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Structural and metamorphic signature of alpine tectonics in the Voltri Massif (Ligurian Alps, North-Western Italy)

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Key words: Deformation, structural analysis, Voltri Massif, Ligurian Alps, Western Alps, North-Western Italy

ABSTRACT

The Ligurian-Piemontese units, widely exposed at the south-eastern end of the Western Alps, display a complex superposition of deformational structures. Despite a very heterogeneous strain distribution, a detailed structural analysis of the whole area allows us to separate the structures into groups that can be considered the signature of four main tectonic phases of the Alpine orogeny. Disrupted folds and foliations in relict eclogitic domains can be related to subduction events; a group of structures evolving from superposed isoclinal folds to shear bands and to ductile and brittle-ductile reverse shear zones can be linked to uplift, exhumation and nappe emplacement. These structures are coeval with metamorphic conditions evolving from Na-amphibole greenschist, to greenschist s.s. to low greenschist facies. During this stage, strain distribution also suggests that a transpressional regime was active along the eastern sector of the Voltri Massif near the boundary with the Sestri-Voltaggio Zone. Open folds with sub-horizontal axial planes and ductile-brittle normal shear zones, record the post-uplift tectonic collapse, under low greenschist metamorphic conditions; E-NE vergent early Miocene thrusts and large scale folds are the expression of a late orogenic backthrusting. On the basis of the structural analysis we have determined a primary role for transpressional tectonics during greenschist facies metamorphism, in the tectonic setting of the sector close to the Sestri-Voltaggio Zone. Comparison with other metaophiolite complexes of the Alps, namely the Monviso zone, highlights analogies in the structural and metamorphic history, but also reveals some peculiar features linked to the tectonic position of the Voltri Massif in the Alpine belt.

RIASSUNTO

In corrispondenza della terminazione sud-orientale delle Alpi Occidentali, le unità Liguri-Piemontesi mostrano una complessa sovrapposizione di strutture deformative. Nonostante una distribuzione dello *strain* molto eterogenea, l'analisi strutturale condotta su tutta l'area ci ha permesso di raggruppare le strutture in quattro gruppi principali, che possono essere considerate il risultato di quattro fasi tettoniche dell'orogenesi Alpina. Una fase di subduzione è testimoniata da pieghe e foliazioni prive di continuità laterale, conservate in domini in cui sono ancora preservate paragenesi eclogitiche. Una successiva fase di sollevamento, denudamento e messa in posto delle falde è testimoniata da strutture che vanno da pieghe isoclinali, a strutture che esprimono estensione parallela alla foliazione, a zone di taglio inverse, duttili e duttili-fragili. Queste strutture si sono prodotte in condizioni metamorfiche che evolvono da Scisti Verdi a Na-anfibolo, a Scisti Verdi s.s., a Scisti Verdi di basso grado. La distribuzione dello strain suggerisce anche che, durante questo stadio, sia stato attivo un regime transpressivo lungo il confine con la Zona Sestri – Voltaggio. Dopo il sollevamento, subentra una fase di collasso tettonico, espressa da pieghe a piano assiale sub-orizzontale e da zone di taglio duttili-fragili a geometria normale, formatesi in condizioni metamorfiche in facies Scisti Verdi di basso grado. Infine una fase tettonica del Miocene Inferiore è testimoniata da pieghe e sovrascorrimenti con vergenza verso E-NE. Questa analisi indica che la tettonica transpressiva sin-metamorfismo a Scisti Verdi ha avuto un ruolo importante nella realizzazione dell'assetto strutturale del settore a contatto con la Zona Sestri – Voltaggio. Il confronto con altri complessi metaofiolitici delle Alpi, in particolar modo con la zona del Monviso, mette in luce analogie nell'evoluzione strutturale e metamorfica, ma anche alcune caratteristiche peculiari legate alla posizione tettonica del Massiccio di Voltri nella catena alpina.

1. Introduction

The SE sector of the Western Alps is commonly known as the Ligurian Alps (Fig. 1). Due to Oligocene-Miocene anticlockwise rotation, the Ligurian Alps have a WNW-ESE structural trend (Giglia et al. 1996), strongly diverging from the orientation of the Western Alps. In central Liguria, at the eastern edge of the Ligurian Alps, the Ligurian-Piemontese units (Vanossi et al. 1984) are widely exposed. They consist of metaophiolitic rocks associated with metasediments; in the regional literature they are often referred to as the Voltri Group (Chiesa et al. 1975) or the Voltri Massif.

The Ligurian-Piemontese units known as the Voltri Massif were involved in the Alpine orogenic events and experienced a

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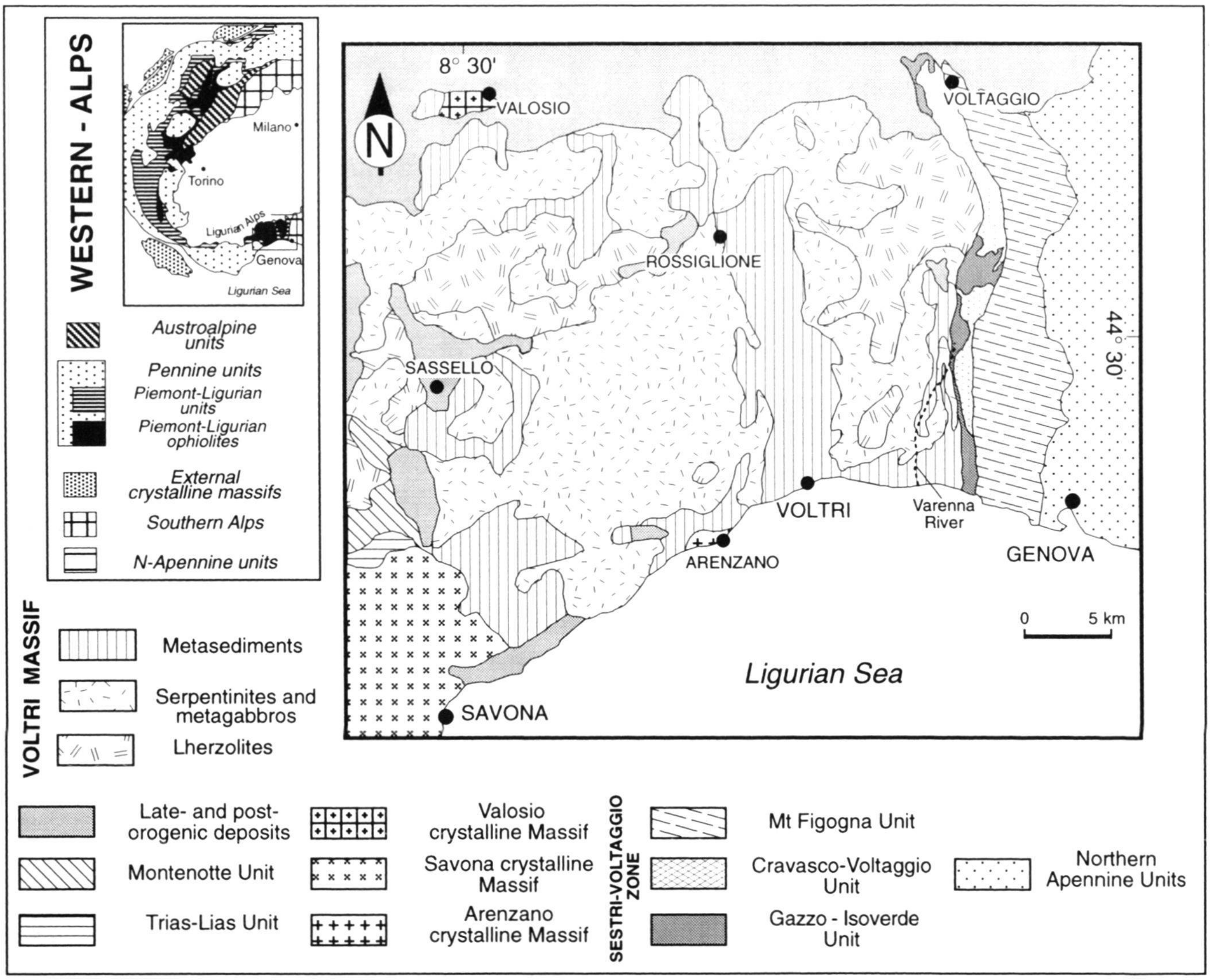

