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Evolution, formation mechanism and kinematics of a contractional shallow shear zone within sedimentary rocks of the Northern Apennines (Italy)

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Key words: Shallow level shear zone, very-low grade metamorphism, kinematics, Northern Apennines, Tuscan Nappe

ABSTRACT

Integration of a meso- and microstructural study with illite-cristallinity, X-ray and SEM analyses allow us to describe the fabric, kinematics, evolution and formation mechanism of a shallow level shear zone, developed during thrusting within the Tuscan Nappe, the deepest non metamorphic tectonic unit of the Northern Apennines. This shear zone has been described for the Mt. Aquilaia area, located west of the Mt. Amiata geothermal area. Shearing took place during the Late Oligocene-Early Miocene and accompanied the superimposition of Middle Jurassic carbonate rocks (Marne a *Posidonomya* Fm.) on the Eocene-Oligocene pelagic sequence (Scaglia Toscana Fm.). The shear zone, less then 10 m in thickness, consists of strongly deformed rocks which can be subdivided in three shear zone domains characterised by different structures indicating top-to-the-east shearing. The A domain (about 1 m thick) is characterised by an highly damaged zone composed of a clayey gouge with centimetre and decimetre limestone clasts dispersed within the clayey "matrix". The B domain (max 4 m thick) consists of detached and overturned decimetre folds, a pervasive tectonic foliation, reverse faults and s-c structures which affected the Jurassic calcareous rocks. The C-domain (max 2 m thick) consists of s-c structures. The shear zone developed through two uninterrupted stages of a single episode of deformation: the first stage produced the highly damaged zone within the Scaglia Toscana clayey rocks (A-domain) and the folding within the Marne a *Posidonomya* calcareous rocks (B-domain); the second stage concentrated on the A-domain, caused the detachment of the folds and the development of reverse faults within the B-domain, and produced the development of S-C structures within the C-domain. During shearing, strain was partitioned. Strain partitioning was mainly controlled by different processes, such as: (a) lithological contrast, (b) lithological anisotropy and (c) fluid-assisted deformation and related metamorphic reactions. Pressure solution and solution transfer are the mechanisms which produced the S_1 foliation. A very low-grade metamorphism affected only the very strongly deformed pelitic dominantly rocks. The metamorphism produced the development of new clay minerals (illite-illite/smectite), only developed within the cleavage domains of the folded rocks. This has been interpreted as mainly due to deformation heating, coupled with fluid circulation associated with cleavage development.

RIASSUNTO

L'integrazione di uno studio meso- e microstrutturale, dell'analisi della cristallinità dell'illite, di analisi a raggi X e condotti con il microscopio a scansione elettronica (SEM) hanno permesso di analizzare il *fabric*, la cinematica, l'evoluzione ed i meccanismi di formazione di una zona di taglio compressiva che si è sviluppata entro la Falda Toscana, la più profonda delle unità non metamorfiche dell'Appennino Settentrionale. Questa zona di taglio affiora nell'area di Monte Aquilaia, ad ovest dell'area geotermica del Monte Amiata. Lo sviluppo di questa struttura ha permesso la sovrapposizione delle rocce carbonatiche del Giurassico medio, appartenenti alla formazione delle Marne a *Posidonomya*, sulla successione pelagica eocenico-oligocenica della Scaglia Toscana. La zona di taglio è spessa mediamente meno di una decina di metri ed è suddivisibile in tre domini caratterizzati da diverse strutture. Il dominio A, discontinuo, è spesso al massimo un metro ed è caratterizzato da un *gouge* di faglia sviluppato a spese delle argilliti e calcari della Scaglia Toscana. Entro tale zona intensamente deformata sono presenti clasti calcarei centimetrici e decimetrici immersi nel *gouge* argillitico. Il dominio B, sviluppato entro i calcari giurassici, è caratterizzato da pieghe decimetriche rovesciate e sradicate, da una pervasiva foliazione tettonica, da faglie inverse e strutture S-C. Il dominio C è caratterizzato da strutture S-C. Tutte queste strutture indicano un senso di taglio verso est. Lo sviluppo della zona di taglio è avvenuto senza soluzione di continuità, mediante due principali stadi: il primo ha permesso lo sviluppo del dominio A deformando il livello argilloso immediatamente a contatto con le rocce calcaree sovrascorse, entro le quali si sono sviluppate pieghe. Il secondo stage ha amplificato la deformazione entro il dominio A, ha causato lo scollamento delle pieghe entro le Marne a *Posidonomya*, lo sviluppo di faglie inverse e delle strutture S-C. Durante la sviluppo della zona di taglio si è verificato una ripartizione della deformazione, principalmente controllata da: a) contrasto litologico, b) anisotropia litologica e c) circolazione di fluidi e processi di metamorfismo ad essi collegati. Lo sviluppo della foliazione tettonica è riferibile a meccanismi di *pressure solution* e *solution transfer*. Entro le rocce più intensamente deformate si è innescato un processo di metamorfismo molto basso che ha dato luogo allo sviluppo di minerali argillosi, quali l'illite-illite/smentite. Lo sviluppo di metamorfismo molto basso entro i livelli a deformazione concentrata è messo in relazione al locale incremento di temperatura dovuto alla deformazione unitamente alla circolazione di fluidi favorita dal meccanismo di *pressure solution*.

