. Schematic tectonic map (A) and cross section (B) of the Voltri Group. Cross section from Chiesa et al. (1975). Approximate location of section (B) shown on map (A).

peridotite xenoliths in basaltic and kimberlitic magmas, and (2) huge peridotite slices included into the crust during orogenesis (Nicolas 1986a, b; Nicolas & Poirier 1976). Studies of these peridotites place constraints on the application of experimental rheological data to natural flow in the upper mantle, but confirm that similar types of mechanisms operate in naturally and experimentally deformed olivine rocks, i.e. dislocation creep and dynamic recrystallization (Kirby 1985; Carter & Tsenn 1987).

This paper documents the first results of a study concerned with the deformation structures and fabrics of an Alpine type peridotite (Den Tex 1969; Nicolas 1986b) from the Voltri Massif in the Ligurian Alps. The purpose of this study is to use field and microstructural information to unravel the deformation history, to determine the fundamental mechanisms operating during deformation and, where possible, to assess the deformation conditions.

Geological setting and metamorphic history

The Voltri Massif immediately NW of Genova occupies the easternmost part of the Ligurian Alps, and is separated from the Ligurian Apennines to the east by the Sestri-Voltaggio zone (Fig. 1). Previous work (Chiesa et al. 1975, Piccardo et al. 1977) has shown that the large-scale structure of the massif is dominated by a number of subhorizontal thrust sheets, which have been dismembered by later N-S trending faults (Fig. 1B). These thrust sheets compose the 'Voltri Group', and include the Voltri-Rossiglione unit of Mesozoic high-pressure (HP) metamorphic calcschists and metavolcanics, overlain by the Beigua unit of dominantly antigorite serpentinites with eclogitic metagabbros, in turn overlain by the Erro-Tobbio (ET) peridotites. The Voltri Group lies tectonically emplaced onto the Savona continental basement.

The metamorphic history of the Voltri Group rocks has been studied in detail (Chiesa et al. 1975; Cimmino et al. 1979; Ernst 1976; Messiga 1987;

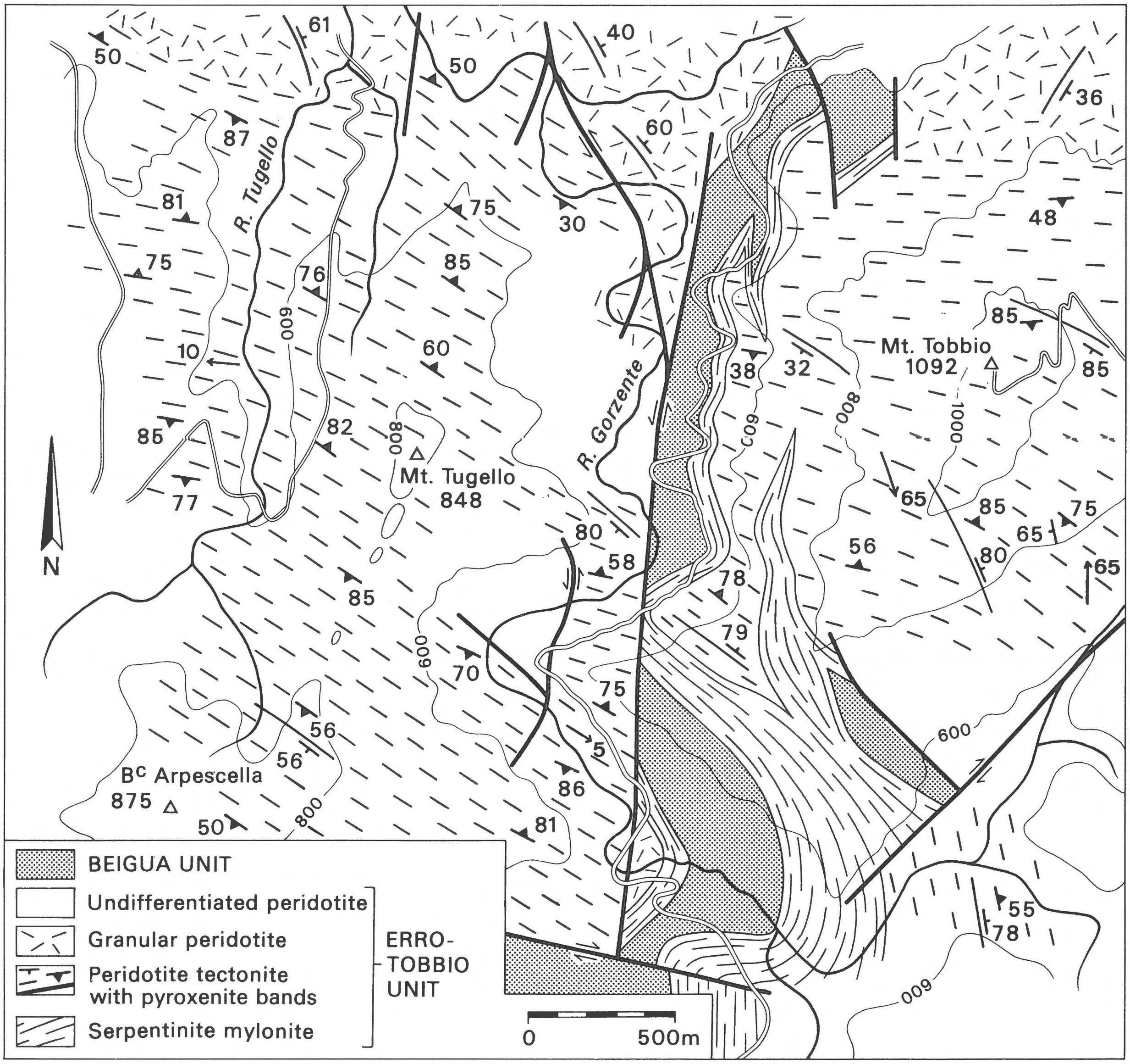

Fig. 2. Schematic geologic map of the Mt. Tobbio area.

