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Polyorogenic deformation history recognized at very shallow structural levels: the case of the Antola Unit (Northern Apennine, Italy)

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Abstract

The Antola Unit is a Ligurian Unit occurring at the top of the nappe pile of the Northern Apennines (Italy). Structural analysis indicates that the Antola Unit has been involved in a complex polyphase deformation history developed at a shallow structural level. The successive deformation phases belong to both the Alpine and Apenninic orogenic cycles. Therefore the Antola Unit is a good example of a polyorogenic tectonic unit deformed at a very shallow structural level. The older part of the structural evolution is assigned to the Meso-Alpine tectonic stage and consists of two deformation phases (D1 and D2) characterized by opposite tectonic transport directions. The D1 and D2 structures were sealed by the middle Eocene-Miocene deposits of the Tertiary Piedmont Basin. The younger part of the deformation history instead can be referred to the Northern Apennines tectonics, and consists of two phases (D3 and D4) both involving the Tertiary Piedmont Basin succession. The structural data indicate that the D1 deformations developed in mostly unlithified rocks, and that these structures can be related to the westward emplacement of the Antola Unit in the Alpine orogenic wedge. The D2 phase is characterized by folds and low-angle extensional faults, and is referable to a gravity-driven deformation. Both the D3 and D4 phases, characterized by folds and thrusts, are due to the compressive regime related to the development of the Northern Apennines.

Keywords: Polyorogenic tectonics; Antola Unit; Northern Apennine; Soft sediment and gravity driven deformations; Shallow structural levels

1. Introduction

A classic problem of structural geology is the study of polyorogenic tectonic units, which are represented in a large number of mountain belts. Usually these studies are performed in metamorphic settings, where the superposition relationships among the different generations of tectonometamorphic fabrics and the use of radiometric ages allow an exhaustive reconstruction of the tectonic history. The polymetamorphic Austroalpine basement units of the Alps represent one example, where the Paleozoic Variscan tectonometamorphic overprint has been widely overprinted by the Cretaceous-Tertiary HP metamorphism related to Alpine subduction (Compagnoni, 1977; Pognante et al., 1987; Biino et al., 1997; Nussbaum et al., 1998). By comparison, the study of polyorogenic nappes deformed at shallow structural levels is hampered by the scarce development of penetrative fabrics and by the lack of metamorphic assemblages to reconstruct the PT-paths.

In the Northern Apennines the Ligurian Units represent an interesting case of tectonic units involved in two subsequent orogenic cycles. These units have been deformed and structured during the late Cretaceous-Eocene Eo and Meso-Alpine W-verging phases and, starting from the Oligocene, they have been involved in the NE-verging phases related to the development of the Apenninic belt (Elter, 1973; Van Wamel, 1987). In this setting the Ligurian Antola Unit (AU) represents an example of a tectonic unit deformed at a very shallow structural level and characterized by a deformation history recording both the Alpine and the subsequent Apenninic orogenic phases. The first aim of this paper is to present the results of

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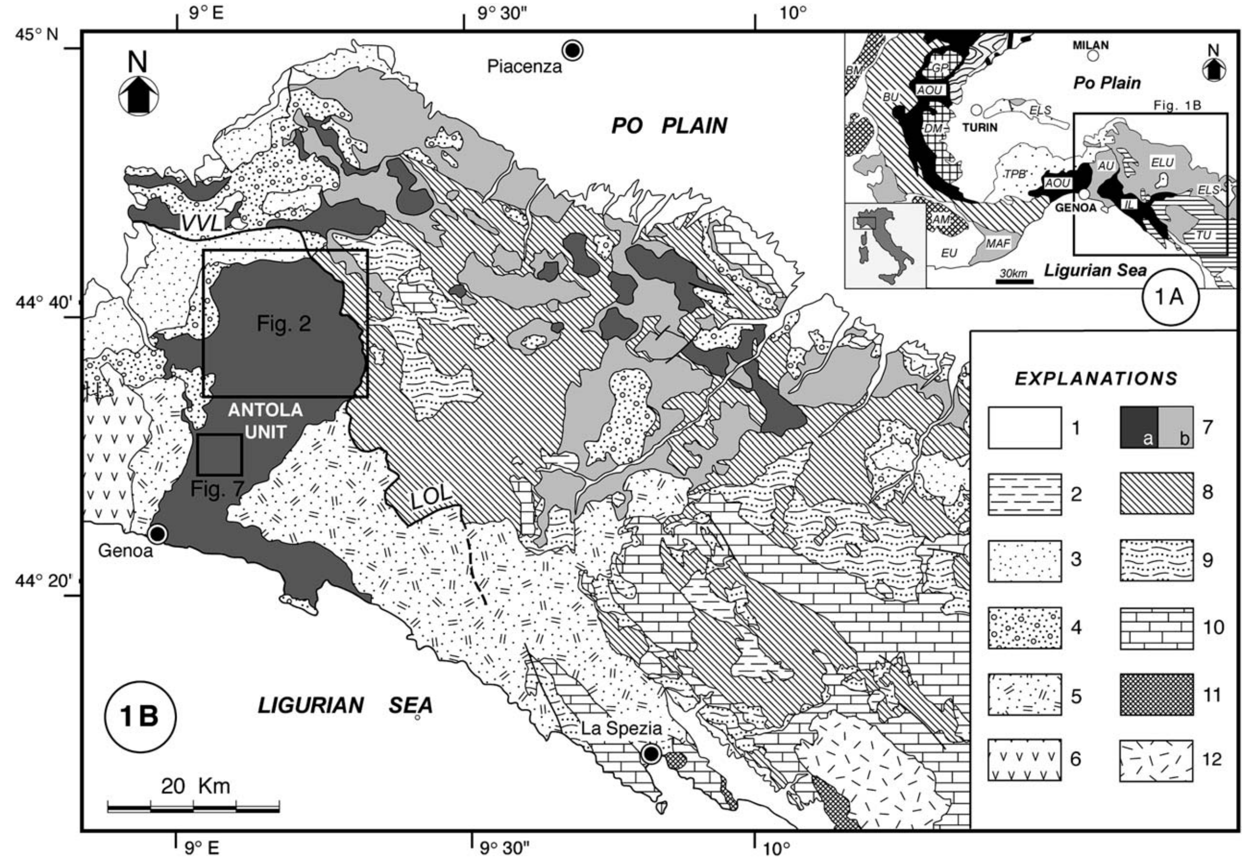

Fig. 1. Tectonic sketch map of the Northern Apennines and Western Alps (A) and close view of the Ligurian Apennines (B). LOL. Levanto-Ottone Line; VVL, Villalvernia-Varzi Line; AM, Argentera Massif; BM, Belledonne Massif; BU, Briançonnaise Units; GP, Gran Paradiso; DM, Dora Maira; AOU, Alps Ophiolitic Units; EU, European Units; MAF, Maritime Alps Flysch; TPB, Tertiary Piedmont Basin; ELS, Epiligurian Succession; AU, Antola Unit; ELU, External Ligurian Units; IL, Internal Ligurian Units; TU, Tuscan Units. 1, Quaternary deposits; 2, Intramontane Basins; 3–4, unconformable Epiligurian Succession and Tertiary Piedmont Basin (3, Miocene; 4, Upper Eocene-Oligocene); 5, Internal Ligurian Units; 6, Voltri Group; 7, Helminthoid Flysch unit without ophiolitic debris (a, Antola Unit; b, Eastern EL Units); 8, Helminthoid Flysch Units with ophiolitic debris (Western EL Units); 9, Subligurian Units; 10–12, Tuscan Units (10, Tuscan Nappe; 11, Massa Unit; 12, Apuan Alps).

a complete structural study performed in the northwestern sectors of the AU (Piedmont-Ligurian Apennines) (Fig. 1); the structural evolution of this unit with regard to the occurring deformation mechanisms and the geodynamics of the Alps-Apennines system is then discussed.

2. Geological background

The Northern Apennines is a collisional belt characterized by tectonic units referable to three main paleogeographic and structural domains: the Tuscan, the Sub-Ligurian and the Ligurian domains (Elter, 1973) (Fig. 1). The Ligurian Units occur at the top of the nappe pile and represent the remnants of the Ligurian-Piedmontese ocean and its transition to the Adria continental margin (Marroni et al., 2001). These units were strongly deformed during the Eo- and Meso-Alpine tectonic phases (late Cretaceous-mid Eocene) recording the oceanic closure and the subsequent continental collision between the Europe and Adria plates (Vescovi et al., 1999; Daniele and Plesi, 2000; Marroni et al., 2002b; Cerrina Feroni et al., 2004). The structures in these units are sealed by the Tertiary Piedmont Basin (TPB) and Epiligurian Successions (middle Eocene-Miocene) (Elter, 1973; Gelati and Gnaccolini, 1982; Ricci Lucchi and Ori, 1985; Bettelli et al., 1987; Mutti et al., 1995). During the mid-late Oligocene and Miocene-Pliocene (Apenninic stage) the Ligurian Units were thrust onto the more external domains belonging to the Adria continental margin (Sub-Ligurian and Umbrian domains) (Elter, 1975; Ricci Lucchi, 1986; Boccaletti et al., 1990; Conti and Gelmini, 1994; Cerrina Feroni et al., 2004).