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Introduction

Deformational processes related to tectonic forces taking place in the crust may localise into zones of high strain characterised by shearing (Ramsay & Huber 1987; Bell et al. 1989; Brown & Solar 1998; Peacock 2002; Montesì & Hirth 2003; Wang & Ludman 2004). For this reason shear zones are common structures in deformed terrains, developed at different scales and crustal levels.

Most terrains experienced multiple deformational events during their tectonic evolution. In this light, superposed structures are the main record of their tectonic history. Detailed analyses on the shear zones in terms of geometry, kinematics, metamorphism and age may reveal much on their evolution and development conditions, providing very helpful information to evaluate the tectonic context in which they took place. Shear zones developed under metamorphic conditions are very useful for this purpose, because the changes in mineral assemblages and the resulting rocks fabrics better contribute to their study. Shallow shear zones within sedimentary rocks may develop by brittle-ductile or ductile deformation (Ramsay & Huber 1987), offering the opportunity for tectonic investigations in sedimentary terrains, difficult to perform in the surrounding less deformed rocks.

This paper deals with a contractional shear zone, up to 7 m thick, with heterogeneous strain distribution, developed at a shallow crustal level during the stacking of the Northern Apennines tectonic units. This shear zone (hereafter named Mt. Aquilaia shear zone) has been described within the Tuscan Nappe, the deepest sedimentary tectonic units of the Northern Apennines, exposed in the western side of the Mt. Amiata geothermal region (Brogi & Lazzarotto 2002; Brogi 2004a, b, c) (Fig. 1). This structure provides a good opportunity to analyse shearing of pelitic and calcareous rocks under conditions of the shallow crust, based on the deformation fabric and the structures found within the sheared rocks. This study is used to reconstruct the formation mechanism and evolution of the shear zone, the progressive shear model and displacement history. It also illustrates that strain concentration and fluid migration in sedimentary rocks may produce localised shear heating and ductile behaviour giving rise to very low-grade metamorphism also at shallow crustal levels.

Geological outlines

Geological features of the Northern Apennines

The Northern Apennines thrust belt formed by convergence and subsequent collision between the African (Apulia microplate) and European (Sardinia–Corsica massif) continental margins during the Tertiary (Carmignani et al. 2001 and references therein). This process resulted in the stacking of tectonic units (Fig. 2) which are, from top to bottom: (a) The Ligurian and the Subligurian Units, composed of remnants of Jurassic oceanic crust and its Jurassic–Cretaceous sedimentary cover

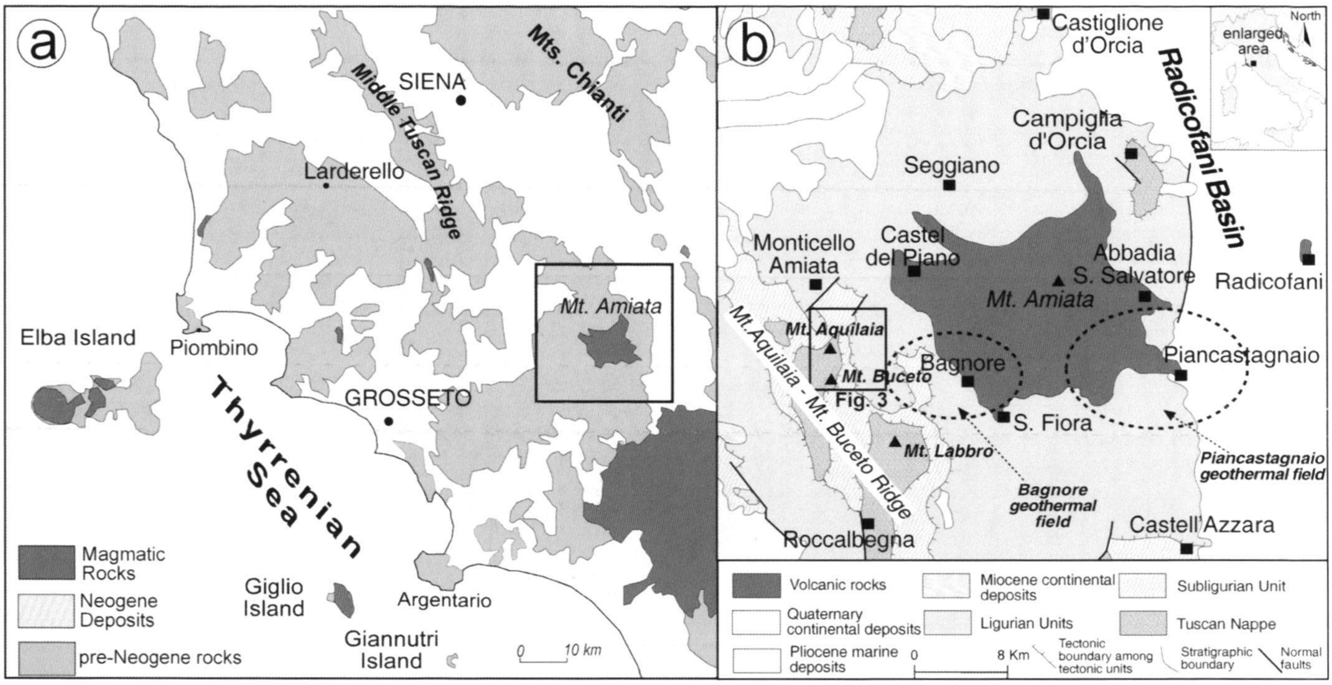

Fig. 1. a) Geological sketch-map of southern Tuscany. b) Geological sketch-map showing the Mt. Amiata geothermal region (see rectangle in (a) for location). The rectangle located on the southwestern side of the Mt. Amiata Volcanic Complex indicates the Mt. Aquilaia area, enlarged in Figure 3.