Messiga & Piccardo 1974; Messiga et al. 1983). These studies clearly demonstrate that the ET peridotites have a different history than the underlying units. The inferred pressure-temperature (PT) path for the peridotites goes retrogressively from high PT conditions in the upper mantle to lower PT conditions at lower-crustal levels (Ernst & Piccardo 1979). In contrast, the rocks of the underlying thrust sheets show a prograde path to peak high pressure – low temperature (HP-LT) metamorphism, followed by uplift along a LT trajectory (Messiga 1987; Messiga et al. 1983; Piccardo et al. 1977). On the basis of these different histories the Beigua and Voltri-Rossiglione units are thought to represent subducted and dismembered oceanic crust of the Piemonte-Ligurian basin, whereas the ET peridotites are interpreted as mantle material from the hanging wall of the subduction zone (Piccardo et al. 1977).

The petrology of the ET peridotites has been

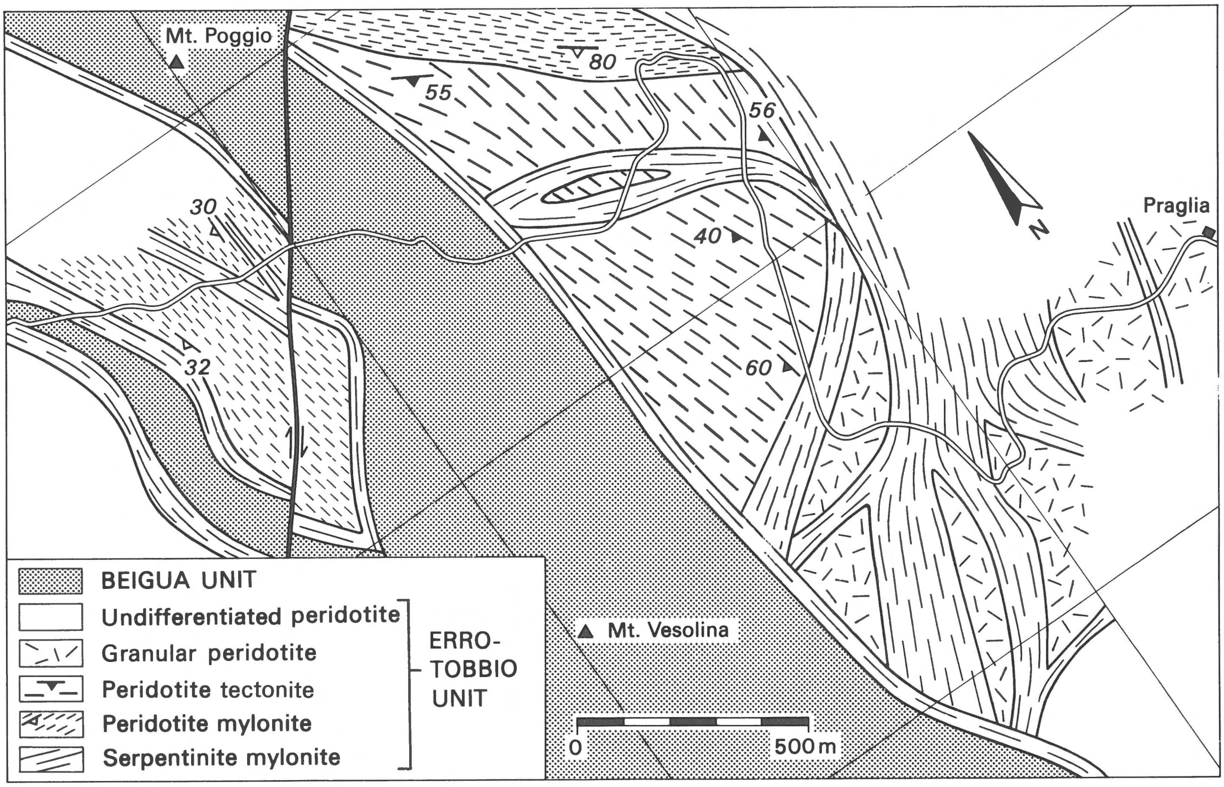

Fig. 3. Schematic tectonic map of the Praglia area.

described in detail by Bezzi & Piccardo (1971) and Ernst & Piccardo (1979). The dominant rock types include spinel lherzolites and plagioclase-spinel lherzolites, with subordinate layers of dunite, pyroxenite and gabbro. Nicolas (1984, 1986a, b) has classified the ET peridotites in his plagioclase lherzolite group similar to the Lanzo and Cap de Corse Massifs.

The ET peridotites are exposed in two main areas in the Voltri Massif (Fig. 1). Our study is restricted to the eastern area and concentrates around Mt. Tobbio and Praglia.

Geological structure around Mt. Tobbio and Praglia

The structural maps of the Mt. Tobbio and Praglia areas are shown in Figs. 2 and 3. In the Mt. Tobbio area (Fig. 2) the primary contact between the ET and Beigua units is strongly dissected by dextral and sinistral oblique-slip faults. In the Praglia area (Fig. 3), the ET and Beigua units are imbricated to form a number of approximately NS trending thrust slices. The western contacts of the Beigua slices are sub-vertical and are defined by serpentinite mylonite zones, whereas the eastern contacts, equally decorated with serpentinite mylonites, represent the reactivated and dissected primary contact between the ET and Beigua units.

Structures in the ET peridotites

Four generations of deformational structures have been identified in the ET peridotites, and the intensity of the deformation associated with these structures varies considerably. In domains of virtually no deformation or very low strain the peridotites only contain primary structures. These structures include a compositional layering (S_o) which is demonstrably overprinted by a tectonite foliation

