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Fluid flow during accretion in sediment-dominated margins: Evidence of a high-permeability fossil fault zone from the Internal Ligurian accretionary units of the Northern Apennines, Italy

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Abstract

We report here a detailed structural study carried out in the Internal Ligurian Units of the Northern Apennines, Italy, formed during the building of the Alpine accretionary complex through subduction of the sediment-filled Ligure–Piemontese oceanic basin. The deformation mechanisms associated with fluid migration across an accretion-related fault zone have been studied through a detailed analysis of different generations of syn-tectonic veins. Hydrofracturing occurred mainly sub-parallel to bedding in unlithified to semi-lithified sediments. Transient, upward-directed fluid injection locally connected the décollement-parallel veins through bedding-normal hydrofractures of lithified sandstone layers. A third vein system comprises fibrous hydrofractures developed on the limbs of accretion-related folds. Crosscutting vein sets and the peculiar features of each identified vein set suggest that deformation was intricately associated with lithification and diagenetic processes. Dehydration-produced fluids transiently injected the lithifying sediments leading to local stress permutations. The proposed model provides a "ramp-flat" migration of fluids in which fluid flow is enhanced along high permeability, less cohesive layers, leading to the development of regional dilated hydrofracture channels like those recognized along the décollement zone of modern margins. The more competent layers are truncated by high angle fractures representing the transient connectivity that existed between horizontal conduits.

Keywords: Northern Apennines; Internal Ligurian Units; Syn-tectonic veins; Décollement zone; Fluid overpressure

1. Introduction

Much of what we know about the evolution of deformation mechanisms and fluid pathways during accretion comes from insights gained from the Ocean Drilling Program, that in the past two decades has monitored and imaged convergent margins with geophysical techniques, laboratory analyses and in situ measurements (for a general review see MARGINS, 2004). These data demonstrate that the décollement zone, the major basal shear zone and plate boundary of subduction margins, is a preferential pathway for escaping fluids generated by dehydration and melting reactions (Moore and Mascle, 1987; Kastner et al., 1991; Moore and Vrolijk, 1992; Bangs et al., 1999; Silver et al., 2000; Moore and Silver, 2002; Maltman and Vannucchi, 2004).

Fluids along convergent margins control heat and mass transport, magmatism, mineral transformations and mantle alteration processes (Ge et al., 2003). Moreover, the fluid escape to the surface from the downgoing plate is intimately linked to deformation and faulting, and to the evolution of the décollement zone. Overpressured fluids in accretionary thrusts help to maintain the critical wedge taper (Davis et al., 1983) and control sediment deformation, mechanical

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sediments properties and fault localization, finally determining the location, frequency and magnitude of seismic ruptures (Sibson, 1990, 1992; Tarney et al., 1991; Scholz, 2002).

As elegantly summarized by Maltman and Vannucchi (2004), horizontal flow generally dominates over vertical flow, as fluids are channeled along the décollement mega-shear zones, which tend to act as hydrogeological seals, isolating the underthrusting sediment section. Moreover, lateral fluid flow along the décollement zone is highly heterogeneous and episodic: fluids are distributed in pockets and lenses characterized either by remnants of highly porous sediments or overpressure and dilation (Bangs et al., 1999). Experimental modeling and geological observations have demonstrated the dynamic character of permeability in accretionary prisms, suggesting feedback between fluid pressure, fluid pathways and dehydration reactions in sediments that can strongly control subduction zone seismicity (Sibson, 1992; Miller et al., 2003).

Despite these efforts, the complex interplay between dewatering, fluid flow and deformation, remains insufficiently characterized. Previous studies have demonstrated that there is a wide structural diversity and heterogeneous fluid distribution along the same margin both in space and time, and, spatially, both along and across the deformation front (von Huene, 1984; Le Pichon and Kobayashi, 1992; Shipley et al., 1995; Maltman et al., 1997; Moore and Klaus, 1998). In particular, it has been suggested that in modern accretionary prisms sediments experience potentially cyclic fluctuation of both tectonic stresses and pore-fluid pressures.

While the direct measurement and monitoring of active processes represents a valuable source of information in the interpretation of accretionary complexes, studies of ancient, fossil structures provide key insights into the structural geometries and mechanisms of deformation, thus facilitating the testing of models built in modern margins and a greater insight into the interpretation of geophysical information. Such geological data also help better target future deep drilling efforts into modern accretionary systems.

We report here a structural study across a shear zone interpreted to lie at the base of a duplex, which imbricated accreting packages of sedimentary rocks from a downgoing oceanic sequence. The chronology of deformation events and the synchronicity with the metamorphic climax in these rocks support the interpretation that the thrust was a once active portion of a décollement or plate boundary thrust, responsible for the incorporation of lower plate rocks into the accretionary prism. Different vein generations are reported and their textures described in detail. Moreover, the crosscutting relationships within the shear zone allow a reconstruction of the fluid pathways across the sedimentary sequence during lithification and diagenesis, immediately prior to underplating.

2. Geological setting

The Northern Apennines (Fig. 1) represent a fold and thrust belt built during the Upper Cretaceous–Eocene subduction of the Ligure–Piemontese oceanic basin that then evolved in the collision between the European and Adria plates.

The westernmost and highest units of the Apenninic belt belong to the Ligurian Domain, representing the remnants of the oceanic lithosphere of the Ligure–Piemontese basin and its transition to the nearby continental margin (Elter, 1975b; Marroni and Pandolfi, 1996; Marroni et al., 2001, 2004). Amongst these rocks, the so called Internal Units expose the most complete and best preserved section of the Ligure–Piemontese oceanic basin lithosphere (Decandia and Elter, 1972; Abbate et al., 1980), comprising: a well-preserved ophiolitic sequence, a Middle Jurassic/Late Cretaceous pelagic and hemipelagic sequence (radiolarian cherts, Calpionella Limestone and Palombini Shale) and a complex, thickening and coarsening upward turbiditic system, ranging in age from Campanian to Early Paleocene. The latter sequences comprise siliciclastic and carbonate turbidites subdivided into the Manganesiferous Shale, Mt. Verzi Marl, Zonati Shale and Gottero Sandstone formations. Early Paleocene, coarse-grained debris flows and slide deposits, referred to as the Bocco Shale Formation, represent the youngest deposits in the succession.

3. The Gottero Unit: accretionary deformation features

The Gottero Unit, cropping out extensively E of the city of Genoa, exposes (Figs. 1 and 2) the best preserved and complete section (around 2500 km thick, see Marroni et al., 2004) of the turbidite and debris flow deposits that infilled the Ligure–Piemontese oceanic basin (Nilsen and Abbate, 1983; Pandolfi, 1997), testifying to the sediment-rich nature of this basin as it became involved in subduction processes.

As with the entire Internal Ligurian complex, the Gottero Unit is affected by complex deformation resulting from two geodynamic events: the Upper Cretaceous-Eocene subduction and accretion of oceanic crust from the Ligure–Piemontese basin, followed, during the Upper Eocene, by the collision between the Adria microplate and the European continental margin. This event is recorded in the Internal Domain by a less intense and penetrative deformation.

Similar to the structural evolution recognized in many examples of fossil accretionary complexes (Moore and Sample, 1986; Fisher and Byrne, 1987; Kusky et al., 1997; Hashimoto and Kimura, 1999), the pre-Oligocene, subduction-related deformation history includes folding and faulting events, each representing a different stage in the accretionary history and subsequent exhumation towards the surface (for a complete description of the pre-Oligocene deformation D1, D2, D3 and D4 events see Marroni and Pandolfi, 1996; Marroni et al., 2004). According to Marrroni et al. (2004), the D1 deformation phase developed progressively through veining, folding and thrusting events that are referred to as the D1a, D1b and D1c sub-phases, respectively. D1b-related folds are decimeter- to meter-scale isoclinal, strongly non-cylindrical folds, with an approximately similar geometry (Fig. 3a). The folds are associated with an S1 axial-plane foliation, that is generally oriented parallel to the bedding surfaces, visible in the shales as a slaty cleavage (Fig. 3b). NE/SW striking shear zones, marked by foliated cataclasites, follow all the structures