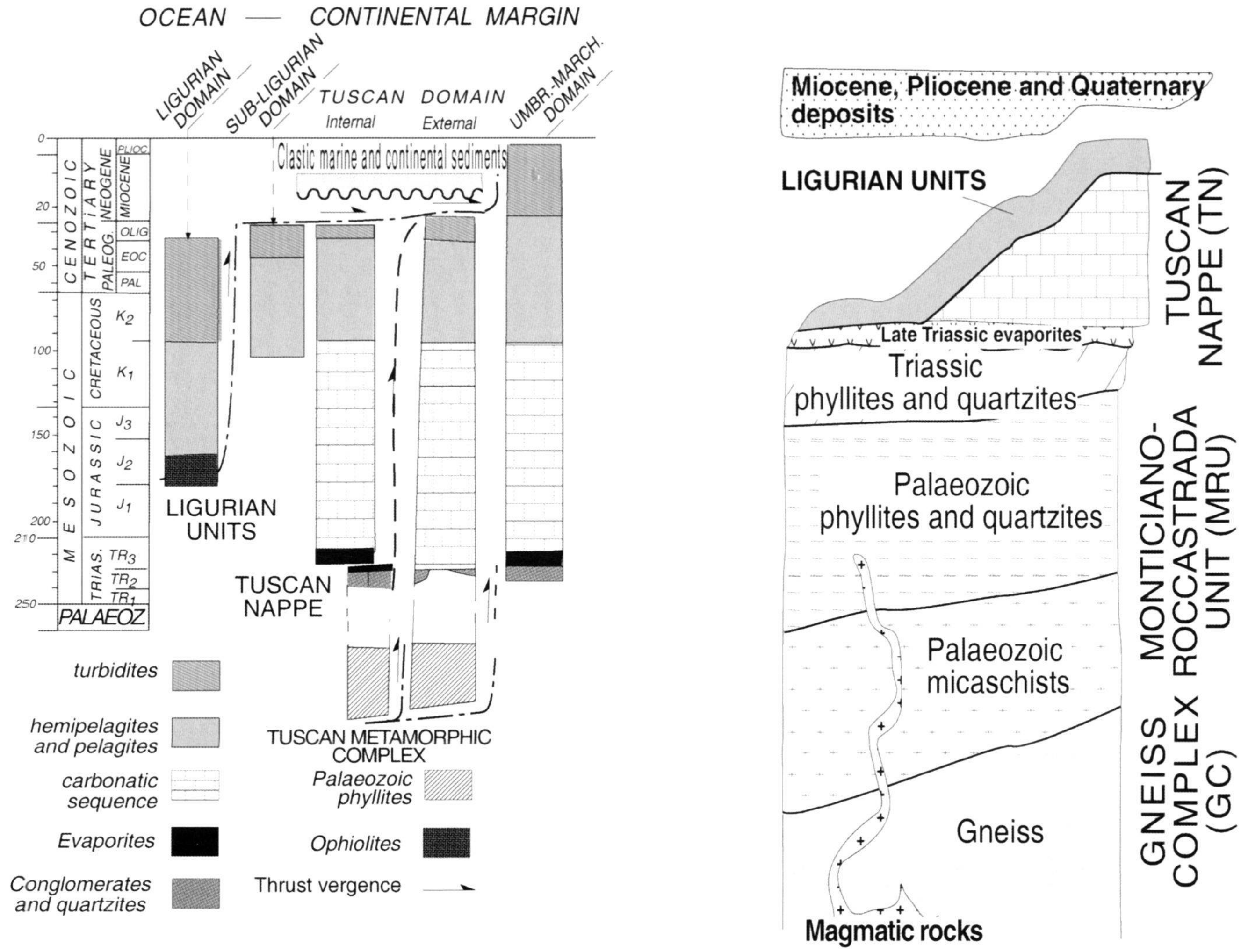

Fig. 2. Left: structural relationships amongst the different tectonic units of the Northern Apennines (from Carmignani et al., 2001). Right: tectono-stratigraphic units reconstructed by means of borehole data in the Larderello geothermal area (from Batini et al. 2003).

and Cretaceous–Oligocene flysch. These Units were thrust eastwards over the Tuscan Nappe during Late Oligocene–Early Miocene times. (b) The Tuscan Units, including sedimentary and metamorphic (HP-LT and greenschist facies) successions ranging from Palaeozoic to Early Miocene in age. The Tuscan sedimentary cover (Tuscan Nappe, including Late Triassic evaporites to Early Miocene turbidites) is thrust eastwards over the outermost Umbria-Marche units. In the hinterland of the Northern Apennines the metamorphic substratum (Tuscan Metamorphic Basement) of the sedimentary cover is mainly known through the drilling of geothermal wells (e.g. Larderello-Travale and Mt. Amiata geothermal areas), penetrating the crust down to about 4.5 km (Batini et al. 2003 and references therein). The Tuscan Metamorphic Basement consists of two units (Bertini et al. 1991): the upper Monticiano-Roccastrada Unit, composed of phyllites, quartzites and metacarbonates, and the lower Gneiss Complex (Fig. 2). After the emplacement of the tectonic units, extension affected the hinterland of the Northern Apennines (*i.e.* Northern Tyrrhenian Basin and Southern Tuscany) from the Early–Middle Miocene (Carmignani et al. 1994; Jolivet et al. 1999; Brunet et al. 2000). Extension is coeval with compression, which had been developing in the outer Northern Apennines, from the Early–Middle Miocene. The extensional tectonic process is well expressed in the structure of the Mt. Amiata geothermal area (Calamai et al. 1970), where extensional structures greatly modified the geometric relationships between the compressional tectonic units piled upon the Adriatic margin at the end of the collisional stage (Brogi 2004a, c).