The Ligurian Units are divided into two groups, representing two different paleotectonic settings: the Internal Ligurian Units (IL), representative of the Ligurian-Piedmontese oceanic basin, and External Ligurian Units (EL), interpreted as derived from the ocean–continent transition area towards Adria continental margin (Marroni et al., 2001).

The IL Units are characterized by remnants of Jurassic ophiolitic sequences and related sedimentary covers, showing pelagic, turbiditic, trench and lower slope deposits ranging in age from late Jurassic to early Paleocene (Marroni and Pandolfi, 2001). The tectonic history of the IL Units is characterized by a pre-Oligocene polyphase deformation, associated with metamorphism ranging from very low-grade to HP conditions, and which has been referred to the underplating and the subsequent exhumation in accretionary wedge, in a pre-collisional setting (Ellero, 2000; Marroni et al., 2004).

The EL Units are made by detached upper Cretaceous-Paleogene successions, characterized by the presence at the base of sedimentary deposits known as "basal complexes", consisting of shales and turbiditic deposits, associated with sedimentary melanges (Marroni et al., 2001 with references). These basal complexes are overlain by late Cretaceous carbonate-rich turbidites, known as Helminthoid Flysch, and by Paleocene-middle Eocene marly-shaly flysch deposits. According to the composition of the basal complexes, the EL units have been subdivided into two main groups (continent-ward and ocean-ward successions), referred to different palaeogeographic positions with respect to the Adria continental margin. The EL Units with basal complexes characterized by abundant ophiolitic and continental debris (i.e. Ottone and Caio units) have been assigned to an ocean-ward part of the basin (Western successions), while the EL Units with only continent-derived deposits have been referred to the outermost edge of the continental crust of the Adria plate (Eastern successions, i.e. Cassio Unit) (Marroni et al., 2001).

The EL Units were affected by polyphase deformation, characterized by pre-late Eocene phases related to the Meso-Alpine tectonics and by post-late Oligocene phases referable to the development of the Northern Apennines (Daniele and Plesi, 2000; Marroni et al., 2002a,b; Cerrina Feroni et al., 2002, 2004). The deformation developed under very low-grade metamorphic conditions (diagenesis/anchizone) (Costa and Bonazzi, 1991; Molli et al., 1992).

The IL Units are juxtaposed with the EL Units along the Levanto-Ottone line, regarded as high-angle thrust surface (Elter and Pertusati, 1973) or as a transpressive fault (Cerrina Feroni et al., 2002, 2004) (Fig. 1). SW this main tectonic line the IL are overthrusted by the AU lying at the top of the Ligurian stack, and characterized by a sedimentary succession analogous to those of the EL. These similarities suggested a common palaeogeographic setting located close to the Adria plate margin (Elter and Pertusati, 1973; Marroni et al., 2001; Cerrina Feroni et al., 2002; Catanzariti et al., in press).

3. The Antola Unit

The AU is a large nappe outcropping in a wide area of the Northern Apennines, NE Genoa to the Po Plain and SE Genoa along the Ligurian sea-coast (Fig. 1).

The AU succession can be roughly subdivided into three parts (Fig. 2; Catanzariti et al., in press): the lower "basal complex" (Montoggio Shale and Gorreto Sandstone, at least 400 m thick), the middle carbonatic turbidites (Mt. Antola Flysch) and the upper siliciclastic-carbonatic turbiditic deposits (Bruggi-Selvapiana Formation and Pagliaro Shale).

The Montoggio Shale (late Cenomanian-early Turonian) is represented by varicoloured and/or black manganesiferous hemipelagic shales, at the base inter-layered with fine-grained turbiditic sandstones. The Montoggio Shale is followed by the Gorreto Sandstone (Campanian), characterized by thin-bedded turbidites showing a mixed siliciclastic/carbonatic composition. The basal complex grades upward to the Mt. Antola Flysch (late Campanian-early Maastrichtian), consisting of thick calcareous turbidites and megaturbidites.

The Mt. Antola Flysch is overlain by the carbonatic megaturbidite sequence of Bruggi/Selvapiana Formation (Maastrichtian) characterized by the alternating of thick turbiditic marls with thin-bedded sandstones and shales. The top of the AU succession is represented by the Pagliaro Shale (late Maastrichtian-beginning of the late Paleocene), consisting of thin-bedded sandstones and calcareous turbidites, alternating with shales.

The absence in the AU basal complex of any ophiolitic debris confirms that the original palaeogeographic position of this succession can be referred to the continental-ward part of the EL domain according to Marroni et al. (2001) and Catanzariti et al. (in press).

The AU is overlain by the unconformable TPB succession (Fig. 2), beginning with the mid-Eocene Monte Piano Marl (Cavanna et al., 1989).

The AU is characterized by a complex structural evolution, deriving from the superposition of several tectonic phases, related to both the Alpine and Apenninic tectonics (Marini, 1981; Marroni et al., 1999, 2002a; Corsi et al., 2001) and developed under very low-grade metamorphic conditions (diagenesis-anchizone) (Costa and Bonazzi, 1991). This history can be subdivided into two parts: the first one is Meso-Alpine in age and is sealed by the middle Eocene TPB (D1 and D2 phases); the second is the Apenninic history and involves also the TPB (D3 and D4 phases).

3.1. The D1 phase deformations

The D1 deformations are mostly developed in the turbiditic sediments characterizing the AU basal complex consisting of the Montoggio Shale and the Gorreto Sandstone. In both the formations, and particularly in the Montoggio Shale, the D1 phase is characterized by a heterogeneous distribution of deformation, with some areas of high strain and others apparently undeformed.

In the shales the most common D1 structural feature is the scaly fabric, consisting of a net of anastomosing shear surfaces separating undeformed shale blocks, which preserve their internal sedimentary structures. The shear surfaces are polished and characterized by thin slickensides, without development of mineral fibers. The scaly fabric and the sedimentary bedding are sub-parallel.

In the sandstone-rich facies (Gorreto Sandstone and the base of the Montoggio Shale) the scaly fabric is associated with F1 folds and boudinage.

3.1.1. F1 folds

The F1 folds are open to sub-isoclinal and intrafolial, with axial planes oriented at low angle to the bedding in the adjacent strata and to the scaly fabric characterizing the surrounding shales. The F1 folds are characterized by curved hinges and axes scattered up to 90°. As a consequence the F1 folds display varied interference patterns with post-D1

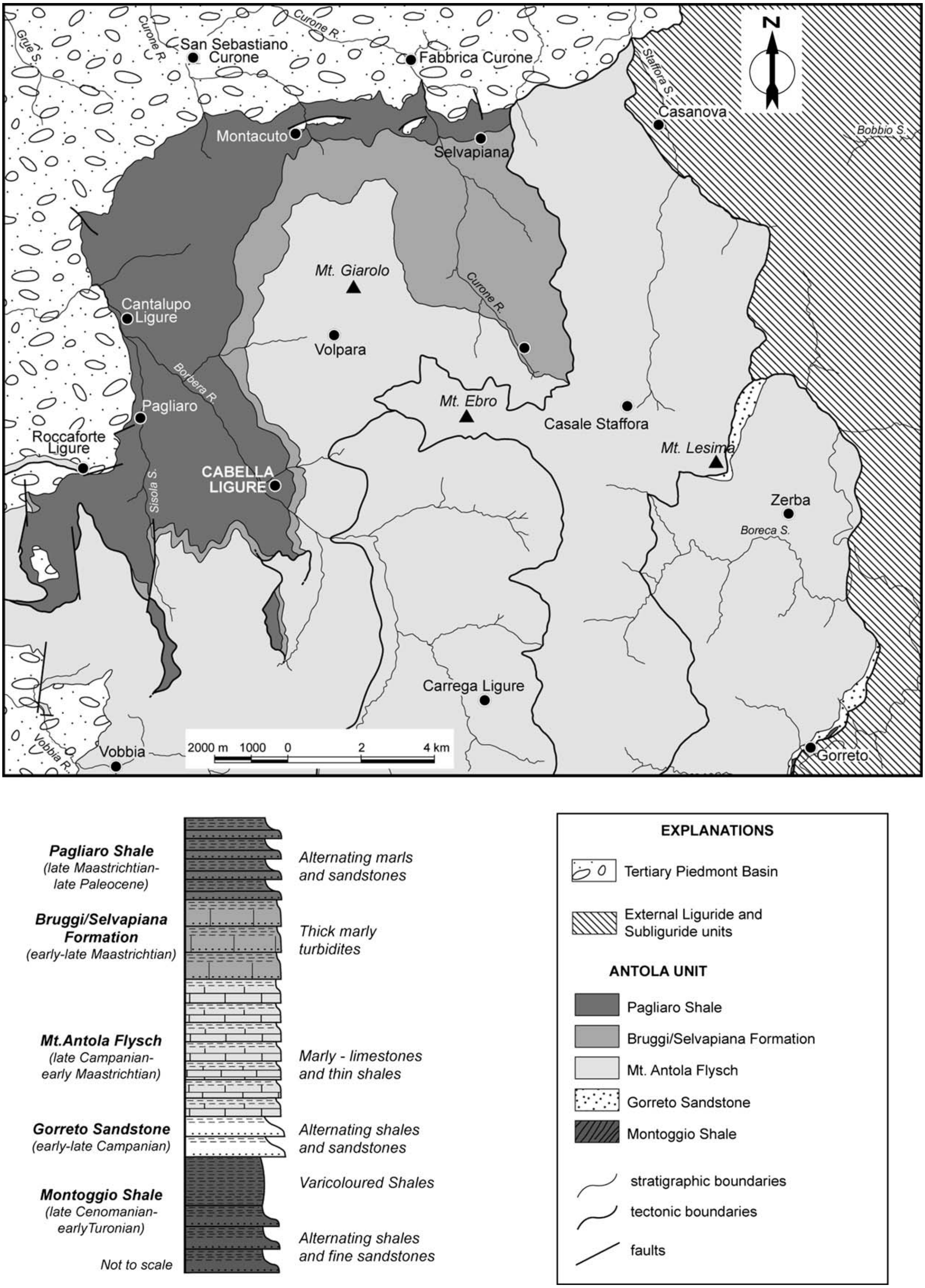

Fig. 2. (A) Geological sketch-map of the Northern Sector of the Antola Unit; (B) schematic stratigraphic column of the Antola Unit succession.