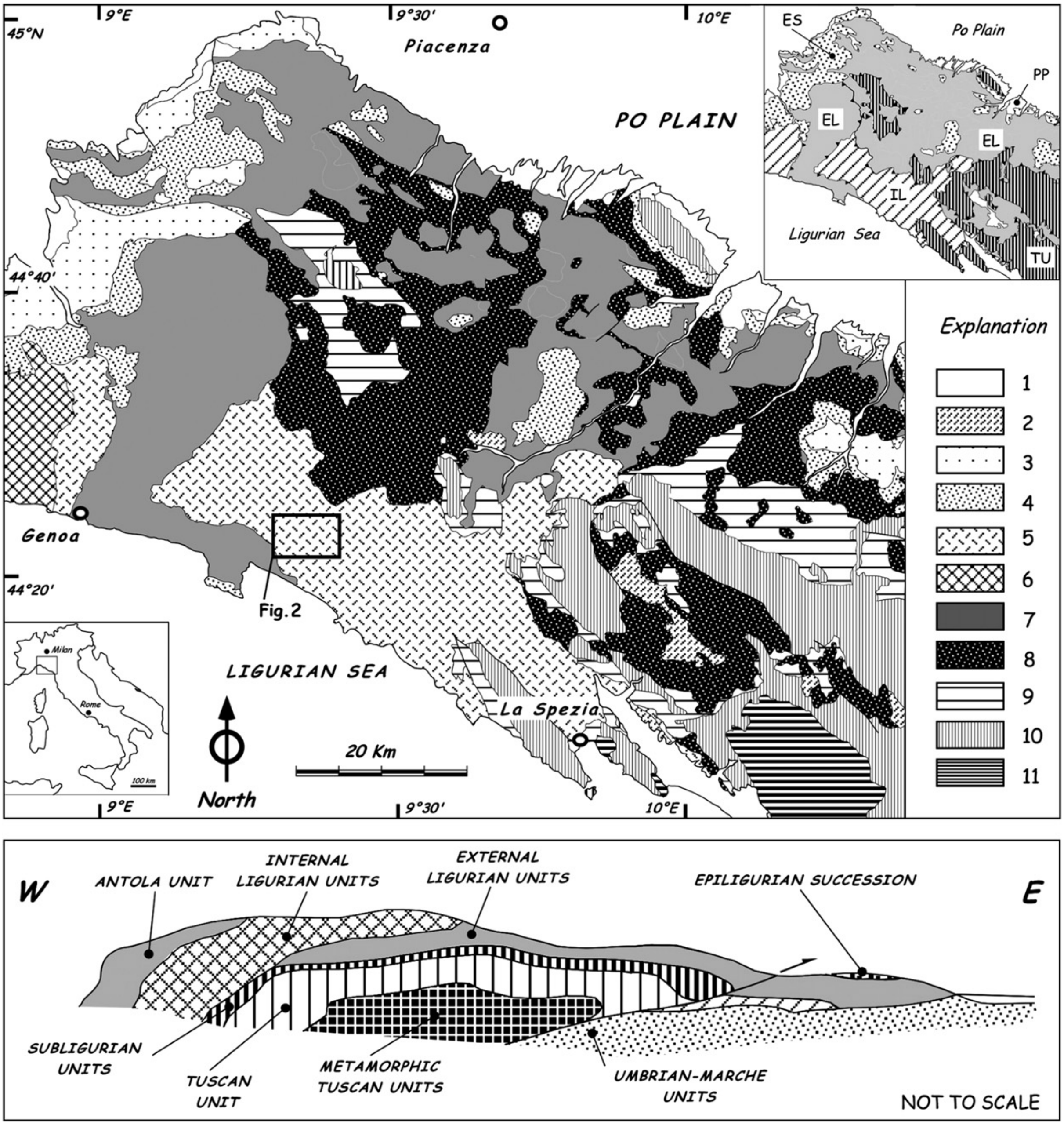

Fig. 1. Tectonic sketch map of the northernmost sector of the Northern Apennines and schematic cross section (redrawn from Elter, 1975a). 1: Plio-Quaternary deposits; 2: Plio-Pleistocene intramontane basins; 3: Miocene Epi-mesoalpine sedimentary sequences of the Tertiary Piemontese and Epiligurian basins; 4: Upper Eocene-Oligocene sedimentary sequences of the Tertiary Piemontese and Epiligurian basins; 5: Internal Ligurian Units; 6: Sestri-Voltaggio zone and Voltri group; 7: External "Eastern" Ligurian Units; 8: External "Western" Ligurian Units; 9: Subligurian Units; 10: Tuscan Units; 11: Low-grade metamorphic Tuscan Units (Apuan Alps window). PP: Plio-Pleistocene successions; ES: Tertiary Piemontese and Epiligurian successions; IL: Internal Ligurian Units; EL: External Ligurian Units; TU: Tuscan and Umbrian Units.

related to the D1b folding phase and are in turn folded by the D2-related folding event (Figs. 2 and 4). The analysis of these shear zones suggests a top-to-the-NW sense of shear, consistent with the vergence inferred for the D1-related mega-structures, once the D2 deformation effects are removed (Thio and van Wamel, 1990; Marroni and Pandolfi, 1996 and references therein). These shear zones are responsible for much of the emplacement and internal imbrication of the units (van Zutphen et al., 1985; Marroni et al., 1988, 2004; Marroni, 1990, 1991; Marroni and Meccheri, 1993; Pandolfi, 1997; Marroni and Pandolfi, 2002). Contemporaneous to this phase is the low temperature and high pressure (HP/LT) metamorphism that, according to illite and chlorite crystallinity and illite b_0 parameters (Leoni et al., 1996), developed under P/T conditions of 0.4 GPa/210–270 °C. The deformational features and the synchronicity with the metamorphic climax recorded in the units (see discussion of van Gool and Cawood, 1994) suggest that this event represents the underplating of the units onto the base of the accretionary prism as a series of imbricated duplexes. Particularly, the non-coaxial character

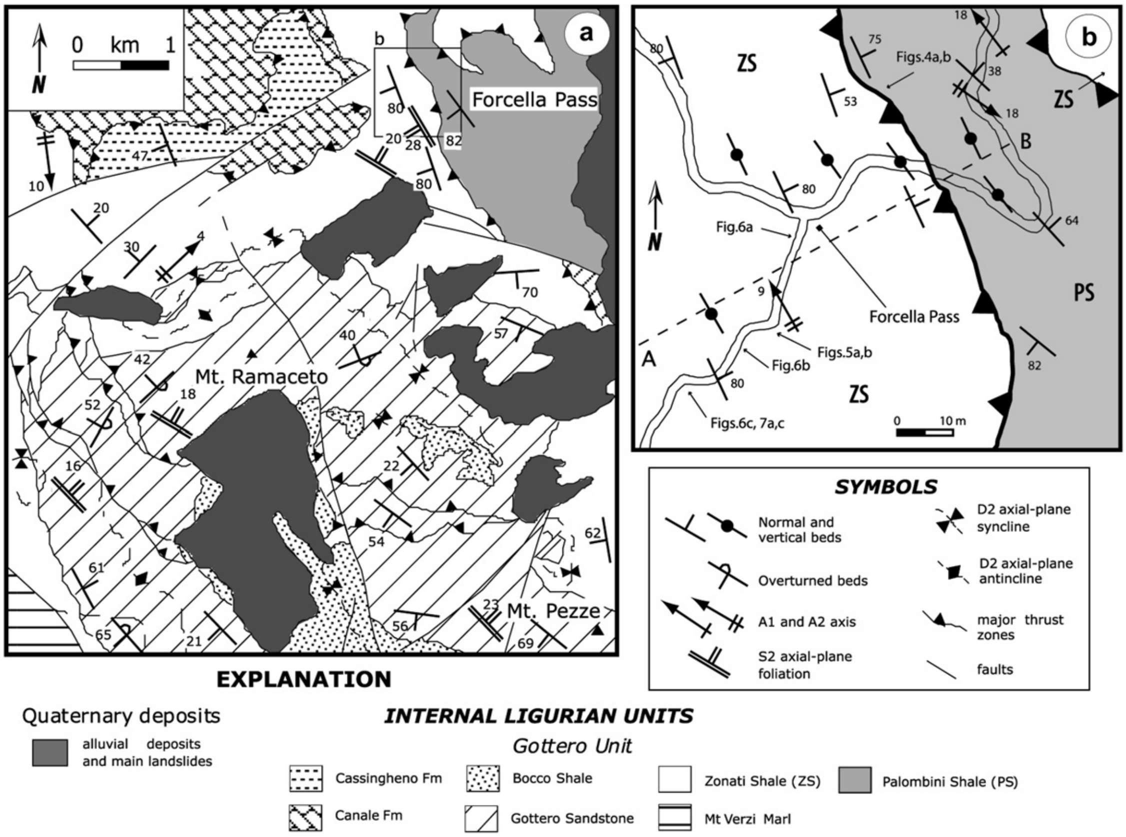

Fig. 2. Geological map of studied area (a) and detailed map of the Forcella outcrop (b). Location of (b) is marked at NE top of (a); dashed line in (b) is trace of schematic section of Fig. 4a. After Marroni et al., 2004.

of the D1 deformation can be interpreted as being generated in response to deformation during the stepping down of the décollement zone at the base of the Ligure–Piemontese accretionary wedge.