Geology of the Mt. Amiata

The studied shear zone is located in a tract of the Mt. Aquilaia-Mt. Buceto ridge, a morpho-tectonic feature located on the

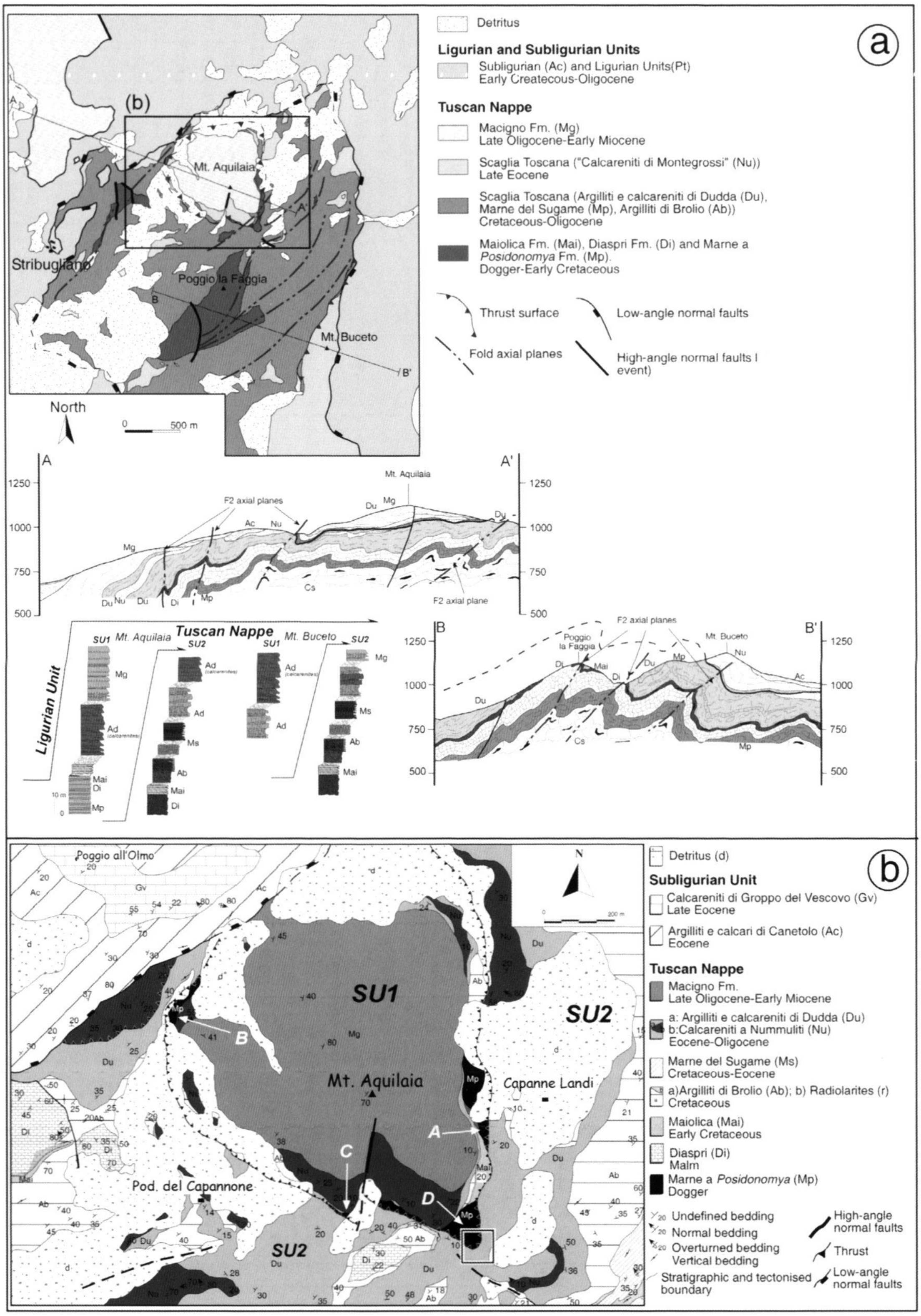

Fig. 3. a) Geological sketch-map of the Mt. Aquilaia – Mt. Buceto Ridge, geological cross-sections and relationships between the Tuscan Nappe subunits (SU1 and SU2) recognised in the Mt. Aquilaia and Mt. Buceto areas, as described in Brogi & Lazzarotto (2002) and Brogi (2004a). The rectangle indicates the detailed geological map given in b. b) Geological map of the Mt. Aquilaia shear zone. The letters A, B, C and D are discussed in the text.