Fig. 1. Structural sketch map of the Voltri Massif and surrounding units. The inset shows the location of the Ligurian Alps in the framework of the Western Alps.

complex metamorphic and structural evolution, from subduction-related events to uplift and exhumation. As a result, these units show several superimposed deformations, evolving from a syn-metamorphic, ductile regime, to a brittle-ductile to brittle regime (Capponi 1987, Capponi et al. 1999a, b).

Both metaophiolitic rocks and metasediments of the Voltri Massif show that they have undergone the same structural and metamorphic evolution; nevertheless, strain distribution has been influenced by lithological contrast and the result is that deformations are best recorded in the metasediments (Capponi 1987), possibly also due to their layered structure.

The aim of this paper is to describe the structures of the Voltri Massif, to examine their role in the present-day structural arrangement and to discuss their structural evolution in the framework of the tectonic history of the Ligurian Alps.

2. Geological setting

The Voltri Massif metaophiolitic rocks (Fig. 1) are serpentinites with metagabbros and eclogitic lenses, and peridotites (lherzolites with minor pyroxenite and dunite bodies). The metasediments are interlayered with metabasite in places and are characterized by a variable percentage of white mica, calcite and quartz, marking the gradation from calcschists to micaschists and quartzschists. The metasediments belong to the "Schistes Lustrés" complex, and include both the sedimentary cover of the Ligurian – Piemontese ophiolitic basement, and slices of the cover of the thinned European continental crust. Serpentinites, metabasites and metasediments belong to the oceanic domain, formed between Paleo-Europe and the Apulian microplate during the Jurassic (Ligurian-Piemontese Domain, according to Vanossi et al. 1984). The lherzolites are in-

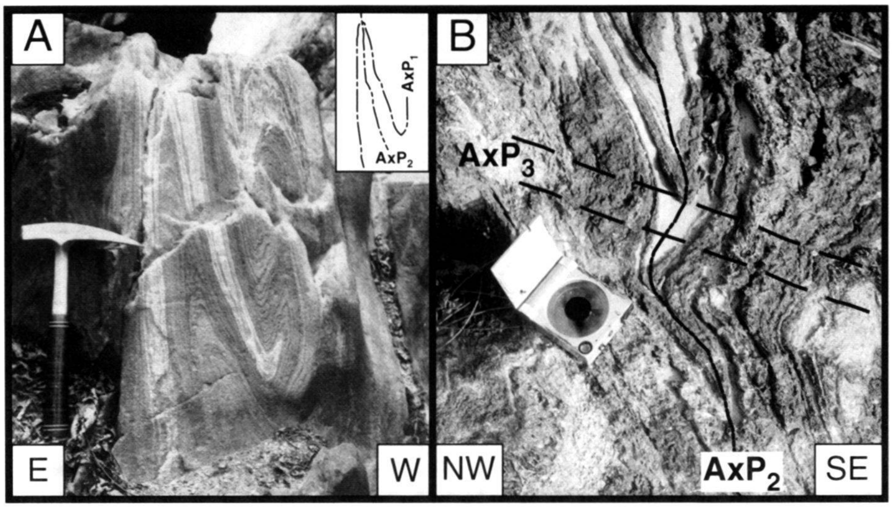

Fig. 2. Type 3 interference pattern in metasediments of the Voltri Massif (south-eastern sector).
A. D_2 overprinting on D_1 fold.
B. D_3 overprinting on D_2 fold. AxP: traces of the axial plane.

terpreted as having originally pertained to the upper mantle of the Apulian microplate (Piccardo et al. 1990).

All these rocks were involved in the Alpine subduction-related tectonic events and brought to a depth exceeding 70 km; they underwent metamorphic re-equilibration under eclogite-blueschist facies (High Pressure – Low Temperature) conditions to different degrees. The HP-LT conditions are testified to by the Na-cpx + grt + rt ± Na-amp ± ep association in the mafic rocks (mineral abbreviations after Kretz 1983), and by the ph + pa + grt + rt ± zo association in the metasedimentary rocks (Cortesogno et al. 1977a, Messiga et al. 1983, Piccardo et al. 1988, Messiga & Scambelluri 1991, Capponi et al. 1994a).

Different metamorphic re-equilibrations at decreasing pressure indicate exhumation of the subducted slices: a glaucophanic stage is testified to by the Na-amp + grt + ttn + Fe-ox association in the mafic lithologies and the phe + cld + chl + rt association in the metasediments; a following albite - amphibolite stage produced the bar + ab + ep + ox ± chl association in the mafic rocks, and the phe + cld + chl + rt association in the metasediments. The final re-equilibration in the greenschist facies conditions is testified to by the act + ab + ep + chl + ttn + Fe-ox association in the mafic rocks, and by the mu + chl + ab + ttn ± ox association in the metasediments.

Due to different rates of re-equilibration, the early metamorphic stages are best preserved in the mafic rocks, whereas the greenschist facies assemblages are dominant in the metasediments.

Km-scale bodies of lherzolite preserve mantle features and mineral associations, showing that they have escaped from the Alpine overprint. These bodies are wrapped in severely deformed serpentinites, testifying to the heterogeneous distribution of the Alpine deformation and metamorphism.