4. The outcrop of the Forcella Pass: mineralization and vein distribution

In the area of the Forcella Pass two imbricates of the Gottero Unit are juxtaposed by a 1–1.50 m thick shear zone (Figs. 2 and 4). The shear zone separates two formations of the Unit: the Palombini Shale (Valanginian–Santonian, Marroni and Perilli, 1990) in the footwall and the Zonati Shale (Campanian–Early Maastrichtian, Marroni and Perilli, 1990) in the hanging wall of the structure. The Palombini Shale comprises up to 3 m thick carbonate or marly turbiditic strata alternating with hemipelagic intervals, generally less than 1 m thick, and represents the final pelagic deposits covering the Ligure–Piemontese ophiolitic sequence (Decandia and Elter, 1972). The Zonati Shale consists of siliciclastic, thin-bedded turbidites (up to 50 cm thick), with a sandstone/shale ratio equal to 1.

The shear zone is marked by a highly disrupted zone including tabular fragments of siltstones derived from the Zonati Shale and carbonate blocks from the Palombini Shale, chaotically mixed with shaly turbidite intervals (Fig. 4). Although dismembered, the competent strata show a roughly sub-vertical alignment, parallel to the general bedding orientation in the two adjacent unit slices (Figs. 2 and 4). The entire stack crops out in the hinge zone of a kilometer scale D2-related fold and, since D2 folds are open with sub-horizontal axial planes, the general dip of both bedding planes and bedding-parallel, D1-related faults is sub-vertical. The strike of the unit stack ranges between N100E and N160E, with a variable dip from almost 90° to 50°, toward both the NE and SW (Fig. 4). Since the S1 cleavage is nearly bedding-parallel and is folded by D2 folds, whose axial-planes are sub-horizontal, the hinges of D2 folds are the zones where D1- and D2-related structural elements lie at the highest angles to one another.

The relative chronology with respect to the D2 phase of deformation, the structural features and the west-ward shear sense allow us to interpret the fault zone as related to the D1c event of Marroni et al. (2004) and similar to shear zones described elsewhere in the Internal Ligurian Domain (van Zutphen et al., 1985; Marroni et al., 1988; Marroni, 1990, 1991; Marroni and Meccheri, 1993; Pandolfi, 1997; Marroni and Pandolfi, 2002).

The thick hanging wall section of the Scisti Zonati Formation, immediately adjacent to the thrust, displays a dense development of veins (Fig. 4a). The Palombini Shale Formation of the footwall shows a few, scattered systems of veins close to the contact. In the Zonati Shale section in the hangingwall, the veins develop in scattered clusters across

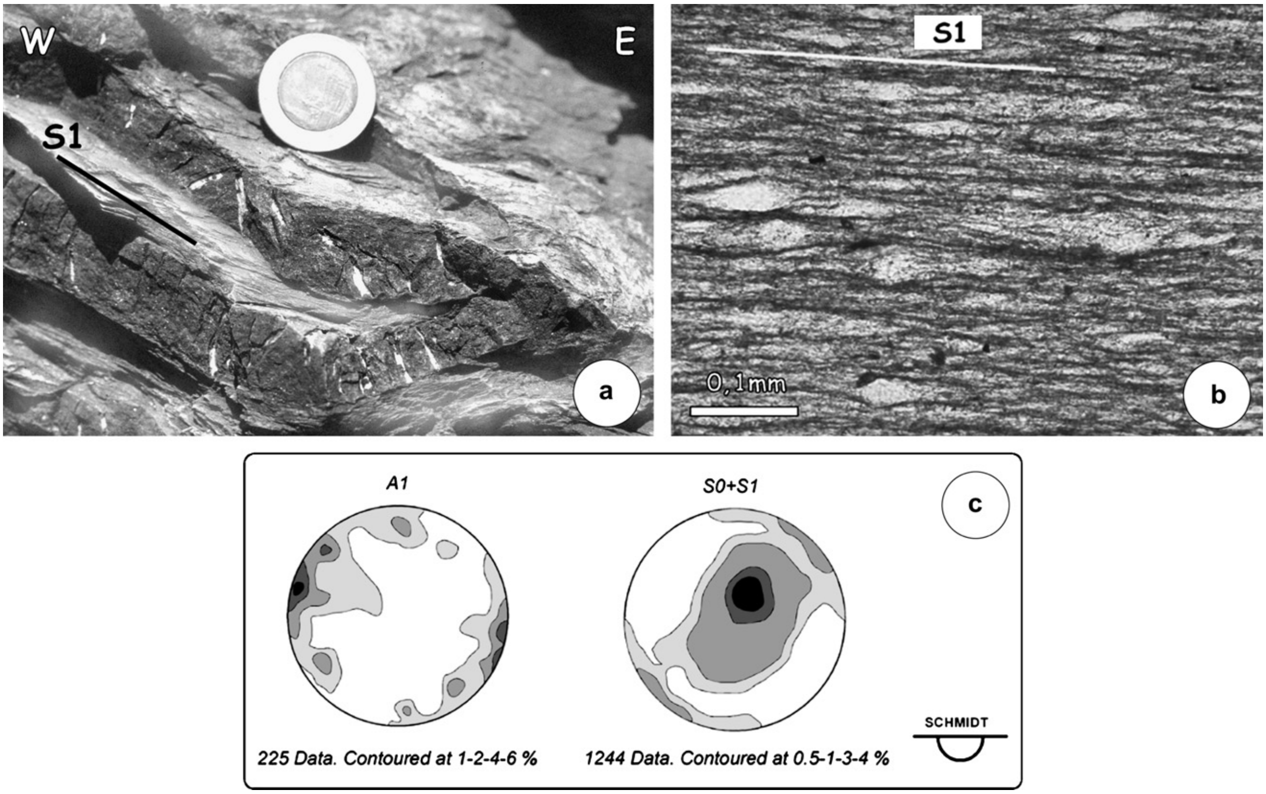

Fig. 3. Summary of D1 deformation phase main features. (a) Isoclinal F1 fold in Zonati Shale. (b) S1, axial plane, slaty cleavage. (c) Lower hemisphere projection of poles to F1 axes (A1), bedding (So) and foliation (S1) in the Gottero Unit. Facing of structures is false due to overprinting by D2 deformation. Once D2 is restored, D1 structures show westward vergence and facing.

the outcrop, each characterized by frequencies ranging from one vein per meter, up to one vein every few centimeters (Fig. 4a). Disrupted veins are found also in the thrust zone, where frequency is typically as high as the highest observed in the hanging wall. No such vein distribution is observed in rocks located far from the shear zone at the Forcella Pass, but a similar vein distribution is observed in other D1c shear zones cropping out in the Gottero Unit, as well as in other Internal Ligurian units.

The veins are arranged in two main groups showing different geometrical relationships with respect to bedding (So) and, consequently, to the S1 slaty cleavage; they also display different infilling textures (Fig. 4, Table 1). The first set comprises tabular vein-filled hydrofractures that lie parallel to So and S1 and located along sandstone/shale interface in the turbiditic sequence. The second group comprises two sets of vein-filled hydrofractures arranged normal to bedding in the more competent intervals (arenitic, silititic and carbonate) that can also be distinguished on the basis of their infilling textures (Table 1). The disrupted nature of veins in the thrust zone prevents in most cases the attribution of them to the different vein sets, except where a difference in textures is preserved.

4.1. Blocky veins parallel to So/S1

A set of thick calcite and subordinate quartz veins is developed parallel to the well-defined bedding of both Zonati Shale and Palombini Shale formations. They range in thickness from few tens of millimeters to 10 cm (Fig. 5a, Table 1) and have a good lateral continuity with the longest veins extending up to 10 m along the bedding, although a progressive decrease in thickness generally characterizes vein tips. The veins display regular boundaries, tabular shapes (Fig. 5a and b) and develop in the shaly intervals, parallel to bedding (average strike around N130E, see Fig. 4), along the interface between sandstone/siltstone and shales.