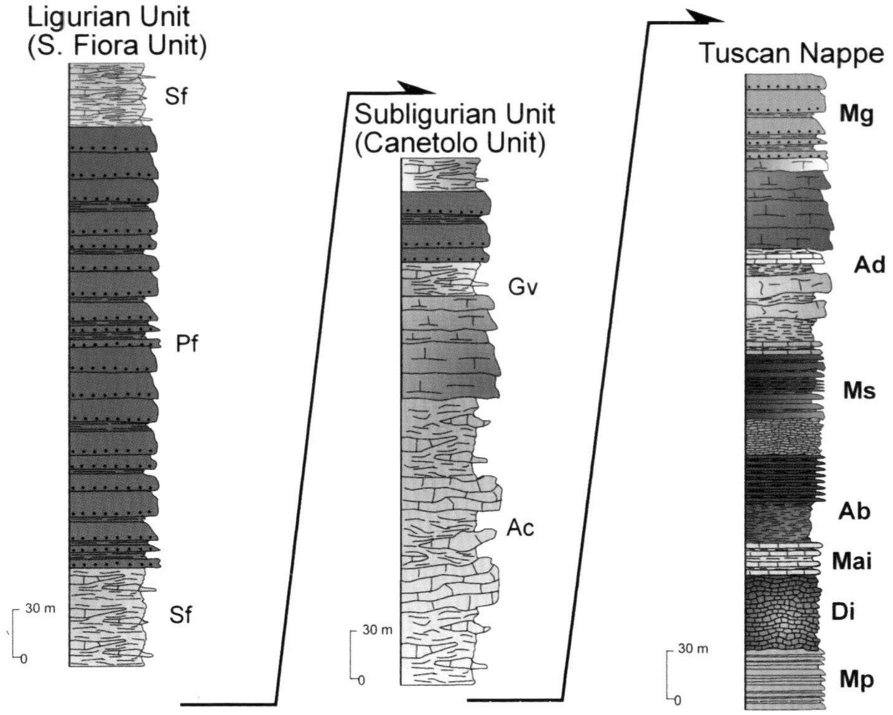

Fig. 4. Tectono-stratigraphic relationships between the Ligurian (S. Fiora Unit), Subligurian (Canetolo Unit) and Tuscan Nappe. The Tuscan Nappe is composed of: Mp – Marne a *Posidonomya* Fm., marls and marly limestones (Middle Jurassic); Di – Diaspri Fm., radiolarites (Late Jurassic); Mai – Maiolica Fm., cherty limestones (Early Createcous); Ab – Scaglia Toscana Fm., Argilliti di Brolio, clays, calcilutites with interbedd radiolarites (Early-Late Cretaceous); Ms – Scaglia Toscana Fm., Marne del Sugame, marls and calcareous marls (Paleocene-Eocene); Du – Scaglia Toscana Fm., Argilliti e calcareniti di Dudda, clays and interbedded calcarenites (Eocene-Oligocene); Mac – Macigno Fm., quartz-feldspar sandstones (Late Oligocene-Early Miocene). The Canetolo Unit (Subligurian Unit) is composed of: Ac – Argille e calcari Fm., clays, limestones and calcarenites (Eocene); Gv – Calcareniti di Groppo del Vescovo Fm., calcarenites and calcirudites (Late Eocene). The S. Fiora Unit (Ligurian Unit) is composed of: Sf – S. Fiora Fm., clays, limestones, marls and sandstones (Cretaceous); Pf – Pietraforte Fm., calcareous sandstones (Late Cretaceous).

western side of the Late Pleistocene Mt. Amiata volcano (Fig. 1). In the Mt. Aquilaia – Mt. Buceto ridge the Tuscan Nappe is widely exposed (Calamai et al. 1970) and surrounded by the Ligurian and Subligurian Units (Fig. 3). The Ligurian Units are represented by the S. Fiora and Ophiolitic Units, whereas the Subligurian Unit is only composed by the Canetolo Unit (Fig. 4). The stratigraphic succession of the Tuscan Nappe has been described by Fazzini (1978), Brogi and Lazzarotto (2002), Brogi (2004a, b) and Pandeli et al. (2005). It consists of, from the top to bottom: turbiditic sandstones and silts (Macigno Fm., Late Oligocene – Early Miocene); limestones, marls, clays and radiolarites (Scaglia Toscana Fm., Oligocene–Cretaceous); cherty limestones and marls (Maiolica Fm., Early Cretaceous); radiolarites (Diaspri Fm., Late Jurassic); marls and clays (Marne a *Posidonomya* Fm., Middle Jurassic). Different lithostratigraphic units have been recognised in the Scaglia Toscana Fm., similar to other areas of southern Tuscany (Canuti et al. 1965; Fazzuoli et al. 1996). They consist of: Calcarenti di Montegrossi (biocalcarenites and biocalcirudites, Late Eocene), Argilliti e calcareniti di Dudda (calcarenites, marls and clays, Middle–Late Eocene), Marne del Sugame (marls, Late Cretaceous–Eocene), and Argilliti di Brolio (clays and limestones – Late Cretaceous). The Tuscan Nappe succession experienced polyphase tectonics during pre-, syn- and post-collisional stages of the Northern Apennines (Pertusati et al. 1977). The detailed description of the tectonic history and related structures has been reported in Brogi and Lazzorotto (2002) and Brogi (2004c), and can be summarised as follows: the oldest deformational event (D_1) is characterised by brittle/ductile east-verging structures. These structures developed during the emplacement of the Ligurian Units on the Tuscan Nappe (Late Oligocene-Early Miocene) and consist of reverse faults, N-S striking folds (F_1) and tectonic foliation (S_1). The D_2 deformational event (Early-Middle Miocene) produced southeast-verging folds, metre to decametre in size, with axes ranging from N20° to N70°, and reverse faults, successively dissected by low-angle normal faults (Middle-Late Miocene). All these structures were deformed by Early/Middle Pliocene – Quaternary high-angle normal and transtensional faults.