The lack of Alpine metamorphism and deformation led most researchers to assign the preserved lherzolite to the so-called Erro-Tobbio Unit (Chiesa et al. 1975), and to postulate its emplacement on the Voltri Massif by post-collisional thrusting phases. More recently, the discovery of eclogite parageneses in Erro-Tobbio rocks (Piccardo et al. 1988, Capponi & Crispini 1990, Scambelluri et al. 1991) has shown their involvement in the Alpine subduction events, making the definition of the Erro-Tobbio Unit questionable.

To the east (Fig. 1) the Voltri Massif is in tectonic contact with the blueschist Ligurian-Piemontese units of the Sestri-Voltaggio Zone (Cortesogno & Haccard 1984, Hoogerduijn Strating 1994); to the west and south the Ligurian-Piemontese units of the Voltri Massif tectonically overlay the Briançonnais Calizzano-Savona Unit. This unit comprises a pre-Alpine crystalline basement and a Carboniferous-Permian cover and is affected by blueschist facies Alpine metamorphism in its inner part. To the north-west the Voltri Massif overlays the Piemontese Valosio crystalline Massif, which experienced Alpine eclogite metamorphism (Cabella et al. 1991a, Messiga et al. 1992). To the north the Voltri Massif is overlain by the sedimentary formations of the Tertiary Piemontese Basin, which is a late- to post-orogenic molassic basin.

Units of the Ligurian Alps experienced subduction and related HP-LT metamorphism at different times: an older subduction involved the Ligurian – Piemontese units, and the Piemontese Valosio crystalline Massif (Desmons et al. 1999). A younger subductive event involved the Briançonnais units, as testified to by the occurrence of HP-LT assemblages in the Eocene sedimentary rocks (Cabella et al. 1991b). Differences in the time of subduction have been explained through the migration of the subduction plane (Vanossi et al. 1984) towards the foreland, thus involving more and more external units. The post-Eocene underthrusting of the Briançonnais units is claimed to have a role in the thermal evolution of the already eclogitized units of the Ligurian Piemont ocean (Messiga & Scambelluri 1991).

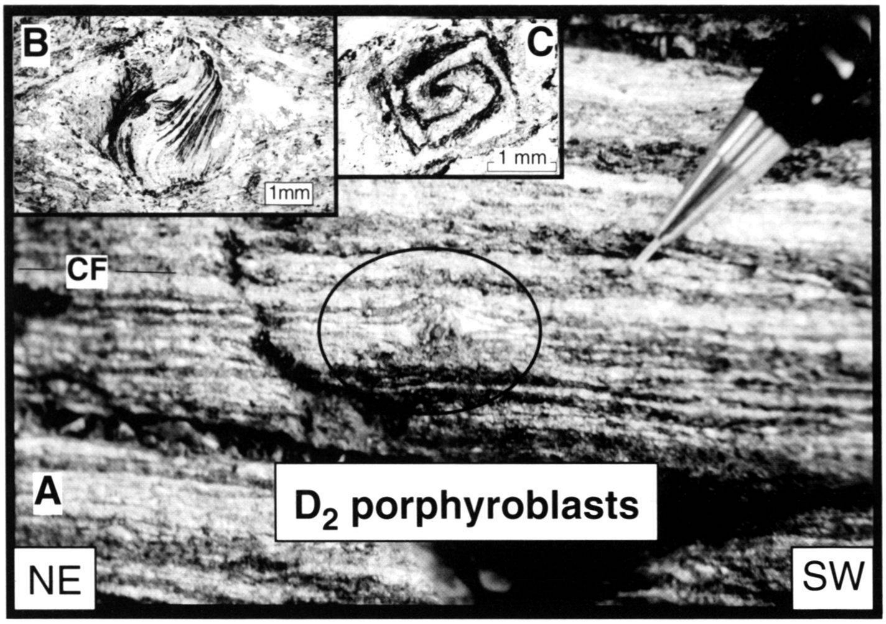

Fig. 3. A. D_2 asymmetrical σ-type mantled porphyroblast in metasediments. CF is the composite fabric.
B. Microphotograph of a pseudomorph on σ-type mantled porphyroblast. ppl.
C. Microphotograph of inclusion patterns in pseudomorph on D_2 porphyroblast. ppl.

3. Structural analysis

Structural analysis shows evidence of several different superimposed deformation events which occurred under different tectonic conditions, ranging from a ductile to a brittle regime. The oldest structures testify to the deformations attained during the HP-LT stage of the tectono-metamorphic evolution; in general the structures linked to the prograde metamorphic path are not preserved. A first glance at the structural setting highlights some differences between the western and the eastern sectors of the Voltri Massif in the orientation of structures, in the magnitude of deformation and in the degree of metamorphic reactivaton.

3.1. High Pressure – Low Temperature deformations

The oldest deformations are related to the HP-LT stage of the tectono-metamorphic evolution. Their main manifestations consist of an eclogite facies foliation and rootless hinges of isoclinal folds, in cm- to km-scale preserved domains. Although these structures occur all over the massif, they have no continuity and they cannot easily be correlated between sites.

3.2. Greenschist facies deformation

3.2.1. D_1 and D_2 structures

The D_1 and D_2 structures (F_1 and F_2 in Capponi 1987, 1991, Crispini & Capponi 1997, or D_1/D_2 folds in Capponi et al. 1999b) are widespread at the scale of the entire massif and are contemporaneous with retrogressive metamorphism, evolving from Na-amphibole greenschist facies to greenschist facies s.s. metamorphic conditions. The D_1 folds are tight to isoclinal similar folds, with pervasive schistosity parallel to the axial plane and occur from mm- to km-scale. Superposition of the D_1 on the pre-D_1 folds locally gave rise to type 3 (Ramsay 1967, Ramsay & Huber 1987) interference patterns. Due to strong transposition related to the schistosity, the D_1 fold closures are obliterated in places.

The D_2 folds are mm- to km-scale similar folds, with interlimb angle ranging from tight to open. Overprinting of the D_2 on the D_1 folds caused type 3 (Fig. 2) and rarely type 2 interference patterns. D_2 schistosity is parallel to the axial plane and is unevenly developed: in tight, similar folds it is pervasive and obliterates fold closures, becoming sub-parallel to the D_1 schistosity. The D_1 and D_2 schistosity, plus the previous foliations (including the lithological surface), caused a composite fabric (hereafter indicated as CF) which is the most evident surface in the field and controls the contacts between the different lithologies.

In places, convergence in style and lack of continuous outcrop make the distinction between the D_1 and D_2 folds problematic. Under these circumstances, the resulting structures have to be referred to as undifferentiated D_1/D_2 folds. From a geometrical point of view, the D_1/D_2 folds can be classified as 1C-2 type (Ramsay 1967) or in the range of D and E shapes and with amplitude number 3-5, according to Hudleston (1973). In three dimensional terms, D_1/D_2 folds range from cylindrical to moderately non-cylindrical, to truly sheath-like shape (Capponi et al. 1994b).