These veins are characterized by mosaic textures made up from irregular arrangements of euhedral calcite crystals of variable size (Fig. 5c and d); the longest crystal axis can locally reach 2–3 cm. The calcite crystals generally display Type I and II twins of Burkhard (1993), suggesting a possible temperature range for calcite formation of between 150 °C and 300 °C. The lack of syn-tectonic re-crystallization during the D2 deformation event (e.g. Marroni et al., 2004) allows us to relate calcite formation (and the associated P/T estimate) to the vein opening event, i.e. sometime before the first folding phase D1b (see D1a event of Marroni et al., 2004).

Quartz is also present in the veins, although it generally shows smaller crystals (Fig. 5d). The biggest crystals display undulose extinction, sub-grains and deformation bands. The mutual relationships between calcite and quartz suggest a possible contemporaneous crystallization. The frequent occurrence of interfingering borders indicates localized dynamic recrystallization.

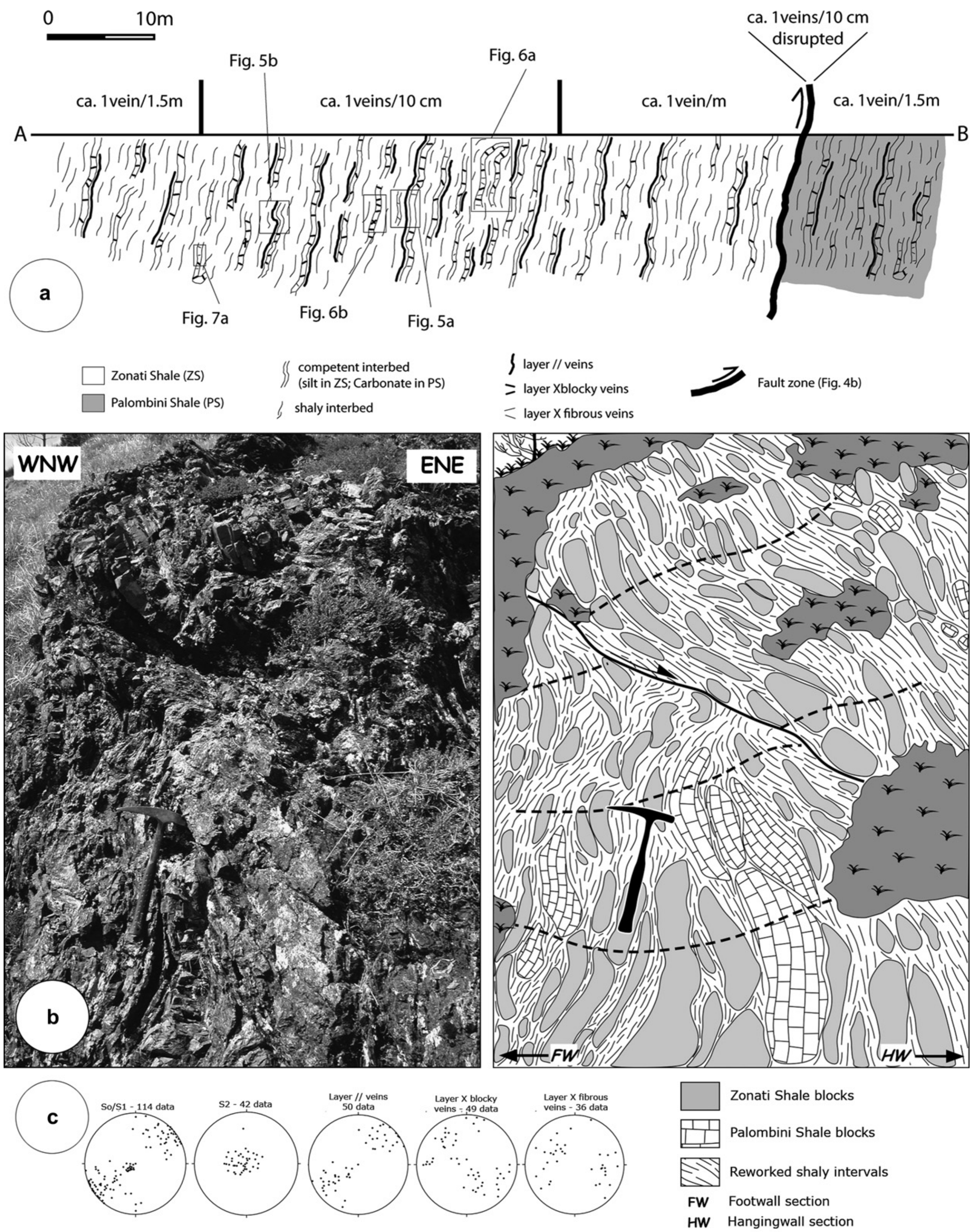

Fig. 4. (a) Schematic geological section through the fault zone at Forcella Pass (location in Fig. 2). General bedding attitude is shown, as well as schematic vein occurrence. Vein distribution across hanging and foot walls is also reported using the number of veins per meter or centimeter. (b) Field photo (left) and interpretive sketch (right) of D1c shear zone at Forcella Pass (Fault Zone of Fig. 4a). The shear zone crops out as a one-meter thick broken formation of Zonati Shale and subordinate Palombini Shale blocks dispersed in a shaly matrix. Although dismembered, bedding and S1 slaty cleavage are still fairly well preserved. D1c thrust is subvertical, lying in the hinge zone of a D2-related fold, and runs approximately from upper to lower side of pictures, as mimicked by S1 slaty cleavage in the matrix (thin black lines). As a reference, footwall (FW) and hangingwall (HW) location far from shear zone is indicated. Dotted lines represent D2 fold axial planes. D2-related low-angle normal faults are also shown with thick black line. (c) Lower hemisphere projection of poles to main structural elements and vein sets attitude across the Forcella Pass outcrop. The first two stereoplots, from left to right, show poles to bedding/S1 foliation and to S2 foliation, respectively.

Table 1
Table summarizing main structural and textural features of vein sets recognized across Forcella Pass fault zone

Relationship with bedding	Wall rock	Shape and lateral continuity	Thickness	Relationship with fault zone	Relationship with F1 folds	Composition	Texture	Crosscutting relationships
Parallel to So	Shale	Tabular, high lateral continuity up to 10 m along So	Up to 10 cm	In the hanging wall section adjacent to fault. Decrease in frequency far from fault zone	Folded by F1 folds	Calcite, +/− quartz. Wall rock inclusion	Blocky	Connected with normal to So, blocky veins. Optical continuity of infillings
Normal to So	Sandstone	Tabular, it usually causes boudinage of sandstone layer	5−6 cm (average)	In the hanging wall section adjacent to fault. Decreases in frequency far from fault zone	Folded by F1 folds. No particular arrangement with respect to fold structural elements	Calcite, +/− quartz. Inclusion free	Blocky	Connected with parallel to So veins. Optical continuity of infillings
Normal to So	Sandstone	Lozenge, it usually tapers out before end of sandstone layer. Locally bifurcating terminations	2−3 mm (average)	Random. Occurs far from fault and elsewhere in the unit	Concentrated only in F1 limbs. Observed elsewhere in other F1 folds	Calcite. Quartz found only as syntaxial growth at the calcite−wall rock interface. Median line	Fibrous, antitaxial. Fibers sub-normal to vein wall and fold axis	Truncates parallel to So veins. No observed relationships with respect to normal to So blocky veins

Vein boundaries are continuous and indented (Fig. 5d), but indention is due to vein opening by grain-to-grain separation, rather than grain breakage.

The main feature at any scale is the ‘‘dirty’’ aspect of the veins due to the occurrence, in the infilling, of variable sized inclusions of dark-green to black shaly material, representing fragments of the host rock (Fig. 5b). Fragments are generally similar to the host rock, showing fine quartz grains surrounded by abundant phyllosilicates that define a continuous slaty cleavage.

The cleavage observed in the inclusions is always parallel to the host-rock S1 foliation. The shale particles are irregularly distributed along the entire length of the veins and intermixed with the blocky quartz/calcite crystals. Generally, where the shale inclusions occur, a gross bedding-parallel layering composed of tabular calcite veins alternating with lenses of shale, is visible (Fig. 5b and c). Some of the wall rock inclusions are only partly detached from the host rock.

In general, different arrangements of the three main components of the vein infillings occur: blocky calcite aggregates, blocky quartz aggregates and wall rock fragments. In a single vein these three elements occur in different proportions and with different geometrical relationships one to each other. One single fracture event can be entirely filled by large blocky-calcite crystals, or by a mixture of calcite and large clear quartz crystals (compare Fig. 5a with b, c and e). Quartz and calcite can be chaotically intermixed or arranged in regular layers defining a compositional layering. Along a single vein, a transition from wall rock fragments-rich zones to ‘‘clear’’ zones is commonly observed. When rock particles occur, they can form discontinuous ribbons alternating with calcite layers, suggesting multiple opening and mineralization events (Fig. 5c).