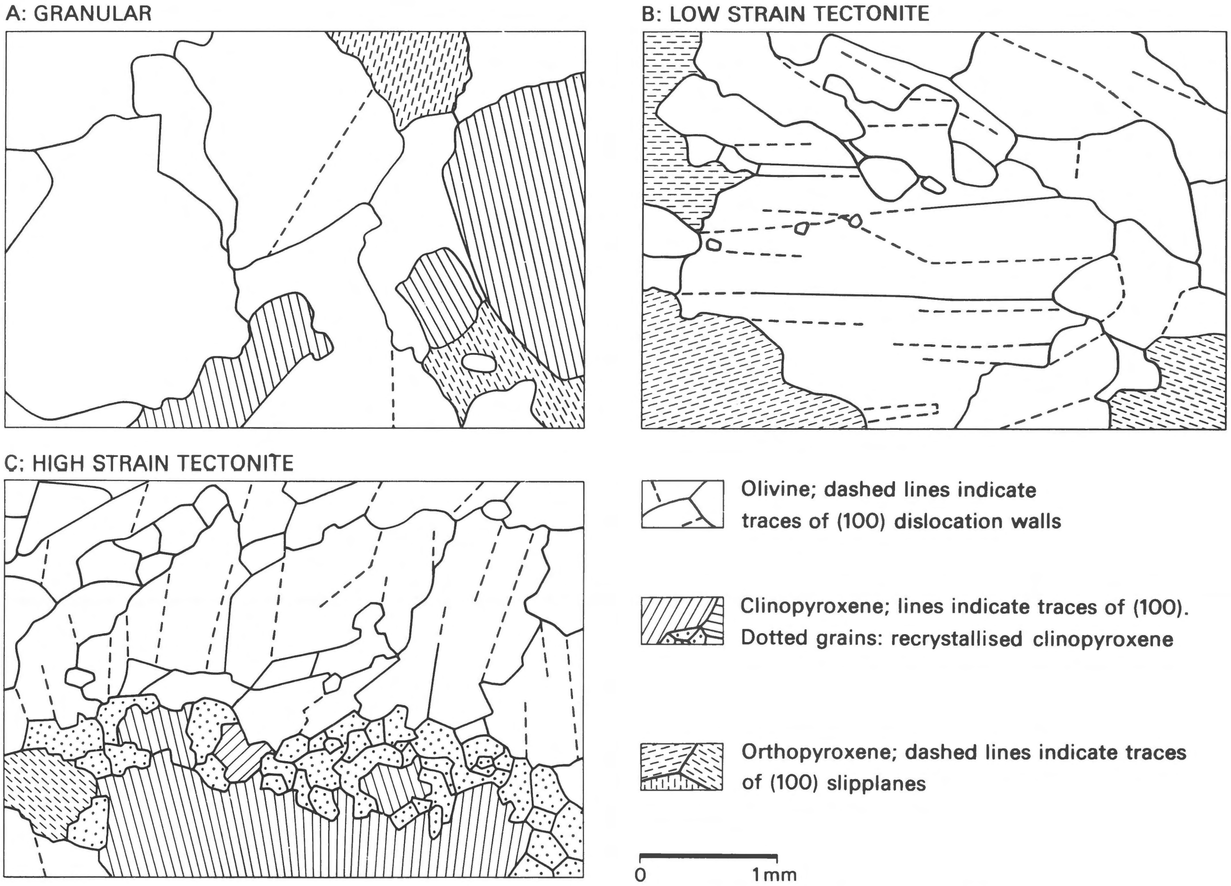

Fig. 4. Line drawings from thin-sections of granular (A), low-strain (B) and high-strain peridotite tectonite (C).

(S_t) developed over a large part of the areas studied. The tectonite foliation is deformed in shear zones, related to two stages of mylonitic deformation and leading to the development of peridotite mylonites and serpentinite mylonites. All ductile deformation structures are affected by younger cataclastic fault zones.

Primary structures and microstructures

Within kilometre-scale regions of low strain no foliations are developed, and the peridotites in those areas have a homogeneous and coarse-grained (0.5–1 cm) granular texture. Primary structures are well preserved in these rocks and include a centimetre-scale compositional banding, 'indigenous' centimetre- to metre-scale layers of dunite and pyroxenite, and 'intrusive' meta-gabbroic dikes similar to those described from other peridotite massifs (Nicolas 1984, 1986a, b; Wilshire 1987). The pyroxenite and dunite layers are commonly sub-parallel and delineate a uniform to smoothly curved pattern on the kilometre scale (Fig. 2). However, examples of dunite layers cross-cutting pyroxenites occur in a few locations.

Granular peridotites are coarse-grained, and all mineral phases have a similar grainsize (Fig. 4A). Orthopyroxene, spinel and clinopyroxene occur in characteristic clusters as described by Mercier & Nicolas (1975) and Nicolas (1986a, b). Surprisingly, plagioclase-bearing assemblages described from the Voltri massif by Bezzi & Piccardo (1971) and Ernst & Piccardo (1979) have not been found in any of the peridotites in the areas studied. 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. The olivine and pyroxene grains show minor internal deformation characterized by very open kinks, and most grains show a weak undulatory extinction (Fig. 4A).

Peridotite tectonites

The bulk of the ET peridotites studied have a well-developed tectonite foliation (S_t) defined by the shape-preferred orientation of deformed pyroxene, spinel and olivine grains. In the areas studied (Figs. 2, 3) this foliation has a relatively constant orientation on a kilometre-scale, and strikes between E–W and SE–NW with steep to moderate dips in the Mt. Tobbio area (Fig. 2), and about SSE–NNW with moderate dips in the Praglia area (Fig. 3). Stretching lineations in the plane of the tectonite foliation are common, but their orientations may vary. In the Mt. Tobbio area (Fig. 2), for example, this lineation is subhorizontal in the River Gorzente section but subvertical on Mt. Tobbio to the east.

In the River Gorzente section, and in two sections north of Mt. Tobbio and north of Mt. Tugello (Fig. 2), 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. 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 primary 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 (Fig. 2). Note that the entire structure is disrupted by later serpentinite mylonites and cataclastic fault zones affecting to an unknown extent the large-scale geometry of the early structures. However, 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 sections indicate, that this transition between granular and tectonite peridotites does not result from juxaposition along a later fault or serpentinite mylonite zone.

Locally, within the tectonites, metre-scale folds occur of pyroxenite layers. S_t is commonly axial planar to these folds but in a few cases refolded folds indicate the existence of more than one fold generation within the tectonite deformation zones.

The microstructure of moderately foliated, presumably low-strain tectonites is dominated by large olivine grains showing well-developed deformation substructures of sharply defined, high angle kink-like subgrains with undulatory extinction (Fig. 4B). Large olivine grains are not preserved in tectonites with a more intensely developed foliation. Instead, the microstructure is characterized by smaller (0.5–1.5 mm) elongate to tabular shaped grains (Fig. 4C) 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 from grain size reduction during dynamic recrystallization. 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.

The orthopyroxene grains in the tectonites have similar sizes as those in the granular peridotites. Undulatory extinction is common and kink bands are well developed but there is no evidence of recrystallization. In contrast, clinopyroxene occurs as highly deformed clasts of original grains, recrystallizing to an equiaxed smaller grainsize (0.1–0.5 mm, Fig. 4C).