The occurrence of asymmetrical δ- and σ-type mantled porphyroblasts and porphyroclasts testifies to a non-coaxial progressive deformation (Simpson & Schmid 1983, Passchier & Simpson 1986, Hanmer & Passchier 1991). In the eastern sector of the Voltri Massif, estimates on some microscopic and

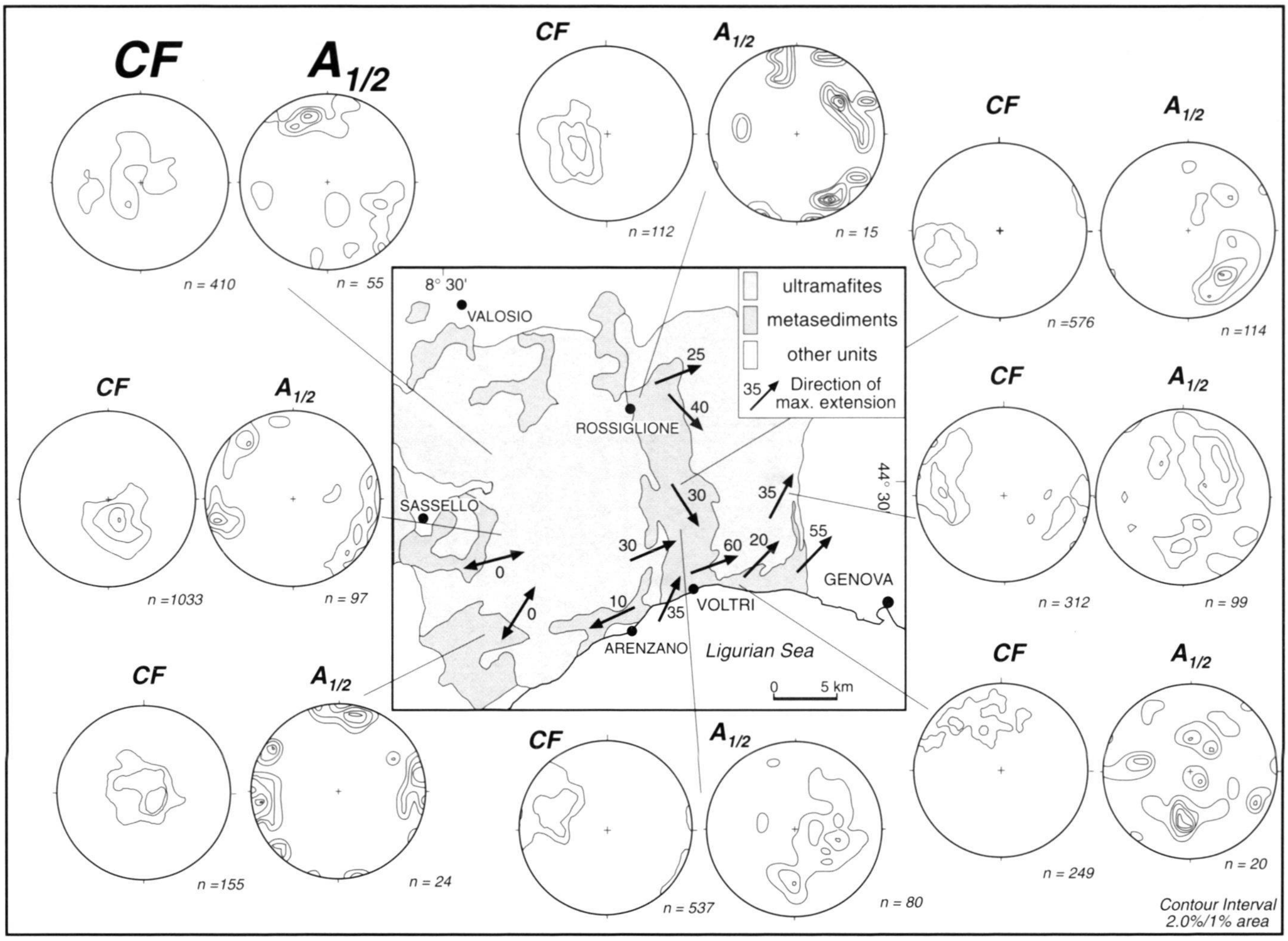

Fig. 4. Synoptic map showing the orientation of the composite fabric (CF) and D_1/D_2 axes ($A_{1/2}$) in different sectors of the Voltri Massif. Arrows indicate trend and plunge of direction of maximum extension deduced from EPF structures.

mesoscopic structures indicate high values of shear strain (γ). Inclusion patterns (Fig. 3) in rotated porphyroblasts (Crispini 1995) return a $\gamma \approx 10$ after the estimate based on the angle of rotation by Rosenfeld (1970). Similar values are obtained analysing the shape of sheath folds (Crispini 1995) following the estimate by Skjierna (1989).

The D_1/D_2 direction of maximum extension (X direction) is indicated by stretching lineations, syntectonic fibres, pressure shadows and oriented re-crystallization of quartz; at many sites the X direction is sub-parallel to the D_1/D_2 hinge lines. The stretching lineations consist of tourmaline, feldspar, aggregates of chloritoid, quartz and graphite, pseudomorphs after pre-existing minerals, such as lawsonite, garnet, chloritoid and epidote, and the lattice-preferred orientation of quartz.

In the western sector the CF is usually low dipping to sub-horizontal, whereas in the eastern sector it is steeply dipping to the east (Fig. 4). Due to the non-cylindrical shape of the D_1/D_2 folds, hinge lines are scattered on the axial plane, which nearly coincides with the CF. Accordingly, they have different attitudes in the western and eastern sectors of the Massif. Due to the complexity caused by the later deformations, no regional information can confidently be derived from the vergence of the D_1/D_2 folds.

3.2.2. Ductile structures recording Extension Parallel to the Foliation (EPF)

The D_1/D_2 folds are overprinted by ductile structures, which indicate extension parallel to the foliation (hereafter EPF), such as shear bands (Fig. 5), boudins, foliation boudinage and extensional veins (Capponi & Crispini 1997). These structures indicate an extension acting sub-parallel to the main planar anisotropy (Platt & Vissers 1980), and are distinctive of areas deformed by high shear strain (White et al. 1980, Ramsay & Huber 1983).

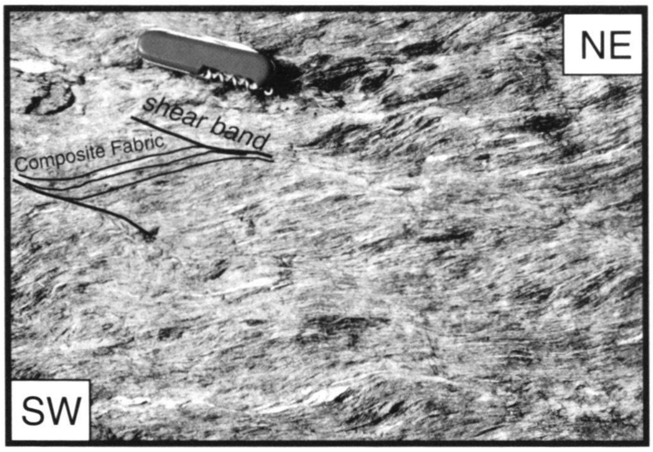

Fig. 5. Shear bands in metasediments with dextral sense of shear. Varenna River, south-eastern part of the Voltri Massif (see Fig. 1). Shear bands extend the composite fabric, which here is sub-vertical; the extension expressed by these structures is not necessarily related to extensional tectonics.