Seams of residual opaque minerals also develop parallel to bedding and to the alternations of calcite and wall rock fragments (Fig. 5e), together with stylolites in the calcite aggregates. This observation demonstrates that pressure solution seams occur parallel to So, to S1 cleavage and to the opening direction of the S1-parallel veins, indicating possible alternation of dilation and compaction in the same direction.

4.2. Blocky veins normal to So/S1

The more competent layers of the Zonati Shale Formation, such as the silty and sandy intervals, preserve frequent pinch-and-swell and boudinage structures. The fractures between the different boudins are filled by sets of calcite veins developed normal or at high angles to bedding (Fig. 6a and b, Table 1): their general orientation is reported in the stereonet of Fig. 4. The veins run sub-normal to bedding and, together with bedding, are folded by both D1b (Fig. 6a) and D2-related folds.

The veins of this set are up to 5−6 cm thick and are generally arranged in two conjugate systems making an angle of 30° to 50° in a plane normal to the So mean strike. On exposed So

Fault zone outcrops in the hinge zone of a D2-related fold, which have sub-horizontal axial plane (i.e. S2 foliation), so that both bedding and S1 foliation are sub-vertical. This makes D2-related fold hinges the best place to observe D1-related structures, because they are always easily distinguished from later deformation. The last three stereoplots show respectively poles to layer-parallel veins, layer-normal blocky veins and layer-normal fibrous veins.

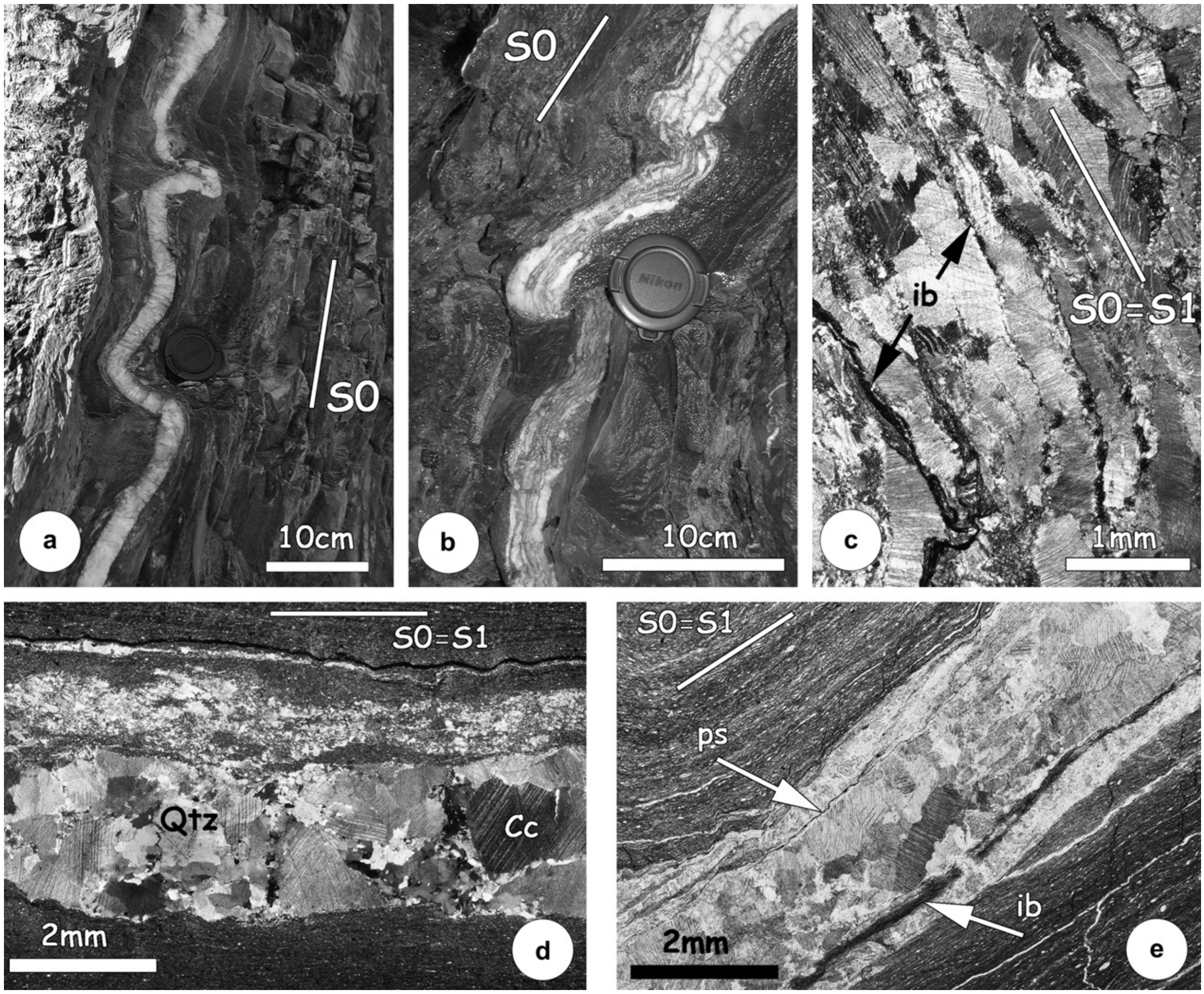

Fig. 5. Meso- and micro-scale characteristics of layer-parallel veins (D1a of Marroni et al., 2004). (a) High lateral continuity of veins. (b, c) Typical occurrence of layer-parallel veins filled by intermixing of calcite, subordinate quartz and wall rock inclusions (ib). Note the frequent layer-parallel alternation of calcite layers and shaly inclusions; mean So–S1 attitude is shown. (d) Blocky quartz (Qtz) and calcite (Cc) interfingered crystals constitute the main vein infilling. Note irregular lower border of the vein (crossed nicols). Mean So–S1 attitude is shown. (e) Dirty aspect of the veins under plane polarized light, due to the presence of wall rock inclusions (ib) and pressure solution seams (ps). Note strong parallelism with mean So–S1 attitude.

planes, the two sets are almost parallel (Fig. 6b). Vein shape varies from thick tabular and parallel-sided to lozenge-shaped. Whilst tabular veins separate completely the boudins, progressive tapering veins generally end before the end of the competent layer. In the shaly interbeds, the veins can be flattened or dissected and sheared by the subsequent development of D2-related folds through flexural slip.

These veins show a clear infilling with a mosaic fabric defined by irregular arrangements of calcite crystals up to 2–3 cm in size, with an almost total absence of wall rock particles in the veins. This last observation, together with the occurrence of sharp vein walls, suggests that the coarse layers were fairly lithified and competent allowing them to deform in a brittle fashion.

When D2-related flexural slip does not dissect veining, thick rectangular-shaped normal veins clearly branch into blocky, layer-parallel veins (Fig. 6c), as demonstrated also by optical continuity of calcite crystals between the two vein sets.

4.3. Fibrous veins normal to So/S1

Syn-tectonic veins with fibrous textures represent the third set of veins recognized in the Forcella Pass outcrop, though their occurrence is reported elsewhere in the Gottero Unit, far from thrust zones (Marroni and Pandolfi, 1996). These veins develop at high angles to bedding in the more competent and coarser layers, striking preferentially in a direction approximately parallel to the axis and usually concentrated in the limbs of the mesoscale D1b folds (Fig. 6a, Table 1). The mean vein thickness is 2–3 mm, with rare centimeter-scale veins (up to 2 cm). These veins are usually lozenge-shaped with sharp walls (Fig. 7a and b), and their length is normally less than the host rock layer thickness, although locally they are associated with bedding boudinage. The veins progressively thin (Figs. 6a and 7a) and show both single and bifurcating terminations. As with the blocky layer-normal veins, the fibrous veins are locally arranged into conjugate sets making angles of 40° to 50° on a plane normal to bedding.