In summary, the tectonite microstructures in the ET peridotites range from porphyroclastic to tabular microstructures (Mercier & Nicolas 1975; Harte 1977). These microstructures are similar to

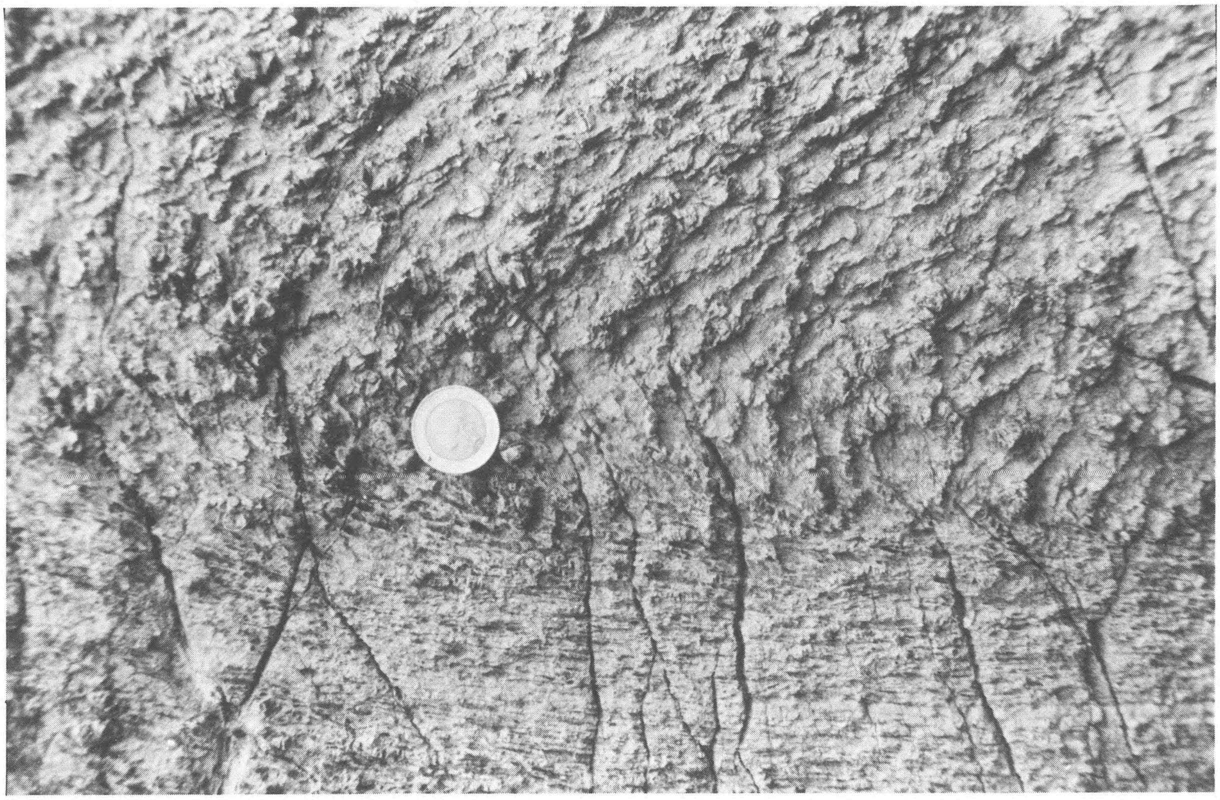

Fig. 5. Detail of a peridotite mylonite in the Praglia area. Note the progressive bending of the tectonite foliation (S_t) into the mylonite zone and the pervasive shear foliation in the mylonite. Diametre coin 2.5 cm.

those developed in the core of the Lanzo massif (Nicolas & Poirier 1976) and some basalt xenoliths (Avé Lallemant 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 strain suggesting that recrystallization involved an important component of subgrain rotation (Poirier & Nicolas 1975). The occurrence of the oblique olivine grain shape foliation suggests that grain boundary migration was also extensive (Lister & Snoke 1984).

Peridotite mylonite zones

In the area around Praglia and Mt. Poggio (Fig. 3) the tectonite foliation is overprinted by a mylonitic fabric developed in peridotite mylonite zones. The extensive development in this area of younger anastomosing serpentinite mylonites described below renders the study of these structures somewhat difficult. Peridotite mylonite structures, however, can be studied within augen-shaped domains bounded by the serpentinite mylonite zones. A detailed study has been made of the structures in the peridotite domains exposed in the road section from Mt. Poggio to Praglia. These domains contain granular peridotite near Praglia, peridotite tectonite structures NE of Mt. Vesolina and peridotite mylonites E and S of Mt. Poggio (Fig. 3). In the peridotites NE of Mt. Vesolina peridotite mylonites occur within distinct decimetre to metre-scale shear zones (Fig. 5) in which, towards the centre of the shear zones, S_t progressively rotates into parallelism with the shear zone boundaries. A wider belt, several hundreds of metres thick, of peridotite mylonites occurs on the slopes of Mt. Poggio. The mylonitic foliation in this belt anastomoses around local metre-scale lenses of less deformed peridotites with a tectonite foliation preserved. The peri-