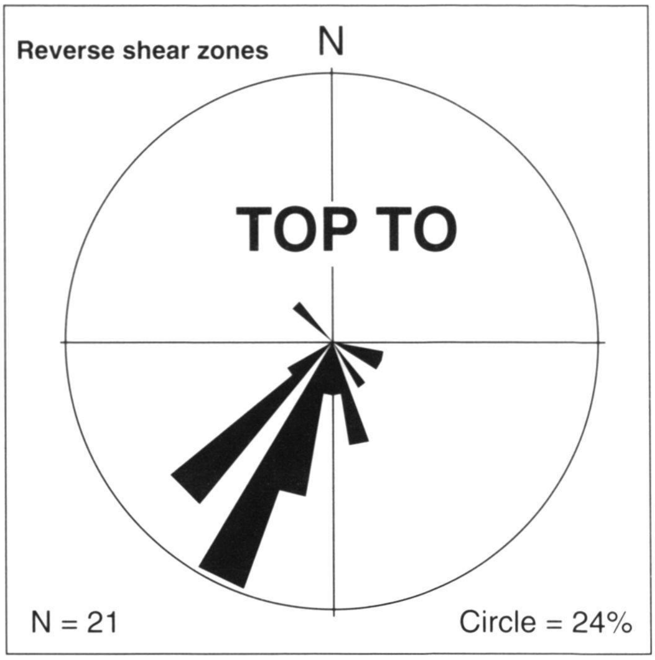

Fig. 7. Rose diagram of the top to directions deduced from RSZ.

The extensional cleavage displays the features of the structures described as C' planes (Berthé et al. 1979), extensional crenulation cleavage (Platt & Vissers 1980), shear bands and shear band cleavage (White et al. 1980), and asymmetrical extensional shear bands (Hanmer & Passchier 1991); hereafter we will use the term shear bands. Recrystallised calcite, chlorite and quartz are often found with dynamic features along the shear bands. Recrystallised calcite, albite and quartz are common in the neck zones between adjacent boudin structures. The same minerals occupy the extensional veins. These metamorphic growth episodes indicate that the EPF structures are coeval with greenschist facies metamorphism. The data on maximum extension derived from the EPF (Capponi & Crispini 1997) indicate an overall shallow dip

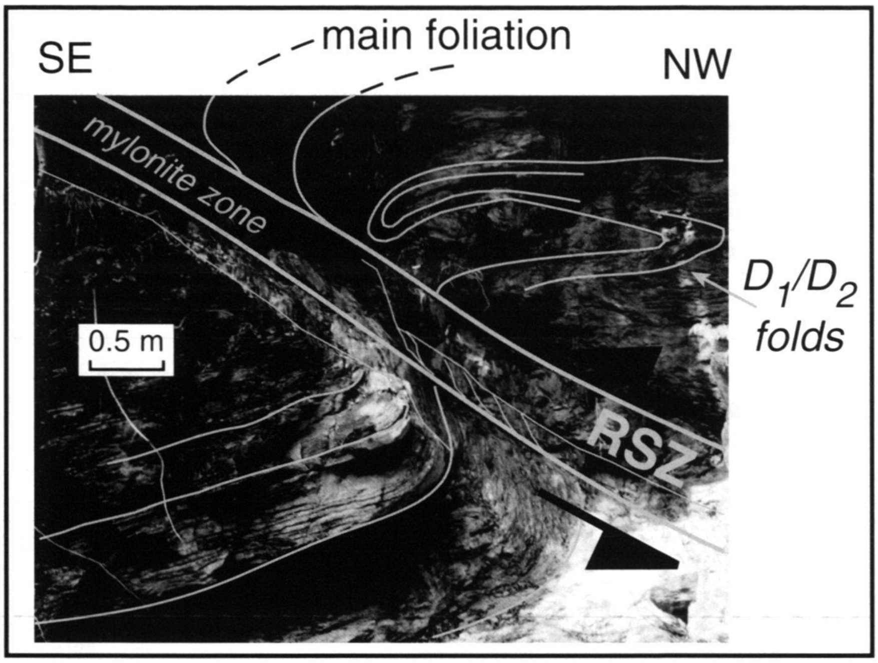

Fig. 6. Reverse shear zone in the western sector of the Voltri Massif.

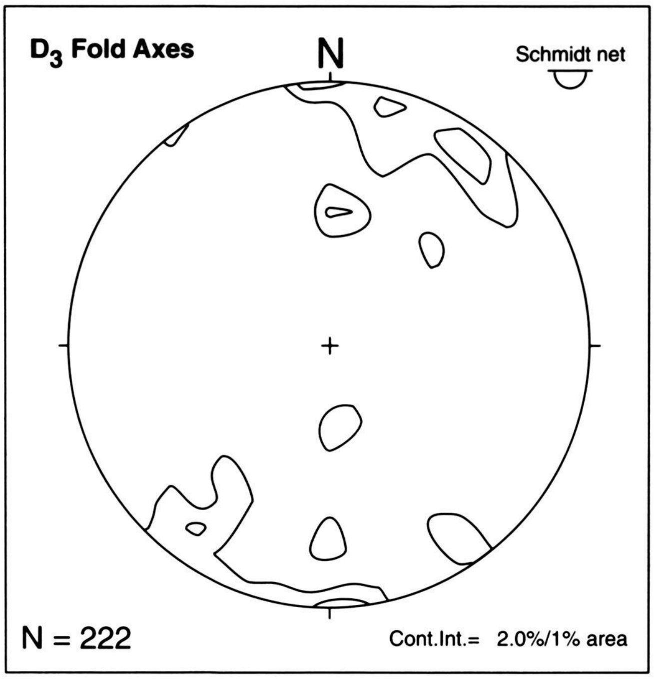

Fig. 8. D_3 fold axes.