The studied shear zone represents a detachment separating two Tuscan Nappe subunits emplaced during the D_1 deformational phase (SU1 and SU2 subunits described in Brogi and Lazzarotto, 2002; Brogi, 2004a). The upper one (SU1) consists of a stratigraphic unit between the Marne a *Posidonomya* and the Macigno Fms., whereas the lower one (SU2) consists of the stratigraphic units between the Diaspri and the Macigno Fm. (Fig. 3).

The Mt. Aquilaia shear zone

Geometrical setting

The Mt. Aquilaia shear zone mainly affected Jurassic marly rocks of the Marne a *Posidonomya* Fm. and clayey rocks of the Scaglia Toscana Fm. (Brogi and Lazzarotto, 2002). This structure, about 7 metres thick, developed during the thrusting of the Marne a *Posidonomya* Fm. on the clayey lithotypes of the Scaglia Toscana Fm. (Fig. 3).

Fig. 5. Panoramic view of the shear zone, as observed in the south-eastern side of the Mt. Aquilaia, indicated by "D" given in figure 3. The black rectangle is enlarged in figure 9.

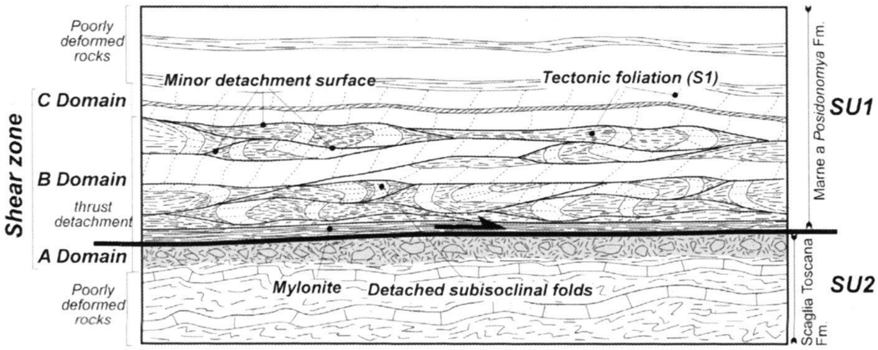

Fig. 6. Unscaled cartoon showing the three domains of the Mt. Aquilaia shear zone. See the text for explanation.

The studied shear zone is broadly exposed mainly on the south-eastern side of the Mt. Aquilaia (D point in Figure 3), where it shows a subhorizontal attitude (Fig. 5). This shear zone may be subdivided into three domains typified by different fabrics, structures, and deformation features (Fig. 6). The deepest domain (A-domain), consisting of about 1m thick of clayey gouge occurring below the thrust decollement surface, developed within the clayey rocks of the footwall block. The middle domain (B-domain), about 4 m thick, is the most extensive shear zone domain. It affected the marly rocks of the hangingwall block located on the thrust decollement. The uppermost domain (C-domain), about 2 m thick, affected the marly rocks and forms the transition to the unsheared and weakly deformed rocks.

This shear zone has also been described for the Mt. Buceto area, where it is characterised by a clayey fault gouge (up to 2 m) separating the arenaceous lithotypes of the Macigno Fm. from the clayey ones of the overthrusted Scaglia Toscana Fm. (Brogi 2004a). No shear domains have been recognised; the deformation concentrated only within a fault gouge composed of clayey and arenaceous much disrupted lithotypes.

Shear zone fabrics and kinematics

The A-, B- and C-domains are characterised by different structures and fabrics. The structures, micro- to map-scale, confirm the top-to-the-east shearing in all the three domains. Shearing was partitioned within the shear domains, mainly in response to the rheological properties of the involved lithotypes. The features of the three shear zones are as follows:

A-domain – This domain is a highly damaged zone composed of a clayey gouge in which primary sedimentary features and rock fabrics have been completely obliterated. Limestone clasts of centimetre and rarely decimetre size are dispersed and randomly distributed within the clayey "matrix" (Fig. 7). These calcareous elements correspond to disrupted limestone beds strongly involved in the shearing. No kinematic indicators, such as rotated clasts, S-C structures, shear bands and folds have been observed.

B-domain – This domain is characterised by strongly deformed rocks, mainly affected by a widespread pervasive tectonic foliation (S_1) accompanying folds (F_1), reverse faults and detachments (Fig. 8). The development of these structures was strongly influenced by the lithology of the rocks involved in the deformation, which belong to a multilayered succession characterised by decimetre beds of calcareous dominant lithotypes interbedded with centimetre thick marly clays. The basal part of the B domain locally coincides with a mylonite zone with variable thickness and laterally discontinuous: about 2 m on the western side of Mt. Aquilaia (point B in Fig. 3), about 1 m on the southern side of Mt. Aquilaia (point C in Fig. 3) and up to 20 cm on the eastern side (point D in Fig. 3). The