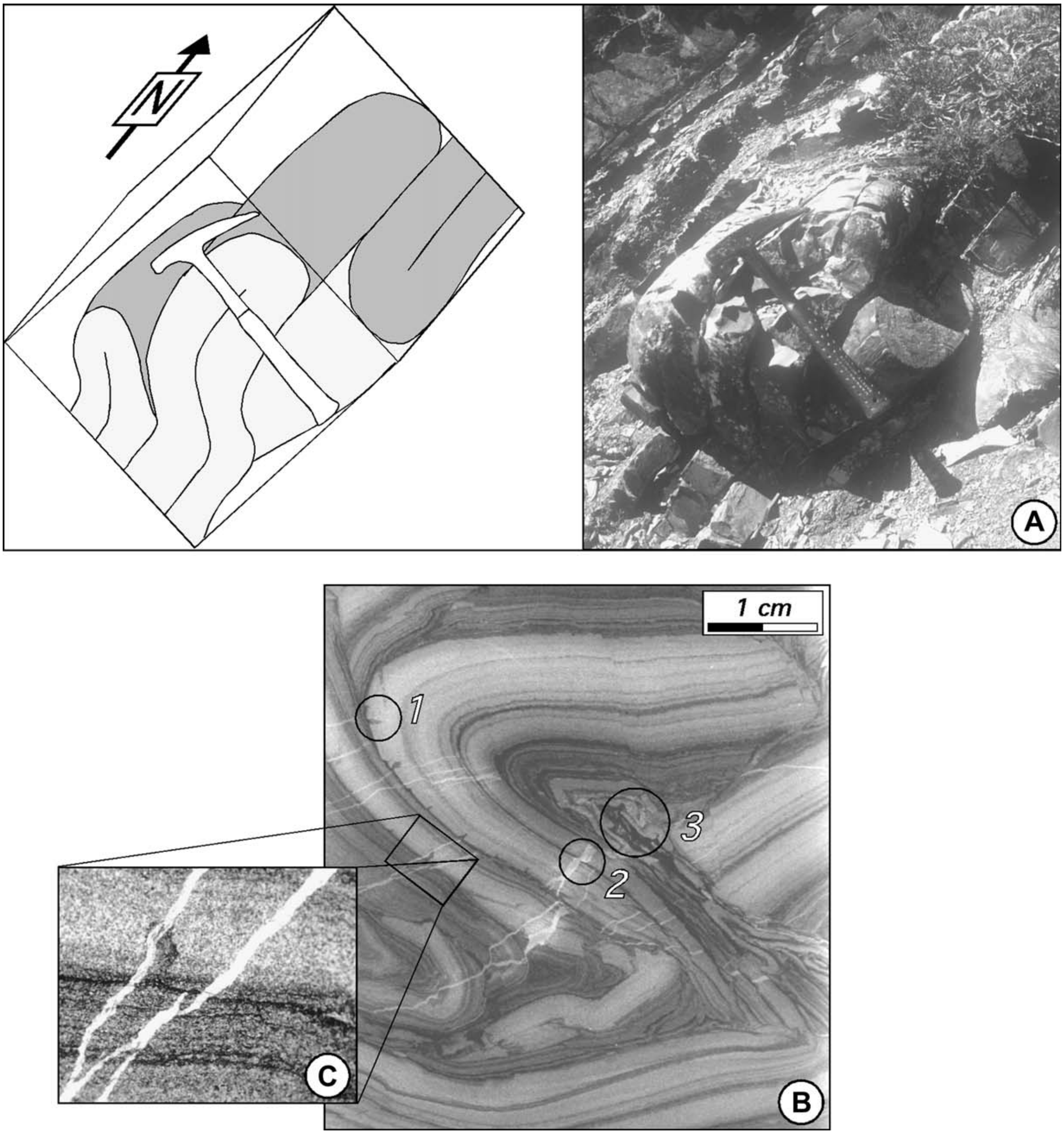

Fig. 3. Structural features of the F1 folds: (A) F1 folds developed in the Gorreto Sandstone, characterized by a strongly scattered axes and related block-diagram of the structure. A large scale F3 fold tilts the AP1 axial planes; (B) Photomicrograph in plane polarized light of a F1 fold. (1) Extension fissures opened at the fold outer limb and filled by siltstone matrix. (2) Veins filled by blocky calcite and slightly deformed by flexural-slip. (3) Injected hinge-zone; (C) Enlargement of area where the superposition relationships between calcite veins and extensional fissures are recognizable.

structures, particularly with the F3 folds (Fig. 3A). The F1 reversed limbs are strongly thinned and sheared, and commonly are cut by small-scale thrust-surfaces. In the high strain areas root-less F1 folds dispersed in the scaly shales are also recognizable. In spite of the high strain ratio, which characterizes the F1 folds and the related small-scale thrusts, these structures are not associated with any cleavage or cataclastic deformation. Only in a few cases in the Gorreto Sandstone, a weak axial-plane pressure solution, made up of thin insoluble mineral seams affecting the carbonatic framework, is recognizable.

Along the outer arc of the F1 sub-isoclinal folds a few microscopic extensional fissures are present. These fissures are filled by the fine-grained siltstone matrix deriving from the surrounding laminae (Fig. 3C). Microscopic injection dikes made up of shale and siltstone debris are also related to F1 folds.

Both the hinge zones and the reversed limbs are characterized by thin calcite vein-arrays, oriented at a high angle to the axial planes. These veins are mostly not folded, but in a few cases they are slightly dislocated by the flexural slip mechanism, indicating that their development is coeval only with the late stage of folding (location 2 of Fig. 3B). These veins are filled by blocky calcite and are characterized by irregular walls, which follow the sandstone grain-boundaries and never cut across them. Microscopic sandstone fragments dispersed in the infilling calcite are also present, giving the vein a "dirty" appearance (Fig. 4A). The extensional fissures filled by a fine-grained matrix are cut by the younger dirty calcite veins (Fig. 3C).

3.1.2. Boudinage

In the sandstone-rich facies at the base of the Montoggio Shale the F1 folds are associated with several extensional structures, for example high- or low-angle normal faults, symmetric and asymmetric boudins (sensu Needham, 1995) and book-shelf sliding structures.

The symmetric boudins are represented by meter-scale sandstone blocks, which preserve the original bed thickness and the internal sedimentary structures (laminations and bedding); the boudin terminations are sharp and oriented at a high angle to the bedding (Fig. 4B). These boudins are characterized by the presence of several veins oriented at a high-angle to the bedding and filled by blocky calcite.

The asymmetric boudins can be represented by elements with an irregular shape (Fig. 4C–E) or by blocks bounded