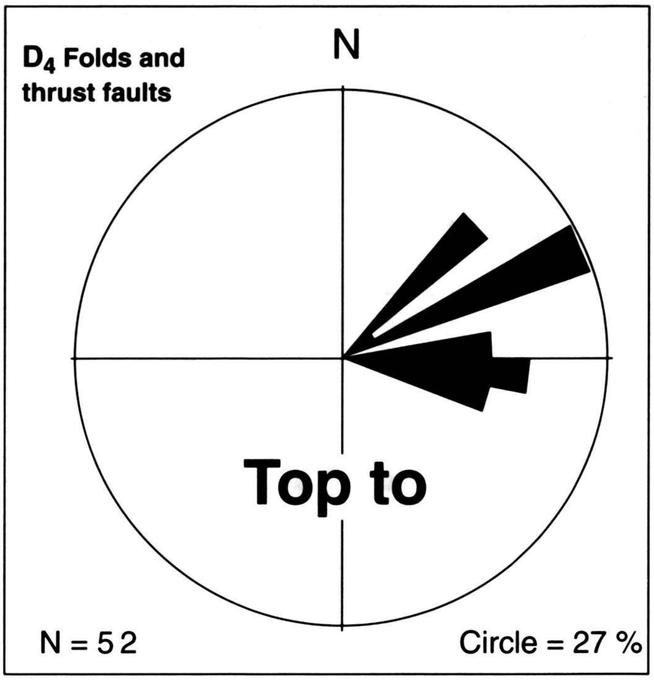

Fig. 9. Rose diagram of the top to directions deduced from D_4 structures.

to the NE or the SW (Fig. 4). EPF structures occur all over the Voltri Massif, but they are more frequent in the eastern sector.

3.2.3. Reverse shear zones (RSZ)

Both the D_1/D_2 folds and the EPF structures are overprinted by contractional, ductile to brittle-ductile shear zones (Fig. 6). The shear planes are decorated with greenschist to low greenschist facies mylonites and protomylonite (nomenclature after Sibson 1977). The kinematic analysis of the shear planes and related mylonites on the whole massif has revealed the occurrence of two sets of structures, with contrasting orientations. A prevalent set of reverse shear zones (hereafter indicated as RSZ) with a top to SW sense of shear coexists with minor top to SE and NW structures (Fig. 7).

3.3. D_3 deformation structures

The D_3 deformation is characterized by folds and extensional shear zones, coeval with low greenschist facies metamorphic conditions. The main structures are m- to 10 m-scale parallel open chevron folds, in places accompanied by a roughly spaced cleavage, rarely schistosity. Pressure solution and rotation of pre-existing minerals are the main deformation mechanisms that contribute to the development of cleavage; pressure solution is prevalent in carbonatic metasediments. The D_3 schistosity is marked by chlorite, calcite, epidote and phyllosilicates. In the Hudleston (1973) classification, the D_3 folds fall into the F2 field. The D_3 fold axes are gently to moderately plunging to both NNE and SSW (Fig. 8), with nearly horizontal axial planes (Capponi 1991, Capponi et al. 1994a).

Ductile to brittle-ductile shear zones occur at the outcrop scale, and are gently to moderately dipping to the east. These shear zones are characterized by S-C mylonites, with crystallization of calcite, chlorite, quartz and talc. Mylonites and the drag of pre-existing foliations indicate an extensional sense of shear.

The D_3 folds occur at sites where the CF is steeply-dipping to vertical (namely in the eastern sector of the Voltri Massif), whereas shear zones occur where the CF is gently-dipping to sub-horizontal (namely in the western sector). The genetic link between the two kinds of structures will be discussed later (Paragraph 5).

3.4. D_4 deformation

The D_4 deformation is expressed by both folds and thrust faults. The D_4 folds are parallel open folds, with wavelengths up to 10 km, and are occasionally characterized by a rough cleavage. In the Hudleston classification (1973) these structures fall into the D2-E2 field. The D_4 fold axes are nearly horizontal and are N-S to NNW-SSE trending; axial planes strike N-S, and dip moderately to the west. The D_4 folds are strongly asymmetrical and indicate a top to E-NE sense of shear. They can evolve in W-SW dipping thrust faults (Capponi et al. 1999a, 1999b), with a top to E-NE sense of shear (Fig. 9). Fold related cleavage domains and thrust surfaces are occasionally coated with zeolite facies minerals. At most sites thrust planes are decorated with crush-breccia, with minor crush microbreccia, protocataclasite and cataclasite, rarely pseudotachylyte. The occurrence of these fault rocks points to a deformation at shallow crustal levels. Crystallization of zeolites (Cortesogno et al. 1977b) and data from fluid inclusions (Crispini & Frezzotti 1998) indicate a temperature lower than 250 °C and pressure around 2 Kb. A similar estimate is obtained from the recrystallization of chlorite and carbonates on the slickensides. The D_4 deformation also involved the Oligocene formations of the Tertiary Piemontese Basin, whose strata are locally folded and tilted to vertical.

The D_4 structures occur over the whole Voltri Massif, apparently with the same features. In the west, however, they more frequently involve the Oligocene formations and hence are bracketed in time.

4. Age of deformations

The youngest rocks involved in the D_1, D_2 and D_3 deformation structures are the metasediments; their age can be inferred by comparison with the Schistes Lustrés of the Western Alps, which are Late Cretaceous (Lemoine et al. 1984). Deformations are clearly younger, but in most cases their age is poorly to not-constrained, also due to the general lack of reliable radiometric data. One relevant exception for the HP-LT stage is a Middle Eocene Ar/Ar age of 45.2±1.8 Ma (Hoogerduijn Strating 1991) of a phengite from an eclogitic metabasite.

The retrogressive metamorphism and the related deformations are poorly constrained by whole rock K/Ar determinations on metasediments, spanning from 41 to 36 Ma (Schamel 1974). Since the deformations D_1, D_2 and D_3 do not involve the formations of the Tertiary Piemontese Basin, they have to be younger than the Late Eocene-Early Oligocene time interval.

On the contrary, the D_4 deformation also involves the Oligocene formations of the Tertiary Piemontese Basin, and its minimum and maximum relative ages are constrained within a short time span: deformed late Oligocene formations are covered by undeformed Burdigalian sediments, thus fixing the age of the D_4 in the Early Miocene (D'Atri et al. 1997, Capponi et al. 1999b).

5. Structural interpretation

The deformational history derived from structural analysis may be considered as characterized by four main tectonic phases (Table I).

The oldest structures occur in preserved domains that escaped later greenschist facies metamorphism and deformations. Heterogeneous distribution of strain related to lithological contrast determined the degree of structural reworking, in turn triggering the retrogressive metamorphism. The dependence of metamorphic reactivation on the structural reworking (Beach 1980) is very common in the metamorphic units of the Alpine belt (Pennacchioni & Guermani 1993), and elsewhere (Shelley & Bossière 1999): as a result, poorly retrogressed domains also retain early structures, whereas no early structure is observed in domains that are severely reworked in greenschist facies metamorphism. Mineral assemblages in preserved domains indicate HP-LT conditions of deformation that can be related to subduction (Scambelluri et al. 1995, 1997). As the greenschist metamorphic reactivation is more widespread and complete in the metasediments, the subduction-related structures are observed in the mafic lithologies, namely in eclogite. Preserved eclogites form discontinuous bodies and pods floating within highly retrogressed greenschist facies metabasite and metasediments: as a consequence, in most cases HP-LT structures and foliations are severely disrupted and have no regional continuity.