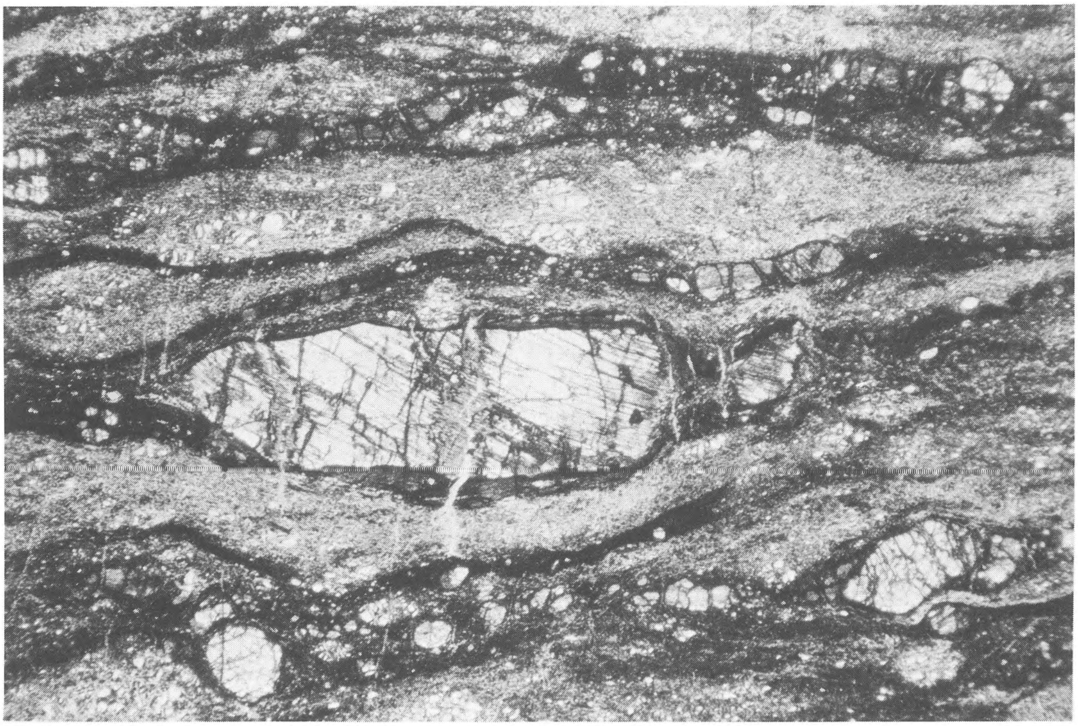

Fig. 6. Photomicrograph of a peridotite mylonite. Width of micrograph 15 mm.

dotite mylonite zones strike approximately NW–SE with dips to the SW, and asymmetric porphyroclast systems of intensely deformed pyroxenes and a distinct stretching lineation defined by these pyroxenes indicate a S to SE sense of movement across these zones.

Both tectonite and mylonite structures in the peridotites are cut by subvertical fine-grained gabbroic dikes. Optical microscopy shows that these dikes have preserved their igneous microstructure in most cases, but that intense greenschist facies metamorphism has led to replacement of a primary assemblage by zoisite + chlorite + garnet. A well-developed deformation microstructure only occurs in dikes that have been affected by serpentinite mylonite zones.

The mylonitic microstructures are characterized by intense grainsize reduction of olivine from a 500–1500 μm initial grainsize to a 10–150 μm recrystallized grainsize (Fig. 6). The old olivine grains show undulatory extinction and contain poorly defined sub-boundaries. There is a remarkably abrupt change in lattice orientation from old grains to new grains instead of the progressive misorientation from subgrains to recrystallized grains often documented from olivine and quartz rich mylonites (see e.g. White 1976). This suggests that other mechanisms than purely rotation recrystallization (Poirier & Guillopé 1979; Drury et al. 1985) have been operative. This is supported by the analogy of the microstructures in the ET mylonites with those produced experimentally where recrystallization has involved local grain boundary bulging (Ross et al. 1980).

The other mineral phases commonly occur as rounded clasts in a matrix of recrystallized olivine (Fig. 6). The small size of these porphyroclasts suggests that original large grains have been fractured and microboudinaged during mylonitic deformation (cf. Boullier 1980). Two types of peridotite mylonite can be distinguished on the basis of the stable mineral assemblage during deformation, i.e. spinel peridotite mylonites and chlorite peridotite mylonites. In the spinel-bearing mylonites py-

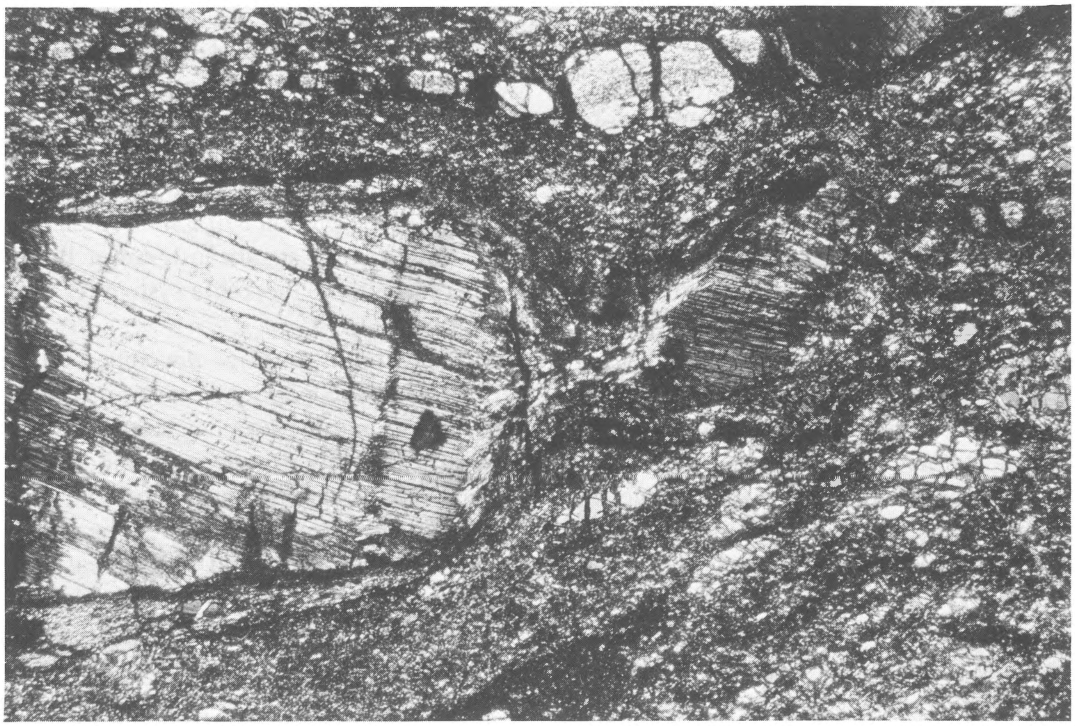

Fig. 7. Photomicrograph showing detail of the peridotite mylonite in Figure 6. Note coarse-grained amphibole fibers in the pressure shadow of an extended orthopyroxene grain. The fine-grained material mantling the orthopyroxene consists of amphibole, the matrix of the mylonite is made up of fine-grained recrystallized olivine. Width of micrograph 5 mm.

roxene clasts have fine-grained 'tails' up to a few centimetres long, defining a 'fluidal' microstructure identical to that in some kimberlite xenoliths (Boullier & Nicolas 1975) and in quartz feldspar mylonites (Passchier & Simpson 1986) (Fig. 6, 7). The material in the tails has a variable microstructure but is commonly very fine grained (d = 10–70 μm), in contrast with the distinctly coarser-grained microstructure in pressure shadows adjacent to elongate pyroxene augen (Fig. 7). Optical and microprobe analysis shows that both tails and pressure shadows consist of edenite (amphibole nomenclature after Leake 1978) which indicates that hydration accompanied deformation.