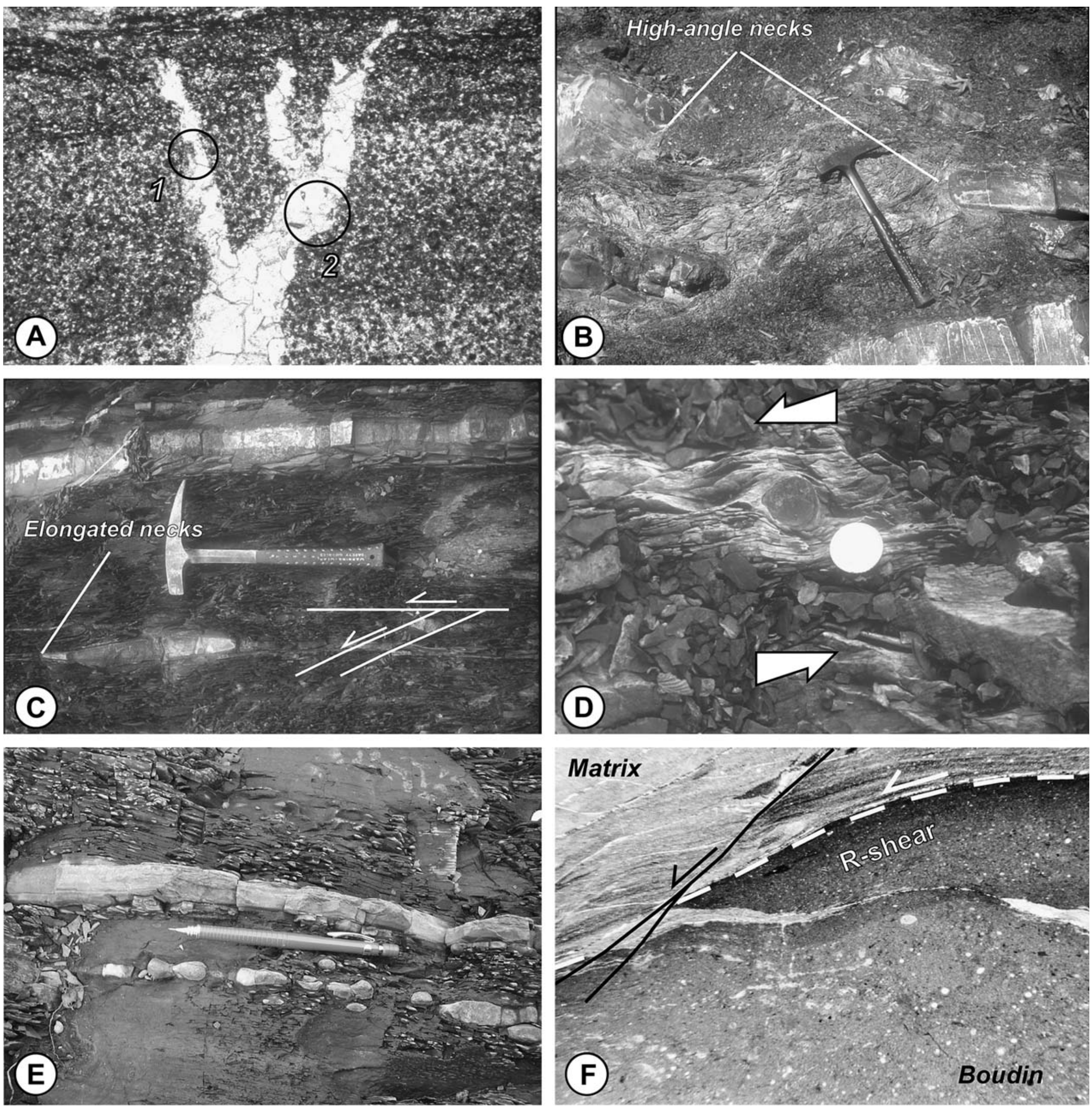

Fig. 4. (A) Photomicrograph in plane polarized light of D1 vein filled by "dirty" blocky-calcite, associated with an F1 fold; (1) irregular vein-walls following the grain boundaries; (2) vein-wall particles scattered in the blocky-calcite; (B) boudinaged sandstone beds in the lower part of the Montoggio Shale; the boudins are symmetric and preserve their internal sedimentary structures (lamination); (C) mesoscale asymmetric boudins developed in the Montoggio Shale, bounded by R-shears (to the right) or showing strongly elongated necks (to the left); the latter do not preserve the internal primary structures; (D) σ-object in the Montoggio Shale dispersed in the scaly matrix; (E) irregular boudins showing pinch and swell structures developed in the Montoggio Shale; (F) microphotograph in plane polarized light of the necking/matrix contact of an asymmetric boudin preserving the internal sedimentary structures. The boudin is bounded by R-shears, with minor synthetic shears.

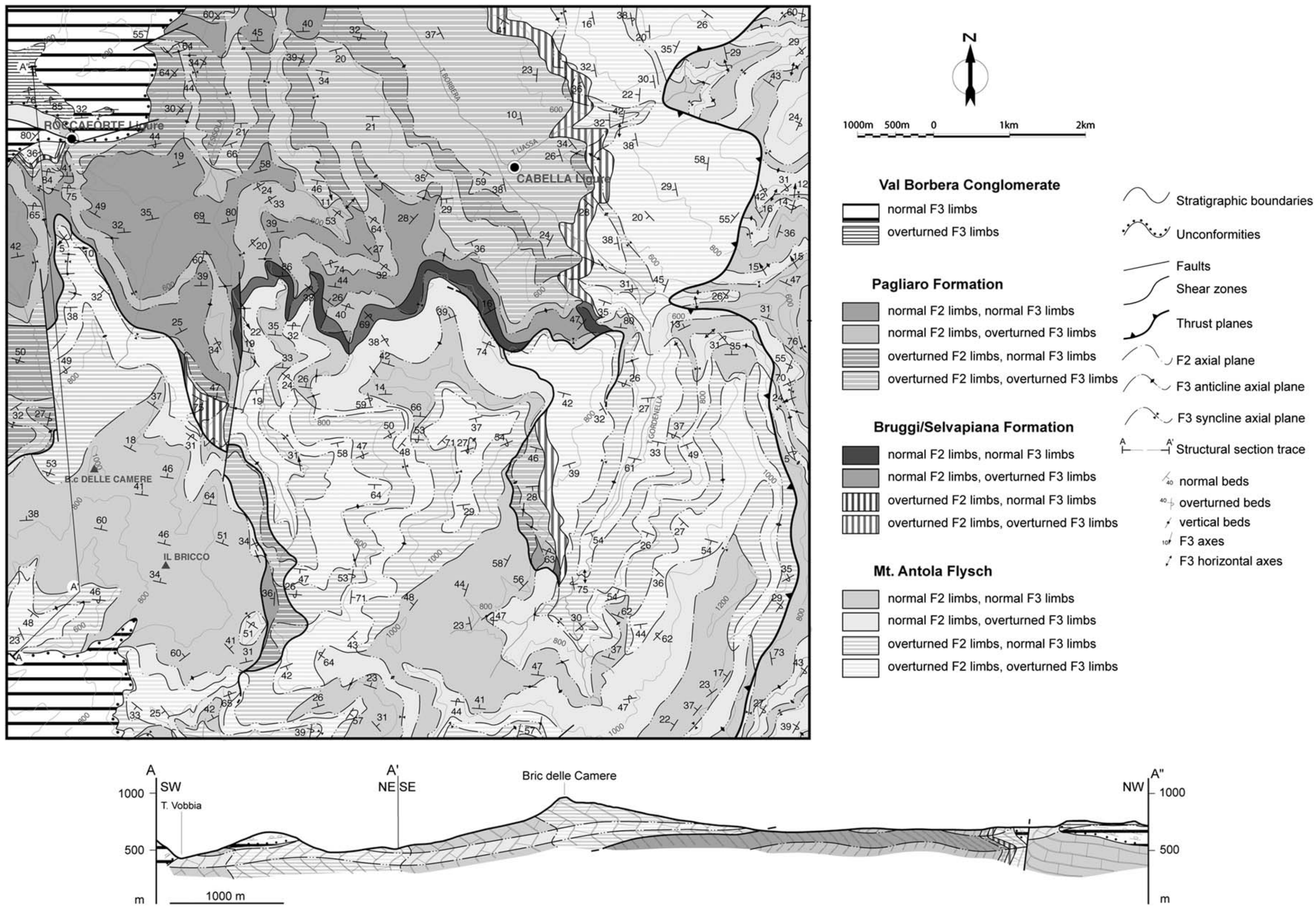

Fig. 5. Structural-geological map and related geological section of the NW outcropping sector of the Antola Unit where the contact with the unconformable deposits of the Tertiary Piedmont Basin is preserved. This map highlights the interference between large-scale F2 and F3 folds. F2 folds are represented by sub-isoclinal folds juxtaposing normal and overturned limbs showing the same orientation.

Fig. 6. Mesoscale F2 syncline developed in the Mt. Antola Flysch. The overturning direction indicates an E-vergence.

by faults oriented at a low angle with respect to the bedding and the scaly fabric (Fig. 4C–F). The boudins with irregular shape do not preserve internal sedimentary structures (bedding or lamination) and usually are characterized by strongly elongated necks affected by pinch and swell structures (Fig. 4C-E). In this kind of boudin no vein development has been observed. Small rounded shale-blocks scattered in the scaly matrix are also present; the surrounding shales are arranged in a σ shape with tails in which no crystal growth has been recognized but only an arrangement of the pre-existing shaly matrix (Fig. 4D).

The asymmetric boudins bounded by low-angle faults preserve the original bed thickness and the primary internal sedimentary structures. The low-angle faults can be regarded as R-shears and along them no mineral fibers have been recognized (Fig. 4C–F).

The structural relationships between R-shear systems and F1 folds are difficult to constrain, because of the strong dispersion due to the post-D1 deformations (in particular D2 and D3 folding). Nevertheless, in the Montoggio area the F1 structural elements, preserved in a vertical limb of a large-scale F3 fold, display coherent sense of shear, which real facing need the restoration of the post-D1 deformations. For a complete discussion of this point see the Cap. 4.

3.2. The D2 phase deformations

The D2 phase is represented by tight to sub-isoclinal strongly non-cylindrical F2 folds with approximately parallel geometry (class 1B of Ramsay, 1967) associated with a weak axial plane foliation S2 and with the development of extensional tectonic contacts.