The D_1/D_2 folds, EPF and RSZ developed during retrograde metamorphism, decreasing from Na-amphibole greenschist facies to greenschist s.s. facies and to low greenschist facies conditions. The tectonic event can be linked to a post-subduction phase of uplift and exhumation of the metamorphic units and nappe emplacement. As D_1/D_2 folds, EPF and RSZ developed during progressively decreasing metamorphic conditions, related to the same tectonic event, they can be regarded as the result of different steps of a progressive deformation instead of phases separated by significant breaks in time and/or in metamorphic conditions.

The first step in this process led to repeated folding of preexisting foliations by strongly non-cylindrical isoclinal folds (i.e. the D_1/D_2 folds) accompanied by the development of a pervasive schistosity and related stretching lineation. The δ- and σ-type mantled porhpyroclasts indicate a non-coaxial progressive deformation (Hoogerduijn Strating 1991, Crispini 1995), with $\gamma \approx 10$.

Continuous and progressive shearing led to the subsequent development of EPF (structures recording extension parallel to the foliation), such as shear bands, boudinage and extensional veins, under decreasing metamorphic conditions. EPF structures evolved when the maximum extension direction of the incremental strain ellipse approached or was contained in the plane of the D_1/D_2 schistosity, during the non-coaxial deformation. As discussed by Wheeler & Butler (1994), EPF structures do not imply regional extension. If the schistosity is close to the horizontal before deformation, then the maximum extension direction is also nearly horizontal. Only in this case, can structures that extend the foliation be related to an extending lithosphere. Nevertheless in the internal parts of mountain belts, such as the Voltri Massif, the structure is complex and no hypothesis on the pristine orientation of the foliation can be made with confidence (Butler & Freeman 1996).

The development of RSZ (reverse shear zones) was the last step during the progressive deformation of the tectonic phase of exhumation and nappe emplacement. Though deformed by younger deformations, the shear planes can be restored to their pristine position with confidence. The coexistence of top to NW RSZ together with top to SW RSZ is puzzling. As the present-day trend of the Ligurian sector of the Alpine belt is WNW-ESE, a top to SW sense of shear is in agreement with tectonic transport towards the foreland during the nappe emplacement; the top to NW shear planes can possibly be interpreted as backthrusts.

The D_3 deformation developed chevron folds and normal shear zones, under low greenschist facies metamorphic conditions. The D_3 folds have sub-horizontal axial planes and occur at sites where the CF is steeply dipping to vertical (i.e. in the eastern sector). These folds collectively achieve a vertical shortening/horizontal extension, that can interpretatively be linked to a phase of collapse; this tectonic phase was probably the response to the crustal thickening resulting from the nappe stacking. D_3 folds were not detected where the CF is gently dipping to horizontal (i.e. in the western sector). On the contrary, at these sites, normal shear zones occur, which result in a vertical shortening. As the two kinds of structure achieve the same result of vertical shortening/horizontal extension, under the same metamorphic conditions, we conclude that they are genetically linked to the tectonic phase of collapse. The development of folds or shear zones is driven by the pristine attitude of the regional foliation.

D_4 deformation points to an Early Miocene tectonic phase, with top to E-NE direction of tectonic transport. This tectonic phase is coeval with the rotation of the Corsica-Sardinia block (which ends at 19 Ma, Montigny et al. 1981) and was defined as "fase Ligure III" by Mutti et al. (1995). This phase led to a general tectonic transport towards the E-NE (backthrusting), expressed by the emplacement of the Internal Ligurian units on the Sub-Ligurian and Tuscan domains and to the early thrusts that accommodated displacements towards the N into the future Po Plain area.

6. Discussion

D_4 folds and thrusts appear to occur with similar features over the whole Voltri Massif. On the contrary, deformations older than D_4 are unequally developed in different sectors of the Massif. D_3 deformation accommodates vertical shortening/horizontal extension by folds or shear zones, depending on the attitude of the regional foliation: as a result, D_3 structures are different in the eastern and western sector of the Voltri Massif.

Significant differences also exist in the greenschist facies progressive deformation, namely in the D_1/D_2 deformations.

i) In the eastern sector:
- stretching lineation is more intensely developed and the rock fabrics are SL-tectonites, suggesting strain at the boundary between constriction and flattening field;
- rocks underwent a more severe metamorphic reactivation and re-equilibration in greenschist facies conditions. Relics of eclogite assemblages are rare;
- sheath folds are frequent;
- strain magnitude, as deduced by rotated porphyroblasts and sheath folds, appears to be high, with values of γ up to 10.

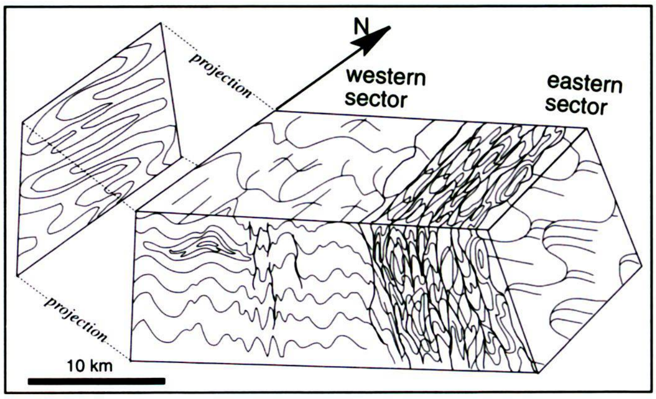

Fig. 10. Interpretative 3D reconstruction of the structural setting of the Voltri Massif, showing different attitude and features of D_1/D_2 folds and schistosity, in the western and eastern sector. D_3 and D_4 removed.

ii) In the western sector:
- stretching lineation is weakly developed and the rocks display the features of S-tectonites, suggesting strain in the apparent flattening field;
- rocks suffered a less severe re-equilibration in greenschist facies conditions: eclogite relics are not as rare as in the eastern sector;
- sheath folds occur, but are not frequent;
- strain magnitude appears to be lower than in the eastern sector.

All these observations indicate an uneven distribution of strain with the D_2 strain intensity increasing from the west to the east. We note that the strain intensity increases moving towards the Sestri-Voltaggio Zone, where the steepness of the regional foliation increases. Both the dip change of the regional foliation and the increase in strain intensity could tentatively be explained by strain partitioning along the boundary with the Sestri-Voltaggio Zone (Fig. 10). This scenario implies the development of a regional scale greenschist facies shear zone along the tectonic contact with the Sestri-Voltaggio Zone during the development of the D_2 deformation; the result is a partitioning of deformation at the scale of the entire massif, with the D_2 strain intensity increasing from the west to the east. The location for the development of this major shear zone was probably driven by the competence contrast between different rock types, i.e. the ophiolitic rocks of the Voltri Massif and the carbonatic rocks of the Gazzo-Isoverde Unit of the Sestri-Voltaggio Zone.