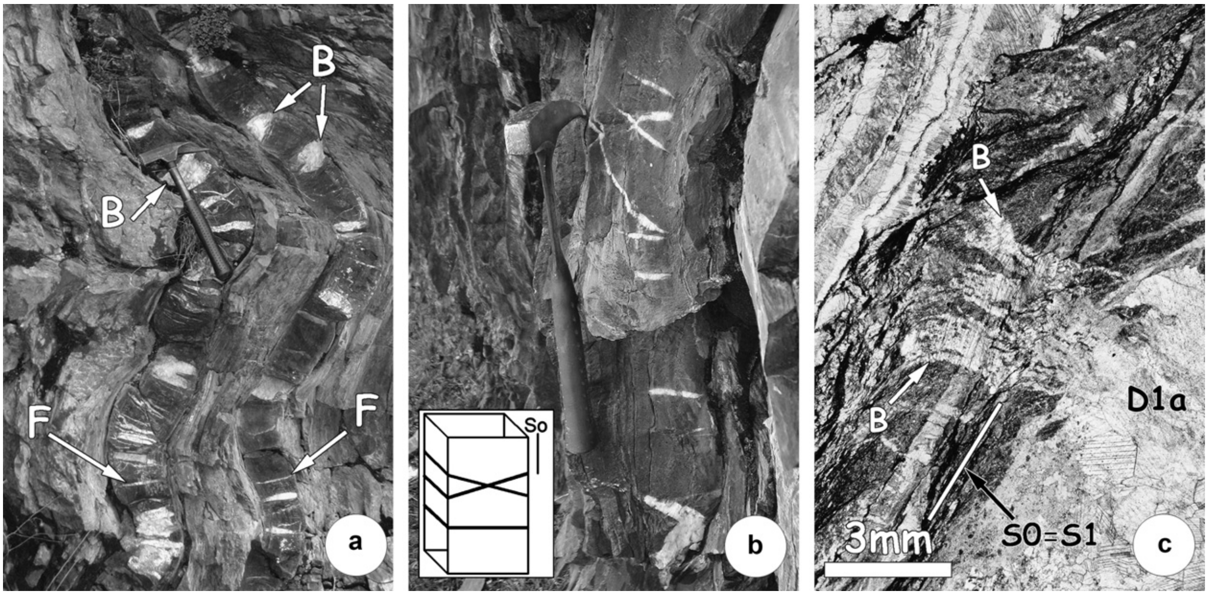

Fig. 6. Meso- and micro-scale features of layer-normal blocky veins (B). (a) Veins (B) lie in the more competent layers between boudins and are folded by D1a folds. Thin fibrous veins (F) are also visible. (b) Veins are generally normal to bedding and are frequently arranged into conjugate systems. In the inset, the orientation of the two systems (lines in the block) viewed in So planes and in planes normal to So is shown. (c) Blocky, layer-normal vein (B) developing from the layer-parallel vein (D1a) at the lower left corner of picture. Vein B in the picture cuts at high angle to both bedding (So) and S1 cleavage. Optical continuity between crystals of the two vein sets is also visible. Mean So–S1 attitude is shown.

In a single coarse bed, fibrous and blocky veins can be found together (Fig. 7a), but clear crosscutting relationships have not been recognized. What is clearly observed is that they are both involved in D1b and D2-related folding (Fig. 6a). The microanalysis of these samples reveals that when normal fibrous veins come into proximity with layer-parallel veins they consistently truncate the layer-parallel system (Fig. 7c).

Fibrous textures are always well preserved, both at outcrop- and microscopic-scales, marked by the antitaxial growth of calcite fibers with well-defined, discontinuous, dark median lines represented by scattered seams of host rock particles aligned in the center of the vein (Fig. 7b). Thin fibrous quartz infillings sometimes limit the calcite fibers adjacent to the vein walls, and represent the only quartz found in these veins (inset of Fig. 7b). The quartz fibers display syntaxial growth from the very fine-grained quartz of the wall rock.

Typically the calcite fibers are oriented almost normal to the vein walls and their direction appears to represent an extension lineation oriented roughly normal to the D1b fold axes. The fibers are generally curved in a plane perpendicular to S1, corresponding to the *xz* plane of the incremental strain field and are elongated parallel to *xy* axis.

5. Discussion

5.1. The Forcella Pass fault zone and its regional accretionary setting

Studies of modern and ancient examples have demonstrated that accretionary prisms grow by repeated transfer of sediments and rocks from the down-going plate to the upper plate through the plate-boundary thrust represented by the décollement zone (e.g., Moore and Sample, 1986; Sample and Fisher, 1986; Hashimoto and Kimura, 1999; Bangs et al., 2004). Each décollement step down isolates packages of rock that are transferred to the prism, underplating and uplifting the previously accreted units; the tectonic contact that separates the two imbricates into the prism is representative of the once active décollement. Thus, the décollement is incrementally and repeatedly preserved along the boundaries of each package that is transferred to the accretionary prism.

As stated above, the studied shear zone is part of an accreted unit (i.e. the Gottero Unit) made up of thrust imbricates of sedimentary rocks covering the oceanic basement. The Forcella Pass fault zone, as well as each of the package-bounding thrust in the Gottero Unit, is folded by D2-related folds and shows the same geometry, shear sense and structural features of pre-D2 shear zones (D1c) observed elsewhere in the entire Internal Ligurian Domain (van Zutphen et al., 1985; Marroni et al., 1988; Marroni, 1990, 1991; Marroni and Meccheri, 1993; Pandolfi, 1997; Marroni and Pandolfi, 2002). The Ligurian units can then be considered to represent underplated duplexes deformed and juxtaposed during accretion (i.e. D1 deformation events of Marroni et al., 2004): in this sense the D1-related shear zones bounding the imbricates can be reasonably seen as examples of paleo-décollements.

5.2. Vein development and possible fluid pathway reconstructions

The detailed analysis of the complex system of syn-tectonic veins associated with faulting-related deformation, suggests a strong interaction between deformation, diagenesis,

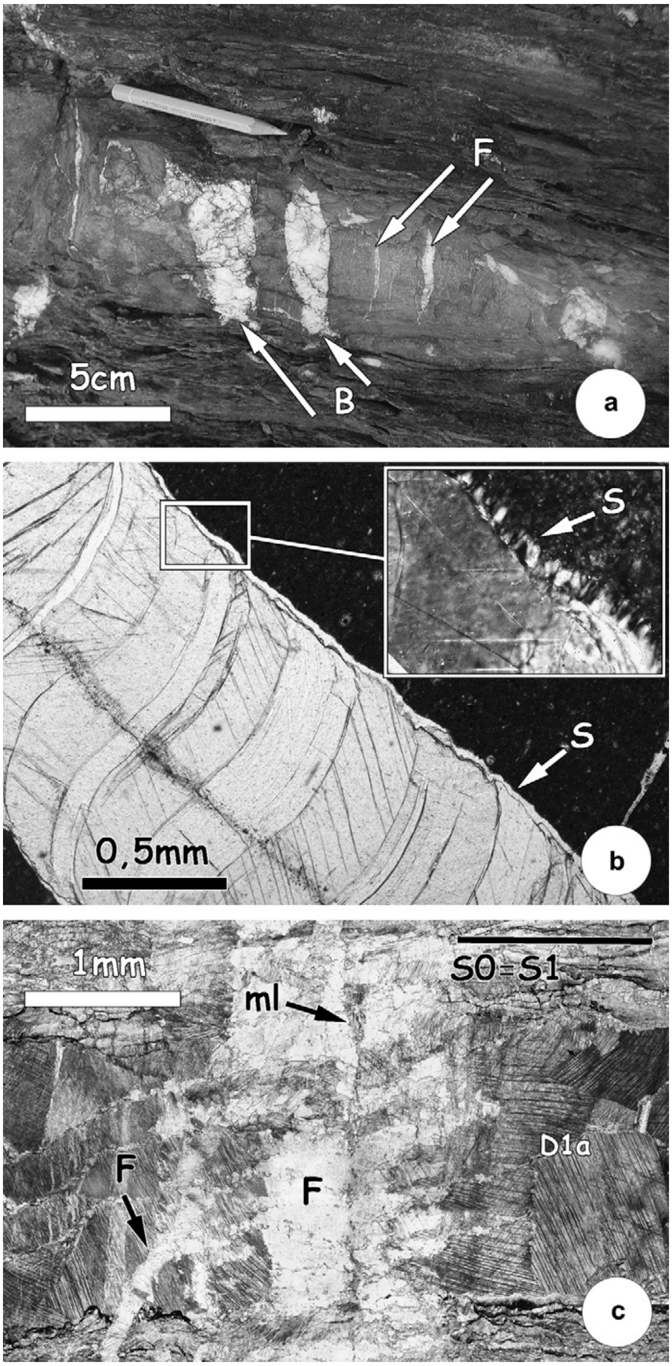

Fig. 7. Meso- and micro-scale features of layer-normal fibrous veins. (a) Coexistence in the same sandy layer of thick blocky (B, on the left) and thin fibrous (F, on the right) veins. (b) Calcite infilling shows antitaxial texture. At vein borders a thin syntaxial filling of quartz (S) develops, as seen in the inset of the picture. (c) Geometrical relationships between fibrous (F) and layer-parallel (D1a) veins: F sharply cuts D1a. Mean So–S1 attitude is shown; ml in F vein indicates the median line.

lithification and anisotropic fluid advection, through an evolutionary sequence that records continuous changes in the mechanical properties of the sediments.