The outcropping hinge-zones are scarce, because of the common development along the AP2 axial plane of cataclastic shear zones. However, the presence of F2 sub-isoclinal folds is recognizable at the map-scale by the occurrence of overturned limbs showing the same direction and plunge as the normal ones (Fig. 5).

The A2 axes are strongly scattered, as a result of both the non-cylindrical geometry of the F2 folds and the post-D2 deformations. Restoring the original E–W trend of the post-D2 deformations (A3 axes) the geometry of the F2 folds indicates a primary prevailing NE facing (Fig. 6).

The S2 axial-plane foliation is a disjunctive cleavage developed only in the F2 hinge zones; it is well developed in the pelitic and arenitic beds, but it is rare in the calcareous marls. In the arenitic beds the S2 foliation is evidenced by aligned elongated grains (lithic fragments or mica grains) and seams of insoluble materials, such as heavy minerals or phyllosilicates.

The F2 folds, particularly the F2 synclines, are characterized by two extension vein systems, mostly developed in the fine-grained arenitic laminae at the base of the turbiditic beds. The first system is oriented in a fan-shape around the AP2 axial planes, while the second is characterized by veins perpendicular to the fold axes. Both these vein systems cut the arenitic grains and are filled by calcite fibers perpendicular to the vein-walls. The fibers are antitaxial and perpendicular to the vein-walls, and the presence of inclusion trails is indicative of a crack-seal growing mechanism.

The D2 phase is characterized by the development of tectonic contacts affecting the whole thickness of the AU. The D2 tectonic contacts lie at low angle to the bedding and to the AP2 axial planes. A map-scale D2-related tectonic surface can be recognized in the Montoggio area (Fig. 7). In the SW sector (location 1 of Fig. 7) this contact is a gentle SW-dipping surface juxtaposing a hangingwall consisting of the Mt. Antola Flysch over a footwall made up of the Varicoloured Shale (upper part of the Montoggio Shale). The related minor faults, regarded as R-shears, indicate a top to E sense of shear (Fig. 8). The eastern sector of the Montoggio area is characterized by a D2 contact lying at a high angle, being deformed by a successive F3 overturned fold (location 2 of Fig. 7). Here, in the hanging wall of the D2 contact, the stratigraphic relationships between the whole basal complex and the overlying Helminthoid Flysch are preserved. In the studied area large-scale F2 structures are not recognizable, but only rare mesoscale isoclinal folds. In spite of the possible mistakes in the interpretation of tectonic contacts starting from finite geometries (Wheeler and Butler, 1994), in the studied case, assuming that the large-scale F2 are lacking and that at the end of the D1 phase the main part of the AU succession was almost subhorizontal, after the restoration of the post-D2 deformations

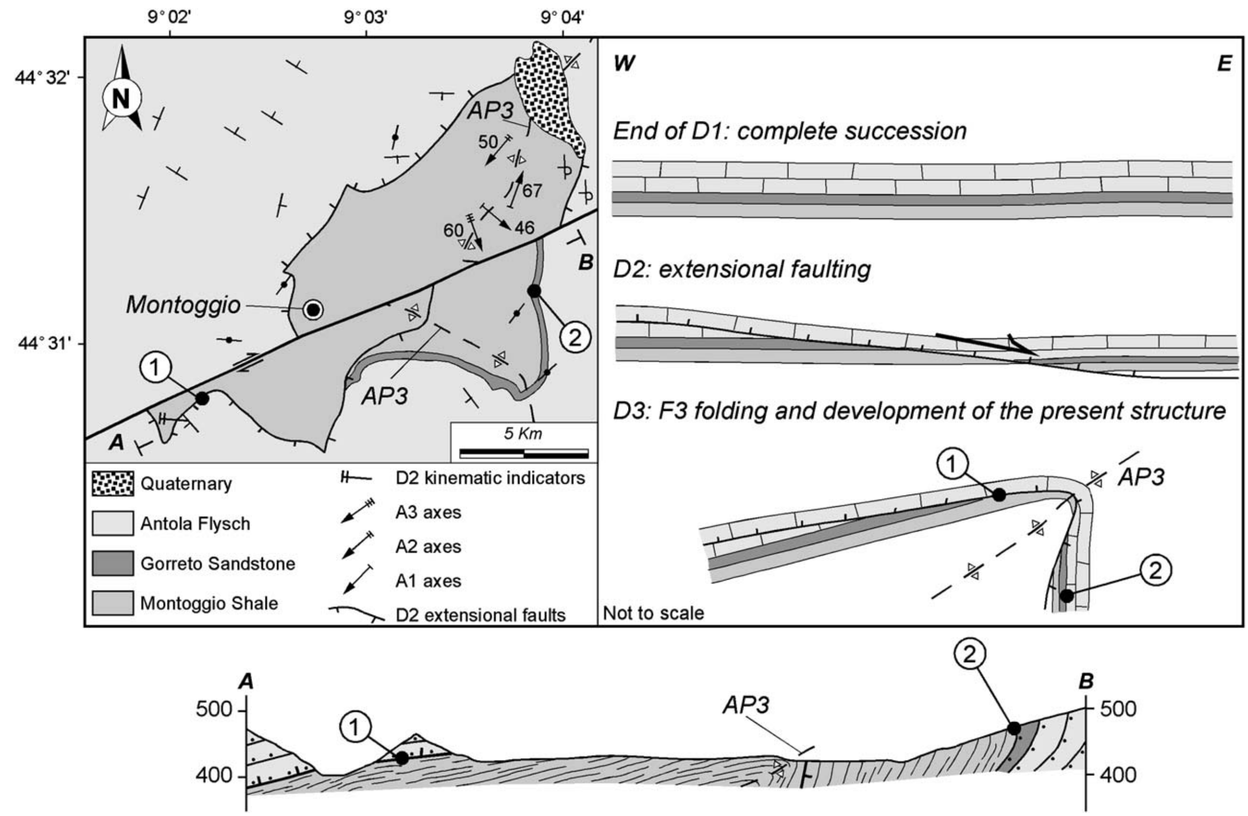

Fig. 7. Montoggio area structural map with geological cross-section, and interpretative evolution of the observed structure (the small-scale F2 and F3 folds have been neglected).

(in particular the D3 folds) the described tectonic contact can be interpreted as an E-facing low-angle normal fault (Fig. 7).

3.3. The D3 and D4 phase deformations

The D3 phase structures are the most widespread and best developed all around the study area (Fig. 5). The F3 folds are open to tight overturned asymmetric folds, with approximately parallel geometry (classes 1b, 1c and 2 of Ramsay, 1967). They formed by a flexural-slip folding mechanism, testified by widespread axes-perpendicular slickensides and calcite fibers recognized on the fold limbs.

The hinge-zones of F3 folds are characterized by the development of a spaced S3 disjunctive cleavage, refracted at the

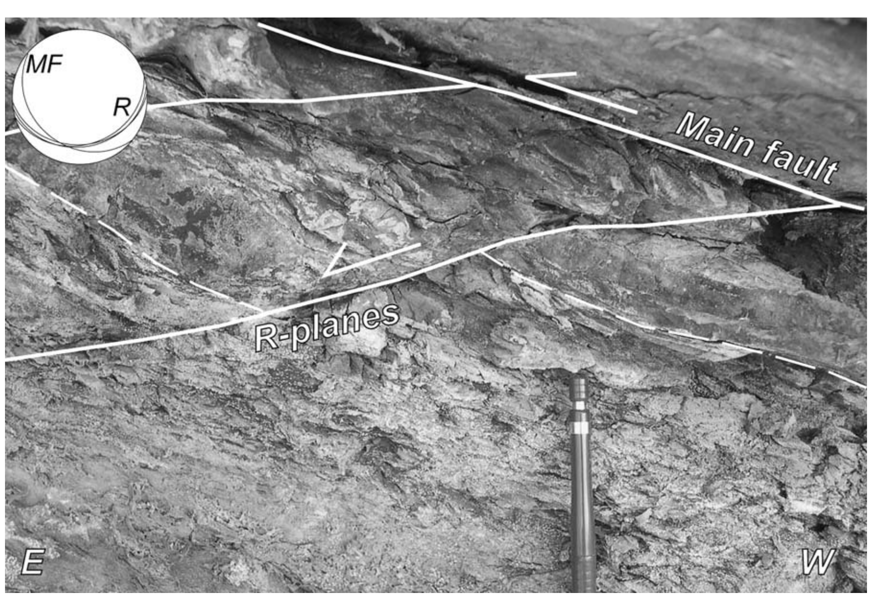

Fig. 8. Particulars of a D2 fault, corresponding to location 1 of Fig. 7, and a stereoplot of the main fault and related R-shears. The hangingwall is represented by the Mt. Antola Flysch, while the footwall correspond to the Montoggio Shale.