In the chlorite-bearing peridotite mylonites the spinel is completely replaced by Mg-chlorite + magnetite aggregates. The recrystallized olivine grain size is smaller in the chlorite-bearing peridotite mylonites (e.g. 10–30 μm) than in the spinel peridotite mylonites (e.g. 80–150 μm).

Serpentinite mylonite zones

In the serpentinite mylonites all primary phases have been transformed into antigorite, iron oxides (magnetite, ilmenite) and Mg-chlorite, and intensely deformed serpentinite mylonites commonly show a strongly foliated microstructure dominated by a millimetre-scale alternation of antigorite, antigorite-chlorite, and chlorite-magnetite bands. Very fine-grained relics of fractured olivine and pyroxene are largely restricted to the chlorite-magnetite bands. These olivine and pyroxene grains often occur in sigma-type porphyroclast systems (Passchier & Simpson 1986) with antigorite and Mg-chlorite in pressure shadows adjacent to the porphyroclasts. The serpentinite mylonites commonly show single and locally multiple sets of extensional crenulation cleavages or shear bands (Platt & Vissers 1980). Within these shear bands antigorite fibres recrystallize into a finer-grained aggregate of equant grains.

Pressure-temperature evolution and conditions during deformation

The mineral assemblages and mineral chemistry of deformed peridotites allow to place some constraints on the conditions during deformation, and on the pressure-temperature evolution of the rocks with time. The occurrence of clusters of clinopyroxene, orthopyroxene and spinel suggests that the peridotites originate from depressurized garnet lherzolites (Green & Burnley 1988; Nicolas 1986a, b, see Fig. 8). The fertile composition of the ET peridotites indicates limited partial melting during ascent from the asthenosphere (Ernst & Picardo 1979; Nicolas 1986a, b). The mineral assemblage in the tectonites is the same as that in the granular peridotites i.e. olivine + orthopyroxene + clinopyroxene + spinel, indicating that tectonite deformation occurred in the spinel lherzolite field (Fig. 8). On the basis of the mineral chemistry of the granular peridotites and peridotite tectonites, Ernst & Piccardo (1979) have inferred temperatures of 1150 ± 50° C and confining pressures of 16 ± 6 kbar. The preservation of the spinel assemblage in the granular peridotites and peridotite tectonites indicate that the peridotite stopped rising along a steep geothermal path at depth between 35 and 70 kilometres, outside the plagioclase lherzolite field (Fig. 8). However, the local occurrence of plagioclase rims around spinel (Bezzi & Piccardo 1971; Ernst & Piccardo 1979) indicates that the ET peridotite did recrystallize close to and locally in the plagioclase stability field, hence that part of the PT trajectory of the peridotites is constrained by the boundary of that field (Fig. 8). Further constraints on the early PT path during uplift are limited. Petrological and geochemical evidence (Ernst & Piccardo 1979; Ottonello et al. 1979) suggest that partial melting of the ET peridotites has occurred in the spinel stability field, but it is not clear if this was in the Mesozoic or earlier. Isotopic studies (Polvé & Allègre 1980; Hamelin & Allègre 1988) indicate that many western Mediterranean orogenic peridotites underwent partial melting sometime during the Proterozoic. If this is also the case for the ET peridotites the Mesozoic PT uplift path need not cross the dry solidus and a number of early PT paths are permissable (Fig. 8). It is emphasized here that this range of possible PT paths includes those consistent with thermal models of pure as well as simple shear extensional geometries (Ruppel et al. 1988). Even if the PT path did not cross the dry solidus, limited partial melting may have occurred in the presence of minor amounts of hydrous fluids which significantly lower the solidus temperature in the lherzolite system (cf. Piccardo et al. 1988).

The mineral assemblage developed during deformation in the peridotite mylonite zones provides constraints on prevailing temperatures as follows. The stability of calcic amphiboles strongly depends on the amphibole composition, and shifts to higher temperatures with increasing ratio of pargasite versus tremolite, and with increasing Ti and Al content (Jenkins 1983; Oba 1980). However, an increasing Fe^{2+} content strongly decreases the maximum thermal stability (Ernst 1968). The composition of the clearly syntectonic edenite (Pargasite$_{45-46}$–Edenite$_{25-27}$–Tremolite$_{27-30}$; TiO_2 < 0.5 wt%, FeO 3.5–4.1 wt%) in the spinel-bearing peridotite mylonites thus indicates that temperatures were less than 925° C (Fig. 8). A lower temperature limit of 750 to 800° C for the spinel-bearing mylonites is indicated by the stability of the assemblage spinel + orthopyroxene + olivine + H_2O which at lower temperatures reacts to form chlorite (Fig. 8). It follows that the chlorite-bearing peridotite mylonites developed at lower temperatures (550–800° C) than the spinel-bearing mylonites. A similar temperature range of 600° to 900° C has been inferred for mylonites in other peridotite massifs (reviewed in Nicolas 1986a).

From the above observations it follows that the ET peridotites crossed the spinel-calcic amphibole stability field close to the spinel-plagioclase transition. This results in a PT path following a steep thermal gradient (about 40° C/km, Fig. 8).