The first set of veins from the Forcella Pass outcrop seems to record fluid expulsion accompanying compaction and lithification during progressive burial. This vein set developed parallel to bedding, in the shales, at the interface between sandy and shaly beds. The veins have been deformed mesoscopically by both D1 and D2 phases with the same geometry and style as the sedimentary layers, so they must have formed early in the history of the units, before any folding occurred (D1a phase, Marroni et al., 2004). However, these veins record important changes in the mechanical properties that point to a syn-diagenetic deformation. The absence of grain breakage at vein boundaries and their indented aspect, indicate that the sediments were weak enough for the veins to open by grain-to-grain separation, suggesting that compaction and cementation were poorly developed. The concentration of veins in shaly beds at the sandstone/shale interface suggests that preferential dehydration and fluid injection were focused along mechanical anisotropies such as bedding surface (Fig. 8a). The arrangement of shale inclusions in layers parallel to the wall rock bedding, and their alternation with clear quartz and calcite filling, indicate multiple opening and mineralization events, with each opening step occurring at the wall rock-vein interface and isolating ribbons of shale inclusions. Parallel to this general layering, seams of residual opaque minerals developed both in the wall rock and in the shale inclusions, together with stylolites in the calcite crystals. Therefore, pressure solution occurred in fracture plane surfaces, suggesting a possible fluid-controlled pressure fluctuation leading to alternating phases of dilation and collapse of the fracture walls (Sibson, 1990, 1992; Miller et al., 2003). In addition, the lack of textures typical of bedding-parallel shear in the veins tends to suggest that they did not form due to bedding plane slip in response to D1 flexural folding (as, for example, described by Fitches et al., 1986).

The second set of veins lie normal to bedding and cut only the sandstone layers. They lack any systematic geometrical relationships or spatial distribution with respect to folding, and are therefore also interpreted to pre-date folding. Locally, this vein set branches directly into blocky, layer-parallel veins (Fig. 8a′), suggesting that they are broadly the same age and represent a change in fluid path from layer-parallel to upward, layer-normal flow that occurred prior to folding. As these veins have clearer infillings and sharper walls compared to the layer-parallel set, this set may have been syn-diagenetic, involving a change in lithification state of the host rock, at least in the sandstone units. The shaly layers were probably less cohesive than the sandstones at the time, and such a difference in lithification state would tend to enhance preferential focusing of fluid circulation within the shales. The further fluid accumulation at the base of the sandstone and the barrier represented by the lithified sandstones themselves may have helped in the build-up of high fluid pressures eventually leading to local brittle fracturing of the more lithified, competent sandy layers and to upward-directed fluid flow.

In order for a hydrofracture to form, the minimum principal effective stress must equal the tensile strength of the rock (Twiss and Moores, 1992; Scholz, 2002). Following the semi-quantitative mechanical models of stress permutation proposed for development and evolution of hydrofractures observed in other Apennine faults (Zuccale Fault, Elba Island,

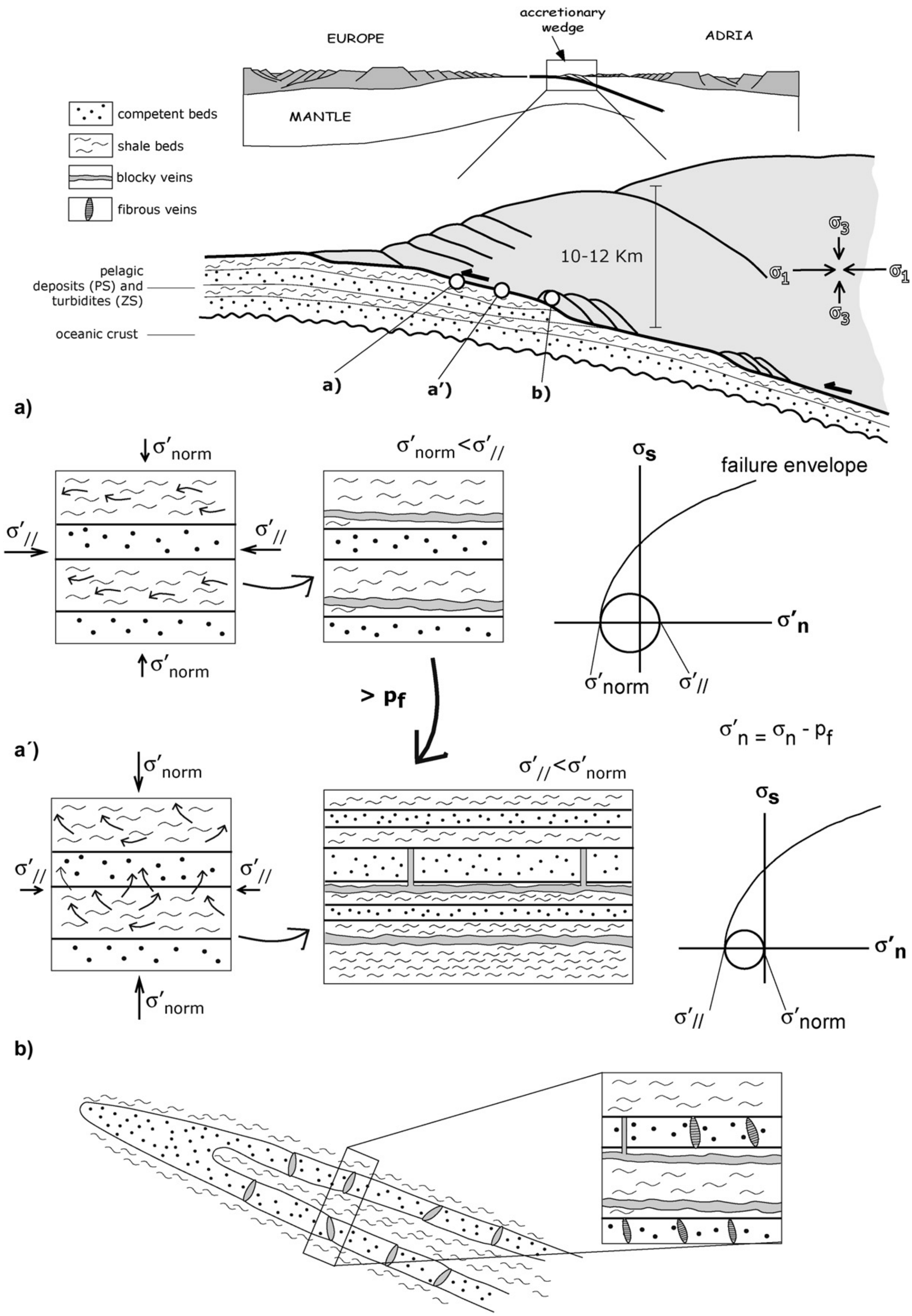

Fig. 8. Schematic reconstruction of fluid pathways, prior and during underplating, in the Gottero Unit. Upper scheme shows the Alpine/Apennine accretionary wedge setting from Marroni et al. (2004). Regional stress field after Mandl (1988). Not to scale. For each veining episode (a, a′ and b) a schematic cartoon illustrates vein formation in the sedimentary sequence. To the right of the cartoon in (a) and (a′) the 2D state of stress is represented as a Mohr circle. σ_s and

Italy, Collettini et al., 2006) and in other analogues of accretionary thrusts (Rodeo Cove Fault Zone, California, Meneghini and Moore, in press), a pressure build up results in a shifting of the Mohr circle toward negative values so that fracture occurs when the minimum effective stress become tangent to the failure envelope. In a 2D model, like the one shown in Fig. 8, with one principal axis lying parallel to bedding ($\sigma_{//}$) and the other one oriented normal to bedding (σ_{norm}), hydrofractures will form preferentially in the shale, since cohesion of shale is lower than that of sandstones and since the higher permeability in the décollement/bedding-parallel direction (if $\sigma' = C + \mu(\sigma - p_f)$ and $C_{sandstone} > C_{shale}$, then $p_{fshale} < p_{fsandstone}$), so that $\sigma_{norm}' < \sigma_{//}'$ and tangent to the failure envelope (Fig. 8a). Fluid pressure in the channel is increased by the progressive compaction and lithification of the sandstones, reducing values of effective stresses, so that $\sigma_{//}' < \sigma_{norm}'$ and tangent to the failure envelope: new fractures develop normal to it (i.e. normal to bedding) and stay open until the pressure gradient is balanced (Fig. 8a). In fact, the opening of normal fractures may have enhanced the connectivity between layer-parallel channels, causing a temporary decay of overpressure, favoring mineral growth and the onset of a new cycle of increasing fluid pressure. The blocky texture of the veins indicates growth at a rate slower than fracture opening, suggesting the instantaneous nature of these hydrofacturing episodes.