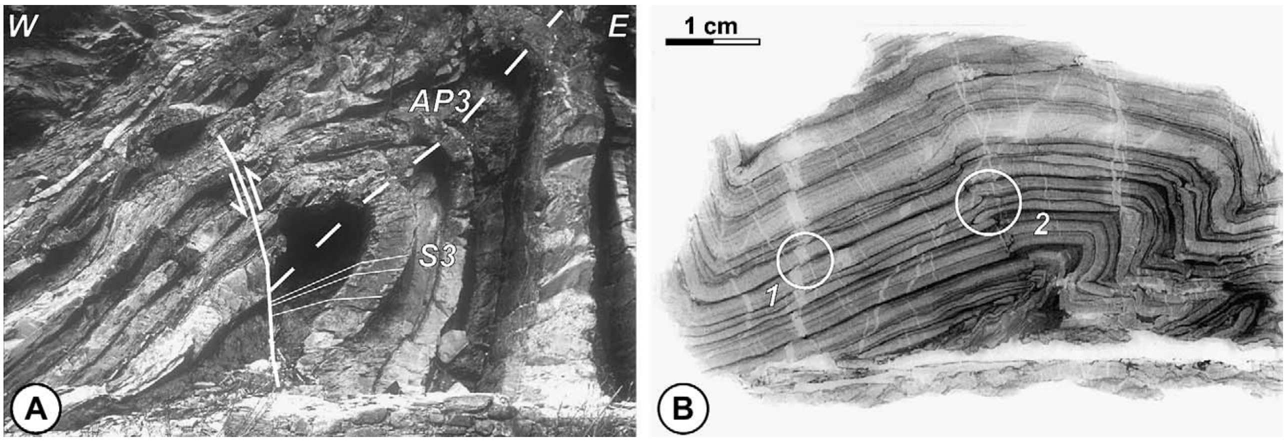

Fig. 9. F3 folds. (A) Mesoscale F3 fold, with sub-horizontal A3 axes, West-dipping AP3 axial plane and fan-shaped S3 foliation, indicating a vergence toward East; (B) photomicrograph in plane polarized light of a micro-F3 fold, showing the flod-microthrust relationships between fold and micro-thrust. 1, V3a veins dislocated by the flexural slip mechanism related to the fold-development; 2, minor back-thrusts.

shale/limestone interface (Fig. 9A). At the microscopic scale in the arenitic beds this spaced foliation S3 is represented by the alignment of elongated grains (detrital quartz, feldspar or phyllosilicates), often affected by an intense cataclastic deformation and by dark seams of insoluble materials, mostly oxides and detritic phyllosilicates.

The A3 axis trend is strongly variable, ranging from E–W to N–S (Fig. 10). The AP3 axial planes are slightly plunging in the NW sector of the AU and more steep in the SE one (up to 45°), and plunge toward S or W, depending on the orientation of the A3 axes. The resulting F3 folds face toward N or E.

The F3 folds are characterized by the presence of two synfolding extension vein systems, developed mostly in the thin arenitic laminae at the base of the turbiditic beds. The V3a veins are characterized by a fan-shaped orientation around the AP3 axial planes, and are filled by calcite fibers growing with an A3-normal orientation. The V3b veins are characterized by an A3-normal orientation and by fibers growing parallel to the A3 axes. Both the extension vein generations cut through the detrital grains of the arenitic beds.

The F3 folds are associated with different scale thrust surfaces (Figs. 5 and 9), commonly characterized by widespread brittle shear zones. The sense of shear deduced from the structural features of the thrusts and the cataclastic bands is coherent with the local facing of the F3 folds, toward N or E depending on the orientation of the A3 axes.

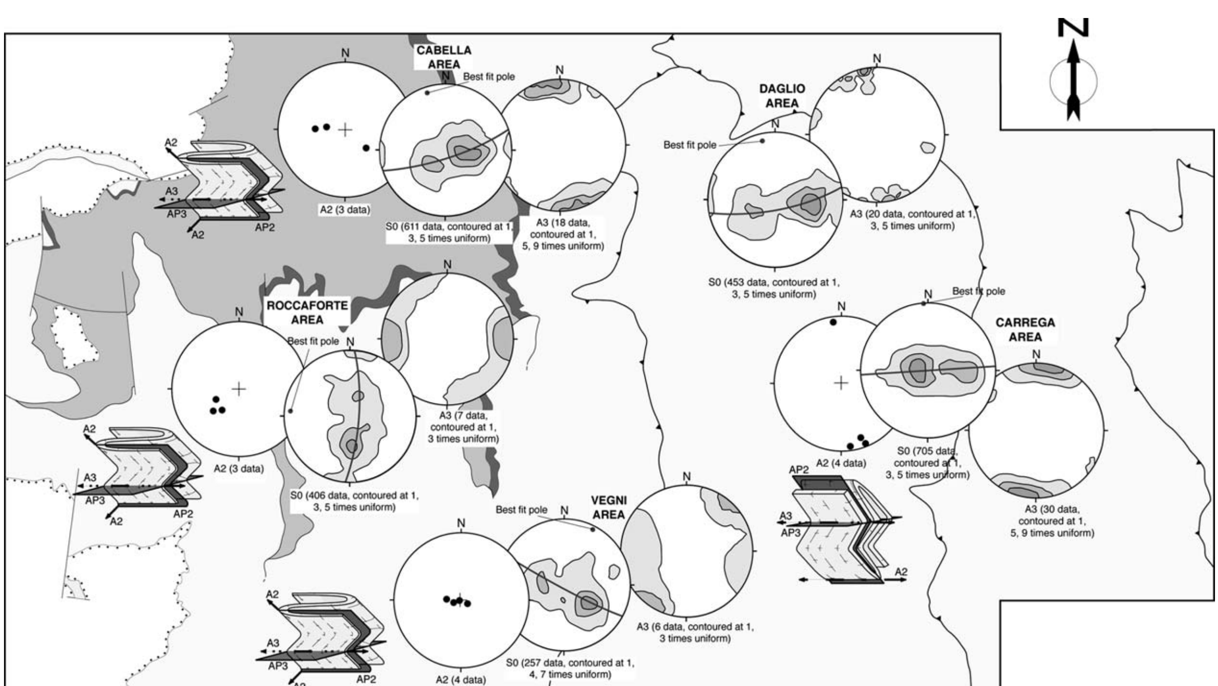

Fig. 10. Variation of the F2/F3 interference pattern in a large sector of the Antola Unit and its relationship to the orientation of the A3 axes, obtained from the stereoplots of the A2 axes, A3 axes and S0 planes (Schmidt net, lower hemisphere).