The mode of development that we describe matches most of the features of the strain partitioning model of Dewey et al. (1998) for transpression/transtension zones. The planar and linear fabrics observed in the eastern Voltri Massif can result from either a transpressional or transtensional regime (Fossen & Tikoff 1998); nevertheless, on the whole, the greenschist facies structures are consistent with a contractional setting and a transpressional regime is more likely to have acted at this stage.

Deformation	Tectonic event	Fabric	Metamorphism	Vergence
Early (preD_1) deformations	subduction	rootless hinges of isoclinal folds and related schistosity	eclogite to blueschist facies	unconstrained
D_1/D_2	exhumation-uplift of the metamorphic units and nappe emplacement	tight to isoclinal similar folds and related schistosity	Na-amphibole greenschist to greenschist facies s.s.	unconstrained
EPF extension parallel to the foliation	nappe emplacement	shear bands, boudinage, foliation boudinage, extensional veins	greenschist facies	scattered directions, with a NE-SW maximum
RSZ reverse shear zones	nappe emplacement	reverse, ductile to brittle-ductile shear zones	greenschist to low greenschist facies	top to SW minor top to SE and NW
D_3	post-uplift collapsing	chevron folds with sub-horizontal axial plane ductile to brittle-ductile normal shear zones	low greenschist facies	vertical shortening horizontal extension
D_4	backthrusting	open folds thrust faults	zeolite facies	top to E-NE

Tab. I. Summary of the relationships between deformations, tectonic events, fabrics and metamorphism.

Although the boundary with the Sestri-Voltaggio Zone was probably the primary site of shearing, the whole eastern sector of the Voltri Massif can be considered as having acted as a regional-scale shear zone. Transpression along this lineament possibly accommodated different rates of deformation and displacement along the Alpine belt, thus acting as a tear fault. This adds complexity to the structural evolution of the Sestri-Voltaggio tectonic lineament, which represented a site in which different tectonic regimes were active at different times and at different structural levels (Crispini & Capponi in press).

The Voltri Massif is one of the widest complexes of metamorphic ophiolites with metasediments in the western Alps; the comparison with other metaophiolite massifs such as the Monviso and the Zermatt-Saas zones is an obvious step. Though the Zermatt-Saas zone has close analogies in the primary lithologies (Bearth 1967), it suffered an Ultra High Pressure metamorphic imprint, testified to by the occurrence of coesite in metachert (Reinecke 1991); furthermore the greenschist facies retrogressive metamorphism is less widespread than in the Voltri Massif. The similarity is more appropriate to the Monviso ophiolites, which display analogies in structural and metamorphic history (Philippot 1990). Structural features point to a polyphase evolution, with different sets of superposed deformations. Older structures show two phases of isoclinal folding (F1 and F2 after Philippot 1990) developed under HP-LT conditions. Philippot (1990) also documented the post D_1/D_2 development of shear bands; the late phase of folding produced open and kink folds. On a large scale, Philippot (1990) questioned the classical interpretation of the Internal Pennine Zone as a backthrusting system superimposed on a westward overthrusting system and suggested coeval development of opposite vergence during nappe emplacement. Strain partitioning is claimed as a fundamental process, at all scales; as in the Voltri Massif, undeformed domains (cm to km scale) are preserved between zones of ductile deformations. In turn, an uneven distribution of strain controls the development of retrograde greenschist metamorphism.

Nevertheless the structural evolution of the Voltri Massif also exhibits some peculiar features with respect to the other metaophiolite complexes of the Western Alps: the uneven D_1/D_2 strain distribution, possibly driven by the transpressional regime at the boundary with the Sestri-Voltaggio Zone, does not match any features of the Monviso structural history. We assume that this is linked to the peculiar position of the Voltri Massif in the Western Alps, during the progressive strengthening of the arcuate shape of the belt. Similarly, the Early Miocene thrust and fold phase (i.e. the D_4 of the Voltri Massif) is not reported in the literature of the Monviso. Top to E-NE structures are related to the Oligocene-Miocene main phase of the anticlockwise rotation of the Corsica-Sardinia block and to the backthrusting of the southern termination of the Western Alps towards the Po plain. The strengthening of the arcuate shape of the Alpine belt is particularly effective in the Ligurian Alps and namely in the Voltri Massif, due to its position at the very E-SE termination of the Ligurian Alps. Hence Early Miocene top to E-NE thrust and folds are again a peculiar feature of the structural evolution of the Voltri Massif.

7. Conclusions

The signature of the Alpine tectonics in the Ligurian-Piemontese units is characterized by the multiple superposition of deformative structures. The complex deformation history can be unravelled by recognizing four main tectonic phases, which can be distinguished by their metamorphic and structural features.

The earliest structures are disrupted HP-LT folds and foliations in preserved domains, which are the tectonic imprint acquired during the involvement of the Voltri rocks in a subduction zone.

The following greenschist facies retrogressive metamorphism evolved from Na-amphibole greenschist, to greenschist s.s. to low greenschist facies conditions and testifies to the tectonic phase of uplift, exhumation and nappe emplacement. The structures related to this stage have been developed in a progressive shear strain regime and evolved from isoclinal folds to structures recording extension parallel to the foliation and to ductile to brittle-ductile reverse shear zones. During the early stage of this phase, the strain gradient from the western to the eastern sector of the Voltri Massif suggests that a transpressional regime acted along the boundary with the Sestri – Voltaggio Zone.

A post-uplift phase of tectonic collapse is testified to by folds with sub-horizontal axial planes and normal shear zones, both of which achieve vertical shortening/horizontal extension under low greenschist metamorphic conditions.

Top to E-NE thrusts and folds are the signature of the Early Miocene phase of backthrusting, linked to the anticlockwise rotation of the Ligurian sector of the Western Alps. This phase led to the emplacement of the Internal Ligurian units on the Sub-Ligurian and Tuscan domains, and marks the transition to Apennine tectonics.

The results of our analysis imply that the Ligurian-Piemontese units of the Voltri Massif bear the inheritance of a complex and multiphase structural evolution, spanning from the subdution stage to the Oligocene-Miocene tectonics of the Ligurian-Provençal basin. Strain features of the greenschist facies deformations suggest the development of a major shear zone along the contact between the Voltri Massif and the units of the Sestri-Voltaggio Zone. This corroborates the observation that the Sestri-Voltaggio lineament was the site of different tectonic regimes at different times and at different structural levels.

Though the structural evolution of the Voltri Massif shows analogies with other metaophiolite complexes of the Alps, namely the Monviso zone, there are also significant differences: the uneven distribution of strain between the eastern and western sectors and the development of Early Miocene top to E-NE structures are peculiar features of the Voltri Massif. We assume that both are related to the position of the Voltri Massif at the very E-SE termination of the Ligurian Alps.

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