In addition to the above estimates of pressure and temperature, paleopiezometer relationships for subgrains and rotation-recrystallized grains in olivine (Karato et al. 1980; Ross et al. 1980) may provide an estimate of the flow stress during deformation. Such stress estimates can potentially be used to constrain the deformation temperature us-

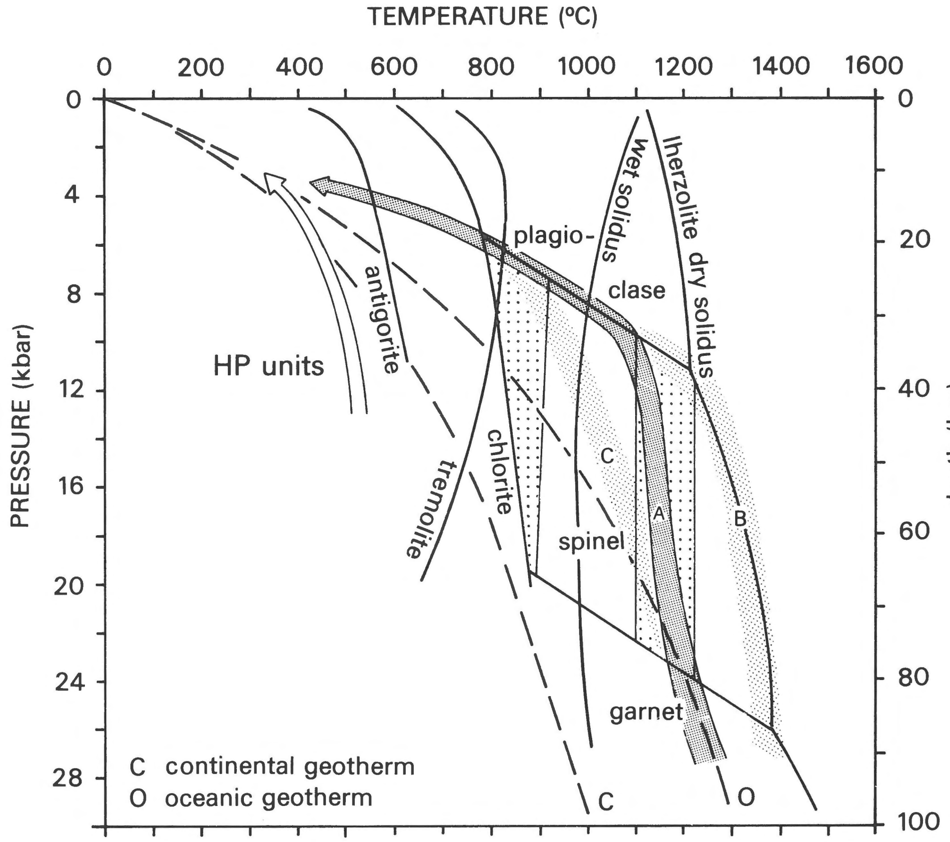

Fig. 8. Possible P-T paths for the ET peridotites in a phase diagram for the reactions in the system CaO-MgO-Al_2O_3-SiO_2-H_2O below 700° C (after Evans 1977) and in the system CaO-MgO-Al_2O_3-SiO_2-Na_2O-H_2O above 700° C (after Jenkins 1983). Lherzolite wet solidus after Wyllie (1979) and lherzolite dry solidus after Takahashi & Kushiro (1983). The geotherms are from Mercier (1980). The stability fields for the peridotite tectonite (1100–1220° C) and for the assemblage spinel + edenite (800–925° C) are dotted. The P-T path with the best fit to the presently available data is indicated with A. The other paths are consistent with thermal models calculated for either symmetric pure shear extension (B) or asymmetric simple shear extension (C; cf. Ruppel et al. 1988).

ing a dislocation creep flow law for olivine (Chopra & Paterson 1981, 1984) for a range of geologically reasonable strainrates (chosen 10^{-12} to $10^{-15}\,s^{-1}$ in this paper). Substitution of a maximum value for the strainrate and a minimum value for the flow stress in a flow law for dry dunite thus leads to an estimate of the maximum value for the temperature during flow, while the minimum strainrate and maximum stress in a flow law for wet dunite provides a minimum temperature. The values of the flow stresses and temperatures, inferred with this method for the peridotite tectonites and peridotite mylonites, are listed in Table 1 and these estimates are in good agreement with the temperatures inferred from the mineral assemblages and mineral chemistry. Both methods indicate that the mylonitic deformation occurred at lower temperatures than the tectonite deformation.

The transitional relationships between some peridotite and serpentinite mylonites suggests that the mylonite zones were active during cooling of the ET peridotite down to temperatures as low as 500° C (Fig. 8). Synchronously with cooling the mylonites were progressively infiltrated by water-rich fluids. Antigorite remained stable during deformation indicating that temperatures did not decrease below 300° C (Evans 1977).

Deformation history and tectonic evolution

The PT path of the ET peridotites obtained above is characteristic for the uplift of asthenospheric peridotites in an extensional setting (Nicolas 1986b). The early history of the peridotites involved depressurization and possibly minor melt production. The development of the tectonite structures is related to this early-stage uplift. Initial uplift along a steep geothermal path was followed by further uplift and synchronous cooling and involved the progressive development of spinel and chlorite-bearing peridotite mylonites and eventually serpentinite mylonites.

A notable consequence of the inferred PT path is, that thrusting of the ET peridotite over the HP-metamorphic subduction complex (e.g. Beigua and Voltri-Rossiglione units) must have occurred under greenschist facies conditions, i.e. late in the Alpine orogenic history of the Voltri Massif. This conclusion is consistent with petrologic and structural studies by other workers in the Voltri Massif (e.g. Capponi et al. 1986; Chiesa et al. 1975; Piccardo et al. 1977).

The deformational and metamorphic histories of the units in the Voltri Group (Piccardo et al. 1977; Ernst & Piccardo 1979) allow a preliminary model to be developed relating the evolution of the ET peridotites to the larger scale tectonic history of the region. We may envisage that the structural and metamorphic history of the ET peridotites is related to the uplift of hot asthenosphere during the rifting and opening stages of the Piemonte-Ligurian ocean. This tectonic setting is supported by the composition of the spinel ($Cr/Al + Cr > 0.1$) and clinopyroxene ($Na/Al + Na < 0.16$), which have chemical characteristics similar to those from transitional to subcontinental peridotites (Ishiwatari, 1985, chemical data from Ernst & Piccardo, 1979). The development of the tectonite foliation in this setting can be ascribed either to the high-strain margin of an asthenosphere diapir (e.g. Nicolas 1986b) or, alternatively, the tectonites represent upper mantle ductile shear zones related to either symmetric or asymmetric extension of the lithosphere. Additional constraints are required to discriminate between one of these options. However, the extension-related petrologic and structural evolution of the ophiolite complexes in the northern Apennines (Cortesogno & Luchetti 1982, 1984; Cortesogno et al. 1987; Hoogerduijn Strating, in press) suggests that the early opening of the Piemonte-Ligurian ocean involved asymmetric crustal extension (Fig. 9; Lemoine et al. 1987).