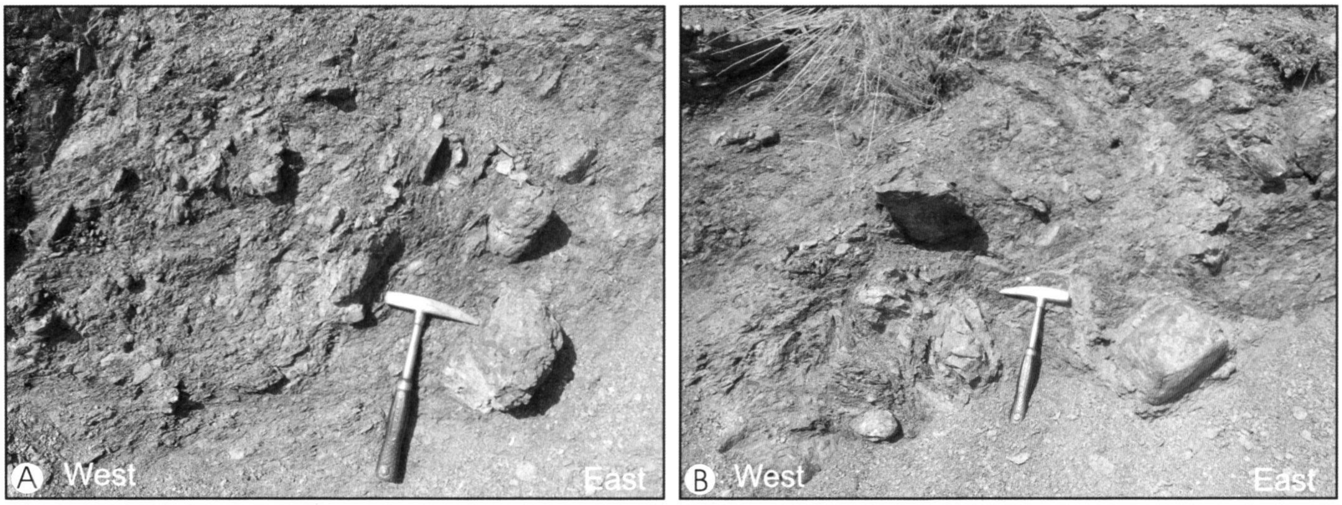

Fig. 7. Very disrupted calcareous beds within a clayey gouge occurring in the A-domain of the Mt. Aquilaia shear zone. The deformed rocks with chaotic fabric belong to the Scaglia Toscana Fm. (Argilliti e calcarentiti di Dudda Fm.), the uppermost formation of the SU1 subunits.

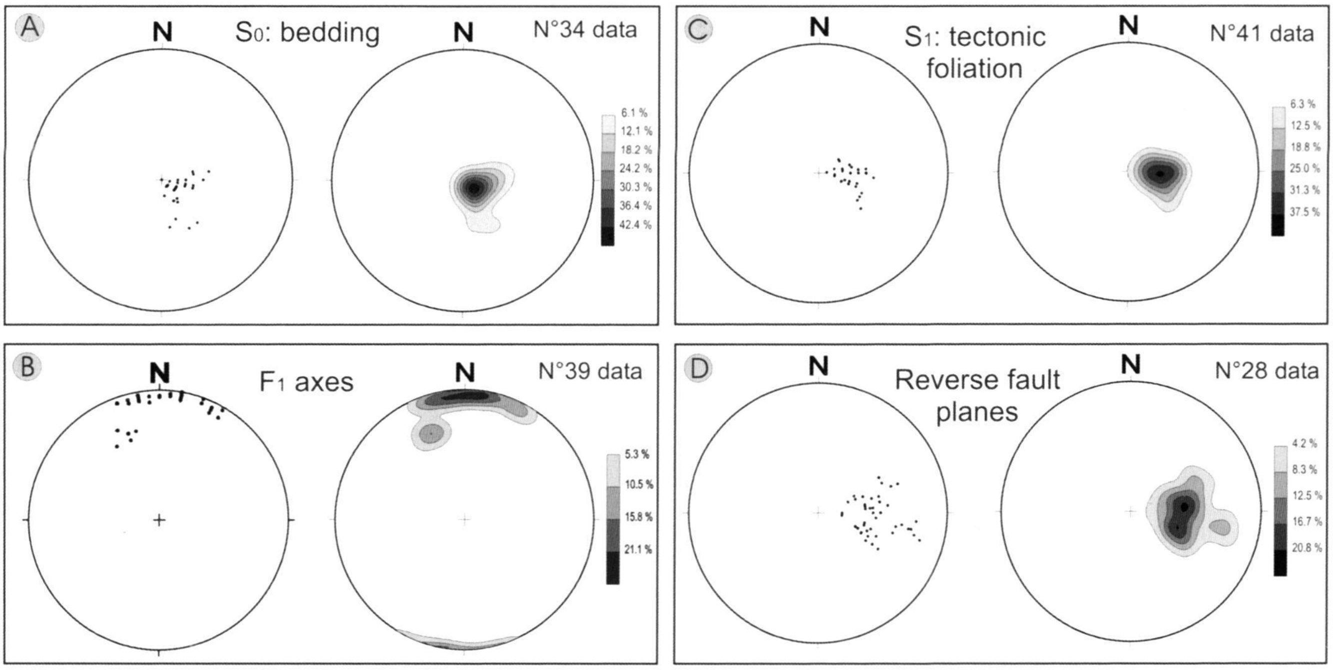

Fig. 8. Stereoplots (lower hemisphere, Schmidt diagram) of primary fabric and meso-structures within the Mt. Aquilaia shear zone. A) poles and contouring of bedding planes (34 data); B) F_1 fold axes and related contouring (39 data); C) poles and contouring of reverse fault planes (28 data); D) poles and contouring of axial plane S_1 tectonic foliation (41 data).

mylonite zone is characterised by foliated and lineated marly rocks showing millimetre lenses and layers, laterally discontinuous, which define the main planar fabric element.