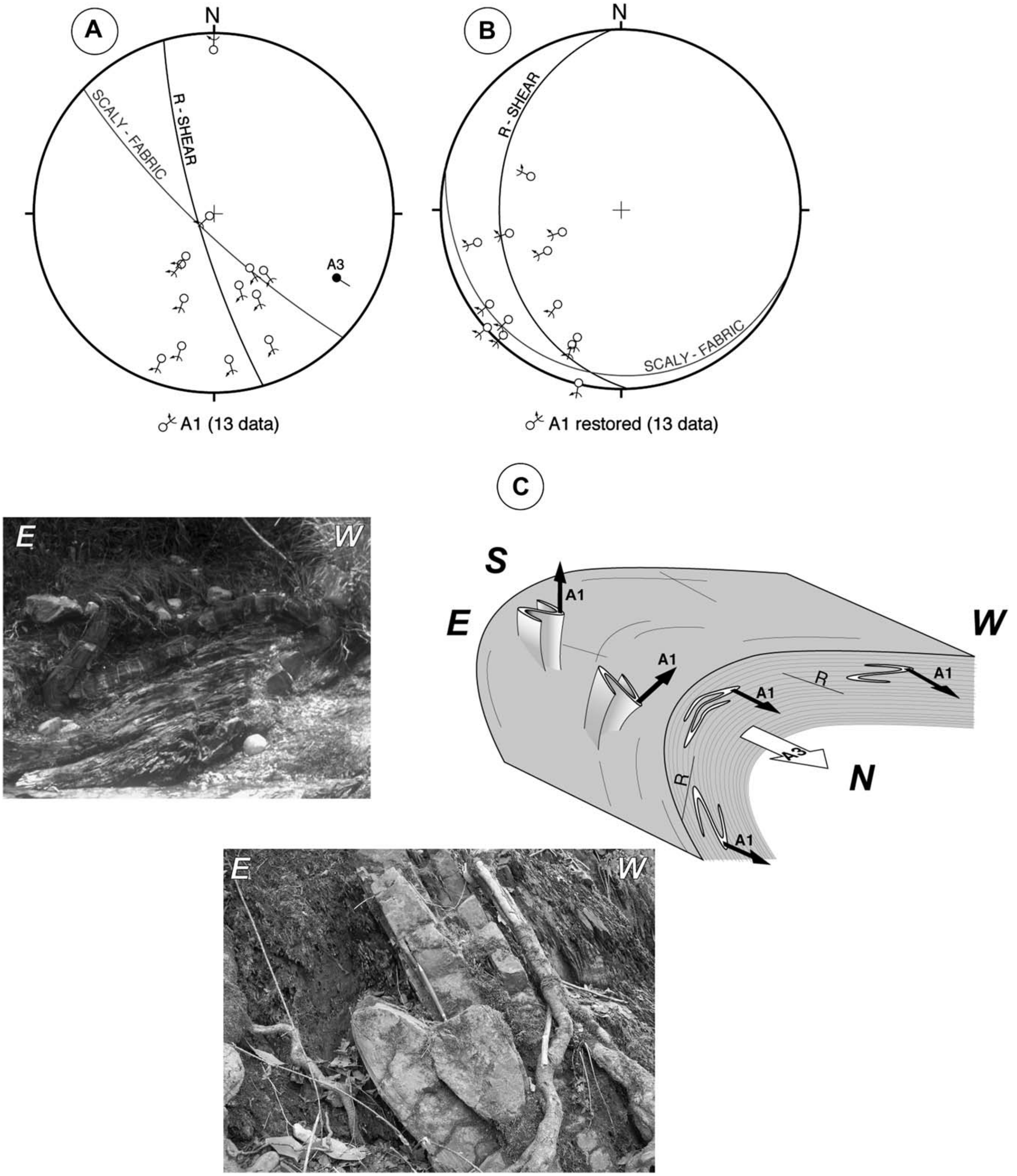

Fig. 11. Reconstruction of the D1 phase vergence starting from the data collected in a large-scale F3 fold in the Montoggio area. (A) Stereoplot of the scaly-fabric and related R-shears, A1 collected in F3 vertical limb and A3 axis; (B) stereoplot of the D1 structural elements (same data presented in (A)) after the restoration of the large-scale F3 fold around the A3 axis, bringing back the bedding to the horizontal. The restored orientation of the D1 structural elements indicates a prevailing vergence toward the western sectors; (C) 3D block-diagram of the F1/F3 interference and the deriving geometries of the mesoscale F1 folds preserved in the different structural domains of the large-scale F3 fold.

The superimposition of F3 and F2 folds produces variable interference patterns, which change from type 2 to type 3 of Ramsay (1967). This variation has been observed in both the N–S and E–W oriented A3 domains and can be considered independent from the dispersion of the D3 structures (Fig. 10).

The D3 phase related folds have also been recognized in the lower middle Eocene-Oligocene part of the unconformable TPB (Fig. 5).

Slight open folds with vertical axial planes and sub-horizontal N–S oriented axes characterize the D4 deformations. The hinge zone of the F4 folds is characterized by a diffusive brittle deformation, testified by cataclastic deformation and large sub-vertical veins filled by blocky calcite. Structures referable to the D4 phase have been recognized in the TPB deposits.

4. Discussion

In spite of the numerous observed structures, the vergence of the D1 phase is difficult to constrain, because of the superposition of two folding phases (D2 and D3) both associated with large rotations and the non-cylindrical geometry of the F1 folds. The facing of these structures has been obtained restoring the post-D1 structures in areas characterized by pinpointed structural setting. For example, in the Montoggio area (Fig. 7), the D1 structural elements are preserved in a vertical F3 limb, where no D2 structures are recognizable. The collected D1 data have been restored to the primary orientation, indicative of a rough vergence toward the Western sectors (Fig. 11).

Some further deductions about the vergence of the D1 phase can be made from palaeogeographic constraints. The paleogeographic reconstruction proposed by Elter and Pertusati (1973) and followed by others (Marroni et al., 2001; Cerrina Feroni et al., 2002; Catanzariti et al., in press) indicate that the AU succession can be referred to the outermost part of the Adria plate margin, and therefore in a position located between the Ligurian-Piedmontese oceanic domain and the Eastern EL Units. The tectonic units deriving from the Ligurian-Piedmontese oceanic domain, the IL Units, underlie the AU. This structure derives from the first main emplacement phase of the AU with a Westward "Alpine" vergence.

4.1. D1 phase deformation mechanism

The development of a scaly fabric in shale deposits has been referred to the deformation of unlithified sediments (Kleist, 1974; Knipe, 1986, 1989; Moore et al., 1986; Maltman, 1988; Labaume et al., 1990; Vannucchi et al., 2003). In this environment the most likely deformation mechanism is independent particulate flow, where the strain takes place through sliding along the grain boundaries (Maltman, 1984; Knipe, 1986, 1989). The presence of extensional fissures filled by fine-grained matrix indicates that during the D1 deformation the muddy sediments were still unconsolidated and able to flow. Moreover, the irregular vein-walls, the grain boundaries and the dirty appearance of the blocky calcite are characteristic of veins developed in unlithified rocks. The absence of grain-cataclasis and foliation, in spite of the high strain ratio related to the sub-isoclinal F1 folds, is consistent with the hypothesis of independent particulate flow developed even in the arenitic beds. The last mechanism is typical of unconsolidated sediments characterized by high fluid-pressure, which in the AU basal complex is testified by the occurrence of micro-scale injection dikes. As discussed by Needham (1995) in the Shimanto Belt, both irregular and symmetric boudins coexist, deriving from the deformation of partially lithified sediments. The irregular boudins are characterized by strongly swelled necks and do not preserve their internal sedimentary structures. These features are regarded as deriving from independent particulate flow, causing the obliteration of the internal sedimentary structure. By contrast, symmetric boudins preserving their sedimentary structures derive from more lithified strata. These conclusions can be applied to the analogous structures recognized in the AU basal complex.

The superposition relationships indicate that the F1 folds mainly developed during the first stage of the D1 phase, while the dirty veins characterize the latest stage. According to Moore and Byrne (1987) this change in deformation style can result from the progressive hardening of unconsolidated sediments, which during the D1 deformation decreases their porosity by fabric collapse. It is still unclear if the development of asymmetric and symmetric boudins is a coeval process related to differently lithified beds, as proposed by Needham (1995) for the Shimanto Belt, or if they represent the result of the different lithification rate reached by the arenitic layers during the different deformation stages which seem to characterize the D1 phase, as suggested by the folds–veins relationships.

The presence of the scaly fabric in the shales is indicative of a predominant shear deformation, and the presence of σ-objects, rounded blocks and strongly asymmetric structures strongly supports this hypothesis. In this case the coexistence of both compressive and extensional structures, as sub-isoclinal folds and boudinage, can be regarded as the result of the progressive deformation connected with this shear deformation.

The structural style of the AU basal complex is similar to that characterizing the base of other Helminthoid Flysch Units of the Northern Apennines, where a soft sediment deformation has been recognized (Bettelli and Vannucchi, 2003).

4.2. D2 phase deformation mechanism

The D2 structures mark a sharp change in the structural style with respect to the D1 phase. In fact, the presence of extension veins cracking through the grains of the arenitic beds and the widespread cataclastic deformation indicate brittle deformation in already lithified host rocks. The prevailing cataclastic flow deformation mechanism is associated with a minor diffusive mass transfer, testified by the common pressure solution foliation recognizable in F2 folds.

The coexistence of two apparently contrasting features, sub-isoclinal folding and extensional faulting, has been described in several tectonic settings (Malavieille, 1987; Carmignani and Kligfield, 1990; Froitzheim, 1992; Orozco et al., 1998, 2004; Harris et al., 2002). In the Alpujarride (Betic Cordillera, Spain) large recumbent folds with rounded hinges are closely associated with extensional faults oriented at a low-angle to the axial planes. Both folds and faults are the result of a gravity driven extensional regime, related to the belt collapse (Orozco et al., 1998, 2004). The D2 structures recognized in the AU are analogous, particularly in the geometrical relationships between isoclinal folds and extensional tectonic contacts, and the axis dispersion. These resemblances support the hypothesis that the D2 phase of the AU could be the result of a gravity driven mechanism. Given the shallow deformation level of the AU, this gravitational mechanism is probably a combination of ductile gliding and spreading. As discussed by Merle (1998), if a tectonic unit is able to deform by ductile gliding, it will able to be deformed by gravitational spreading too; thus, the two mechanisms are coexisting. In this case the F2 folds can be regarded as passive folds produced by the irregularity in the sliding surfaces corresponding to the extensional normal faults. The strongly non-cylindrical geometry of the F2 folds of the AU can be regarded as the result of the amplification of original perturbations, induced by irregularities in the sliding surfaces (Brun and Merle, 1988; Merle, 1998). Similar variations in the trend of passive fold axes in relation to gravity-driven mechanism have been recognized in the Parpaillon Unit (Helminthoid Units, Western Alps) (Merle, 1984).

4.3. D3 and D4 phase deformation mechanism

As for the D2 structures, the main D3 deformation mechanism is the cataclastic flow, with minor diffusive mass transfer testified by pressure solution of the carbonate framework.

The structural features of the D3 phase, characterized by the systematic association between folds and thrusts, are compatible with a typical fold and thrust belt. The dispersion of the A3 fold axes can be explained by a post D3 torsion due to a subsequent transcurrent deformation phase (Marroni et al., 2002a).