The peridotite mylonite zones were formed during ongoing extension under decreasing conditions of pressure and temperature at gradually shallower levels. The intrusion of gabbroic dikes transecting the peridotite mylonites suggests that the mylonites developed while, during progressive extension, the peridotites became transported towards a position above a melt-producing, ascending asthenosphere (Fig. 9). It may be noted here that the syntectonic hydration in the mylonite zones, implied by the development of edenitic amphiboles, may have had an important effect on the bulk rheology of the lithosphere. The mylonitic microstructures are similar to those in fluidal kimberlite xenoliths (Boullier & Gueguen 1975; Boullier & Nicolas 1975) which are thought to have developed by 'superplastic' flow in finegrained orthopyroxene. It follows, that if a similar mechanism has operated in the ET mylonites, the rheology of the mylonite zones would have been controlled by 'superplastic' amphibole. In this respect it is important to note

Table 1. Calculated flow stresses and temperatures

Deformation fabric	Grain size (μm)	Flow stress (MPa)	Temperature (°C)
Peridotite tectonite	1500	4.3	1219
	500	10.8	896
Spinel peridotite	150	93.3	914
mylonite	80	153.0	676
Chlorite peridotite	30	331.1	822
mylonite	10	786.0	577

Paleopiezometer for recrystallised grains:
Peridotite tectonite – Karato et al. (1980)
Peridotite mylonites – Ross et al. (1980)
Flow laws:
Dry Aheim dunite (Chopra & Paterson 1984)
Wet Aheim dunite (Chopra & Paterson 1984)
Strain rates:
Maximum: $10^{-12}sec^{-1}$
Minimum: $10^{-15}sec^{-1}$

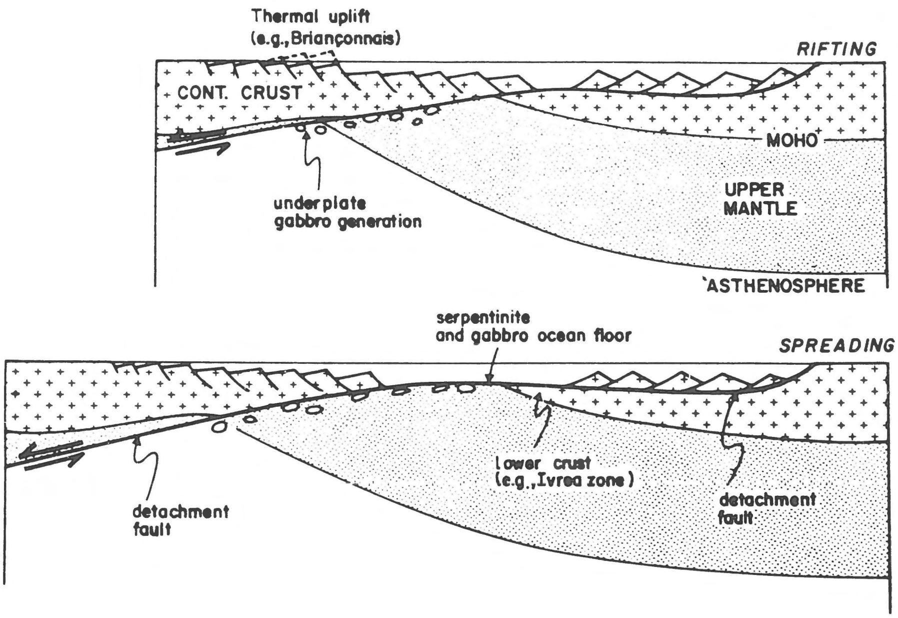

Fig. 9. Asymmetric extensional model for tectonic exhumation of upper mantle peridotites as inferred for the Piemonte-Ligurian ocean (from Lemoine et al. 1987). The P-T history and mylonitic deformation of the ET peridotites are thought to be related to progressive uplift along a low-angle extensional detatchment fault.

that the very large strains reflected by the amphibole tails are inconsistent with the high strength of amphiboles during dislocation creep (Hacker & Christie, 1987; Nicolas & Poirier 1976), but that they are certainly consistent with the operation of a superplastic mechanism.

In some peridotite mylonite zones transitional relationships occur between peridotite and serpentinite mylonites which possibly reflect continued movement along these mylonite zones during extension and cooling of the lithosphere below 550° C. However, most serpentinite mylonites cut the peridotite mylonite zones (Fig. 3) and these structures are clearly associated with imbrication of the ET unit with the ophiolitic units of the Voltri Group during Alpine collision.

Recent deep seismics in extensional terrains (i.e. BIRPS, COCORP, etc.) reveal ubiquitous horizontal and dipping reflectors underneath the seismic Moho, indicating that the upper mantle is far from homogeneous. There is an increasing belief that these reflectors represent tectonic features (i.e. faults, shearzones) possibly associated with compositional variations (Warner & McGeary 1987; Klemperer 1988). We suggest that the shear zone structures in the ET peridotites represent ancient equivalents of these features and that similar structures as those in the ET peridotites may act as upper mantle seismic reflectors.

Conclusions

1. Three generations of ductile deformation structures developed in initially granular peridotites, i.e. peridotite tectonites, peridotite mylonites, and serpentinite mylonites. Inferred PT conditions during the development of these structures are as follows:

Peridotite tectonites	1100–1220° C	16±6 kbar
Spinel peridotite mylonites	800– 925° C	6–8 kbar
Chlorite peridotite mylonites	550– 800° C	4–6 kbar
Serpentinite mylonites	300– 550° C	±4 kbar

2. The PT path of the peridotite is characteristic for the uplift of asthenospheric peridotites in an extensional setting.
3. The development of peridotite tectonites involved kilometre-scale shear localisation in up-

per mantle granular peridotites. An oblique fabric similar to that in certain S-C mylonites in quartzites (Lister & Snoke 1984) also occurs in peridotite tectonites.

4. Syntectonic hydration of peridotite mylonite zones may have had a drastic effect on the bulk-rheology of the extending lithosphere. 'Fluidal' microstructures (Boullier & Gueguen 1975; Boullier & Nicolas 1975) in these mylonites involving intense deformation of very fine-grained amphiboles suggest a superplastic mechanism.

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