When fractures open normal to bedding, σ_{norm}' is higher than $\sigma_{//}'$, so that bedding-parallel fracture walls experience collapse, with consequent formation of pressure solution seams and stylolites.

The onset of the D1b folding phase, which is likely related to the transfer of thick packets of sediments into the accretionary prism, led to the formation of further extensional fractures perpendicular to So during non-cylindrical isoclinal folding (Fig. 8b). In fact, a clear extensional origin related to D1b folding is consistent with the geometry of the fibrous layer-normal vein system. The D1b folds (Figs. 3 and 6a; see also Marroni et al., 2004) have typically developed thickened hinges and boudinaged limbs, implying that sediments must have suffered a strong layer-parallel extension. Moreover, the fibers in these veins grew in directions oriented at high angles to the fold hinges, recording the progressive stretching of the competent layers during folding and suggesting that competency contrast again played a key role. At the time of folding, the sandstones must have been completely lithified, as also suggested by blocky layer-normal veins, creating the right competence contrast relative to the shales to favor fracturing of the sandy layers. The lozenge shape of the veins seems to confirm this hypothesis. In contrast to the instantaneous open of the blocky-infilled veins, the fibrous texture of the second group likely indicates mineral growth at a rate commensurate to fracture opening in a regime of progressive extension.

The onset of deformation through isoclinal folding determines a dominant stress field that prevents further layer-parallel hydrofracturing, favoring layer-parallel extension (Bettelli and Vannucchi, 2003).

The depicted pre-folding scenario describes fluctuations of fluid pressure during progressive burial that may have occasionally produced instantaneous and cyclic local switching of the principal stress axis and the formation of the two sets of veins lying at high angles, as shown schematically in Fig. 8a–a′. In fact, the occurrence of a conjugate system of layer-normal veins making an angle of 30° to 50° and consequent boudinage of the competent layers, are consistent with layer-parallel extension and a layer-normal σ_1 (e.g. Twiss and Moores, 1992).

5.3. The anisotropic and dynamic character of permeability and the cyclic build-up of fluid pressure in accretionary prisms

The analysis of accretion-related structures in the Gottero Unit indicates that the tectonic processes active during accretion of thickly sedimented units interplays with the processes of burial, dewatering and lithification, with fluctuations of pore fluid-pressure and injection of over-pressured fluids as the main controlling parameters. This observation fits with the available data from modern accretionary prisms, where heterogeneous distributions of overpressured fluid pockets have been mapped along the décollement zone (Brown et al., 1994; Shipley et al., 1994, 1995; Moore et al., 1995a,b; Moore and Klaus, 1998; Bangs et al., 1999; Screaton et al., 2000; Bourlange et al., 2003). The intimate relationship between tectonism and diagenetic processes has been inferred also from studies of numerous ancient analogues (e.g. Fisher and Byrne, 1987; Maltman, 1995; Vannucchi and Maltman, 2000; Labaume and Moretti, 2001; Bettelli and Vannucchi, 2003).

Prior to accretion (i.e. prior to the D1b folding phase), while sediments were moving to depth, fluid transport in the Ligurian sediments was mainly accomplished via the layer-parallel system of hydrofractures (D1a), suggesting that enhanced fluid flow was focused in bedding- and décollement-parallel fractures. Although most of the data on fossil accretionary prisms indicate that underthrusting is dominated by layer-parallel extension, which contributes to mélange formation (e.g. Fisher and Byrne, 1987; Cowan, 1985; Needham, 1995), the opening of the pre-folding, bedding-parallel

σ_n axes are respectively shear and normal stresses, with the (′) sign that conventionally refers to effective stress as defined in the figure. σ_{norm} and $\sigma_{//}$ are defined as the stresses acting normal and parallel to bedding, respectively. (a) Lower cohesion of shales favors layer parallel hydrofracturing and preferential fluid circulation along the weak planes represented by the sandstone/shale interfaces. At this time $\sigma_{norm}' < \sigma_{//}'$ and tangent to failure envelope. The regional stress field helps layer-parallel hydrofracturing of unlithified shales at the interface. (a′) Progressive lithification of the sandstones allows different mechanical conditions and competency contrasts to exist between the shales and sandstones. The sandstones act as barriers for fluids causing fluid pressure build up. Subsequent hydrofracturing of sandstones allows upward migration of fluids and local and temporary flipping of stress axes: σ_{norm}' is now bigger than $\sigma_{//}'$ and tangent to failure envelope. (b) The third set of fibrous veins develops due to fold limb extension during F1 isoclinal folding and regional layer-parallel extension.

hydrofractures in the Gottero Unit sediments indicate a local stress field with sub-horizontal (i.e. almost layer-parallel) σ_1, whilst σ_3 was oriented at high angles to bedding. This stress orientation is consistent with the regional stress field characterizing the upper plate in subduction margins, in which the regional σ_1 lies at a very low angle to the décollement zone and the sedimentary layering (e.g. Davis et al., 1983; Mandl, 1988). Confirmation of this idea can be found in studies of modern convergent margins, where fracture porosity estimations and monitoring have been conducted along the décollement zone (e.g. Brown et al., 1994; Shipley et al., 1994; Moore et al., 1995a,b; Bangs et al., 1999; Bourlange et al., 2003). The models derived from these studies indicate that the décollement is a site of concentrated and episodic fluid flow. To account for this sub-horizontal flow, high décollement permeability must be expected, which, in turn, requires the development of regional dilation through hydrofracturing, with fluid pressures in excess of lithostatic pressures, in surfaces oriented parallel to the décollement. Similarly, permeability experiments and modeling conducted on large strike-slip fault zones demonstrate anisotropy: the permeability in directions parallel to the fault is up to three orders of magnitude greater than in the direction normal to it (Faulkner and Rutter, 2001).

In the case of subduction of thickly sedimented packages, we can expect the fluids to migrate through the thick slices of sediments, before reaching the décollement, using the weak, more permeable bedding layers.

The reconstructed veining history also suggests a transient hydrofracturing and fluid injection regime that seems to be strongly controlled by diagenesis and lithification processes, which determine changes in the mechanical and permeability conditions of the sediments. Dynamic changes in permeability, causing cyclic build-up of fluid pressure, are suggested in the models for modern prisms, and in conceptual models from natural and experimental examples, where hydrofracture dilatancy is associated with fault-valve action (Sibson, 1990, 1992) or dehydration kinetics (Miller et al., 2003). All these examples emphasize the coexistence of dilatant and compactive deformation modes, as suggested also in this study for the early layer-parallel vein set.

In all cited models, this cycling has been tentatively correlated with the seismic cycle along modern subduction zones, through migration in the décollement of overpressured fluids pockets as "pressure waves" upward from the seismogenic zone. The injection would occur during or after the fracturing associated with the slip event, with compactive shear localization ahead of the injection (Bourlange et al., 2003).

6. Conclusion

The ancient fault zone cropping out at Forcella Pass is interpreted to represent a paleo-décollement, i.e. a once-active portion of the plate-boundary thrust that imbricated a sheet of sedimentary rocks into the Ligurian accretionary prism. Faulting is associated with an extensive and complex vein system comprising three sets: (1) early hydrofractures parallel to bedding, (2) broadly contemporaneous hydrofractures normal to bedding, and (3) D1b fold-related fibrous veins concentrated along the fold limbs.

Geometrical relationships with the D1b folding phase and crosscutting relationships constrain a chronology of vein formation, suggesting a complex history of dehydration of sediments and fluid flux, interplaying closely with lithification and diagenesis. The two earlier systems of hydrofractures oriented at a high angle to one another suggest cyclic fluctuations of fluid pressure and repeated hydrofracturing events resulting in switching of local principal stresses around the fault. The most efficient dewatering system appears to have been active prior to folding and developed preferentially along the high permeable bedding- and décollement-parallel fractures. The Forcella Pass Fault zone can therefore be considered as a good fossil analogue of the dilated, high-porosity and high-permeability channels observed along décollements in modern accretionary prisms.

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