The D4 phase is difficult to constrain. The presence of F4 open folds with sub-vertical axial planes is the result of compressive tectonics, possibly related to a transcurrent regime (similar folds have been produced in analogue models by Odonne and Costa, 1993). North of the study area runs a main regional transcurrent fault, the Villavernia-Varzi line

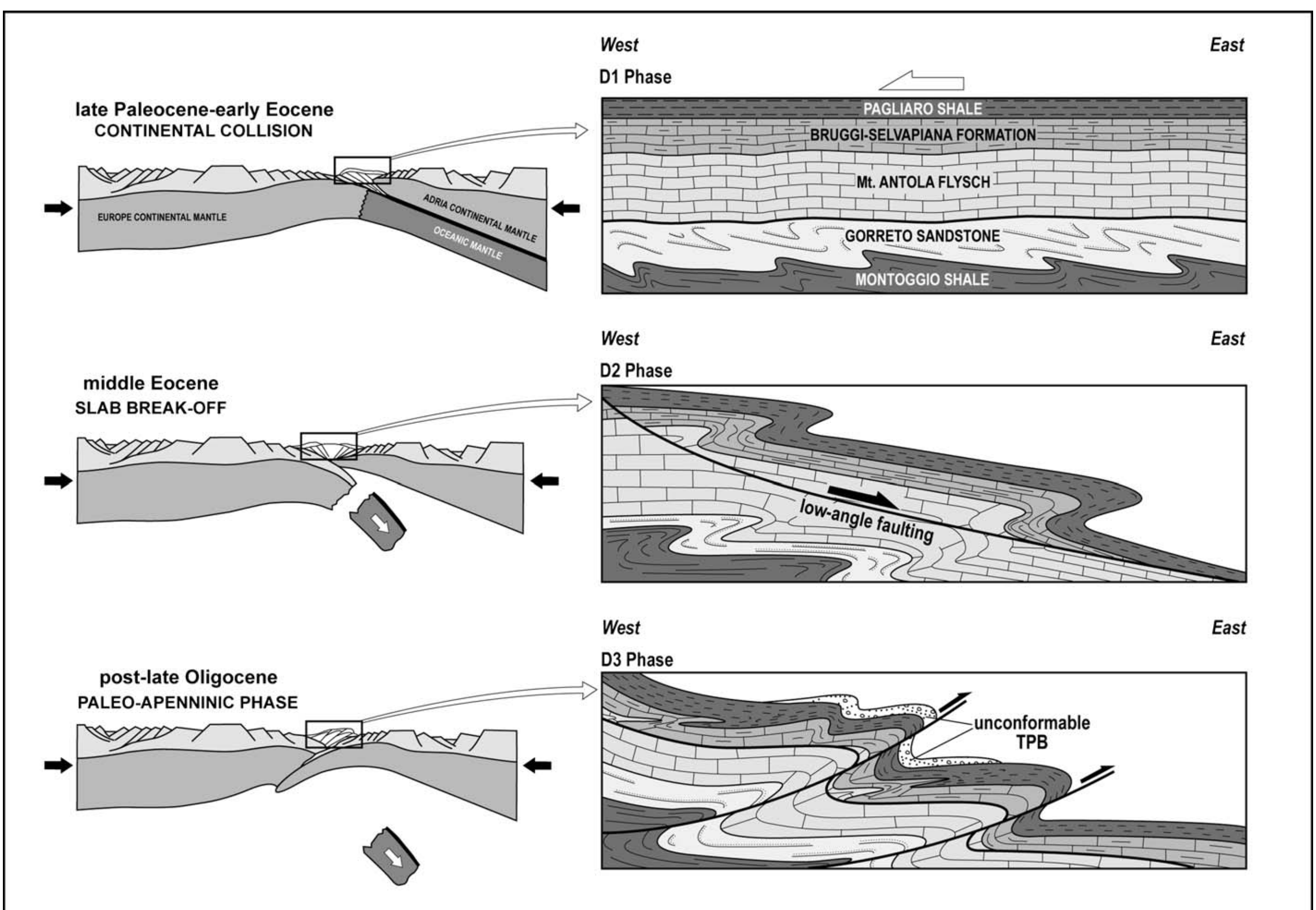

Fig. 12. Synthesis of the AU deformation history and relationships with the Alps-Apennines geodynamic setting.

(Fig. 1), activated probably during the late Oligocene (Schumacher and Laubscher, 1996). The development of the D4 structures can be tentatively related to this regional structure.

4.4. Geodynamic setting of the Antola Unit deformations

The most peculiar feature of the AU structural evolution is the strongly heterogeneous distribution of the deformation in the whole thickness of the unit, characterized by high strain levels at the base (basal complex, i.e. Montoggio Shale and Gorreto Sandstone) and very-low strain in the upper part (Mt. Antola Flysch, Bruggi-Selvapiana Formation and Pagliaro Shale). This anomalous distribution could indicate that the D1 deformation predates the deposition of the whole calcareous flysch succession. Nevertheless this hypothesis is invalidated by the occurrence of a continuous biostratigraphic record through the whole AU succession (Catanzariti et al., in press). This heterogeneous deformation distribution is possibly related to the high contrast between the lithification rate of a shaly-dominated basal complex and the overlying calcareous succession during the D1 deformation. In fact in an underwater environment the diagenesis process is a combination of mechanical compaction and chemical reactions; in the case of the shales the last ones are not time-dependent and start with temperatures higher than 70–80 °C, corresponding to a few kilometers depth (Hower et al., 1976; Bruce, 1984), while for the fine-grained carbonates the cementation can be a process closely following the deposition (Lasemi et al., 1990). Consequently, during the D1 phase the shales could be still unconsolidated, while the calcareous flysch succession was already lithified.

The D1 structural features indicate a prevailing simple-shear regime, which can be related to the translation of a rigid sheet (corresponding to the already lithified calcareous succession) over a relatively thin soft level, consisting of the mostly unconsolidated basal complex (Fig. 12). This mechanism can explain the absence of deformations in the upper part of the AU even with a long-distance translation with an Alpine-vergence. Thus, the D1 phase can be related to the AU emplacement at the top of the Ligurian stack, representing the shallower part of a West-verging accretionary wedge (Fig. 12). The youngest deposits of the AU succession are referred to the beginning of late Paleocene (NP5 zone, Catanzariti et al., in press), while the oldest deposits of the unconformable overlaying TPB are from the upper part of the mid-Eocene. The emplacement of the AU must be related to the oldest part of this time-span, corresponding to the Meso-Alpine phase, which records the beginning of the continental collision as testified by the subduction of the European plate margin beneath the Alpine accretionary wedge (Rubatto and Hermann, 2001; Meffan Main et al., 2004). In agreement with the occurrence of an unconsolidated level at the AU base, the stratigraphic data indicate that until the Eocene, the Alpine belt was essentially a submarine accretionary wedge. The first deformation of the Helminthoid Flysch domain has been attributed to the first stages of continental collision, causing the readjustment of the stress field at the belt-scale involving the outermost part of the Adria plate margin.

The subsequent D2 phase marks a sharp variation in the structural style due to a change in the geodynamic setting and in the rheological characteristics of the involved rocks. The TPB deposits seal the D2 structures, so that this phase can also be referred to the Meso-Alpine tectonics, related to the occurrence of a West-verging accretionary wedge. As discussed above, the D2 structures suggest a gravity-driven mechanism, characterized by the close association of E-verging sub-isoclinal folds and low-angle normal faults (Fig. 12). This phase can be correlated with the Meso-Alpine phase of Marroni et al. (2002b), which has been regarded as a gravitational process that results from the uplift of the orogenic wedge induced by the break-off of the subducting slab.

As discussed by Marroni et al. (2002a), the youngest TPB deposits involved in the D3 structures are late Oligocene, and thus can be referred to the NE-verging Apenninic tectonics (Fig. 12). This remark agrees with the documented vergence of the D3 structures and with the occurrence of fold and thrust structures. This post late Oligocene N-facing phase has been correlated by Marroni et al. (2002a) with the thrusting of the metamorphic Alpine Units, represented by the Voltri Group, over the Northern Apennines during the so-called Paleo-Apenninic phase (Piana and Polino, 1995; Schumacher and Laubscher, 1996).

5. Conclusions

Despite its shallow structural level, the AU records a complex deformation history, referable to two superposed orogenic cycles (Alpine and Apenninic). The first part of this deformation history (D1 and D2 phases) is referable to the Meso-Alpine tectonics due to the beginning of the continental collision between the Adria and Europe plates. The second one (D3 and D4 phases) is post-late Oligocene and is referable to the development of the Northern Apennines collisional belt. The two orogenic cycles are separated by at least 15 My, during which the sedimentation of the lower part of the TPB succession took place.

The late Paleocene-lower Eocene D1 structures are developed only in the AU basal complex and are characterized by a marked soft-sediment style and shear strain regime. These structures are referable to the West-ward translation of the whole AU over the Ligurian orogenic wedge, gliding on the poorly lithified shaly basal complex. The middle Eocene D2 phase, affecting the whole thickness of the AU is characterized by NE-verging sub-isoclinal folding and low-angle normal faulting. This deformation phase is referable to a gravity-driven mechanism, probably due to the exhumation of the deeper part of the orogenic wedge. Both the D1 and D2 structures are sealed by the TPB deposits.

The late D3 and D4 phases affected both the AU and the overlying TPB deposits. Particularly the post-late Oligocene D3 phase is characterized by NE-facing folding and thrusting, and is correlatable with the Paleo-Apenninic phases.

The results from the mesoscale structural analysis of the AU, connected with the successive geodynamic settings in the Alps-Apennine system, can therefore provide a useful contribution for a better understanding of the evolution of high structural level units, which suffered a polyorogenic deformation history.

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