The S_1 foliation, normally west-dipping, frequently exhibits a subparallel attitude with respect to the bedding (hereafter S_0) in the pelitic dominant lithotypes (Fig. 9), whereas in the calcareous rocks S_1 intersects S_0 at high angles (Fig. 10). The S_1 morphology and distribution depends on the affected lithotypes: this is typically a strongly penetrative foliation within the pelitic lithotypes (slaty-cleavage) which, locally, obliterated the primary sedimentary fabric. In contrast, the S_1 foliation within the calcareous beds is a stylolytic cleavage, defined by

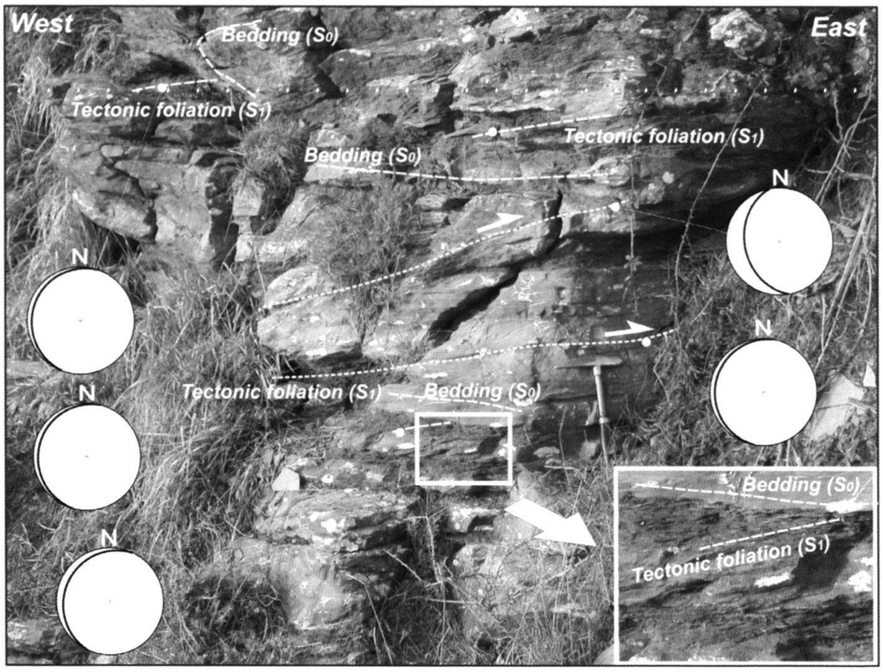

Fig. 9. Details of the B-domain and related structures occurring on the south-eastern side of the Mt. Aquilaia. Enlarged zone is indicated in figure 5.

millimetre thick and little deformed lithons, separated by dissolution surfaces with very wriggly shape, up to 1 mm thick, in which mainly Fe-oxides and clayey minerals concentrated (Fig. 10c). F_1 folds, decimetre in size, are detached and asymmetric, and show interlimb angles ranging between 5° and 30°, resulting in subisoclinal to tight folds (Ramsay 1967; Ramsay and Huber 1987) (Fig. 10b). Their hinge lines mainly strike N-S, mostly plunge to the N, both vergence and facing (*sensu* Bell 1981) being east-directed. F_1 folds are concentred within decimetre thick horizons, delimited at the top and the base by detachments defining strongly sheared levels. The F_1 axial planes are sub-parallel or very gently inclined to the west with respect to detachment horizons activated along bedding surfaces in response to lithological contrasts. On the detachment surfaces, kinematic indicators consisting of very well developed striations are orthogonal to the orientation of the F_1 hinge lines (see stereonets in Figs 8 and 10).

Within the calcareous beds, reverse faults developed at a low angle with respect to S_0 (Fig. 10a). Such structures show west-dipping fault planes, decimetre displacements and east-directed tectonic transport. They are typified by flat-ramp-flat geometries: the flats coincide with the pelitic interstrata which are connected by ramps crosscutting the calcareous beds. Centimetre-scale drag-folds occur mainly in the footwall blocks. Calcite veins infilling millimetre thick cracks, geometrically coherent with shear strain, developed both in the footwall and in the hangingwall blocks (Fig. 10a). Furthermore, millimetre to centimetre scale en-echelon calcite veins developed within calcareous beds, mainly orthogonal with respect to S_1.

C-domain – This domain is mainly characterised by S-C structures developed within the pelitic rocks. These structures define minor centimetre and decimetre thick shear zones. The s surfaces, gently west-dipping, intersect S_0 at angles ranging from 10° to 30°. C-surfaces correspond to detachment planes parallel to the bedding surfaces which developed at the transition between calcareous and pelitic strata. The intersection between S and C planes is parallel to the F_1 fold axes occurring within the B domain. In addition, widespread striation developed along C planes. These kinematic indicators, mainly consisting of slickensides, are coherent with the direction of tectonic transport of both reverse faults and folds occurring in the B-domain (see stereonets in Figs 8 and 10).

Microstructures

Microstructural analyses have been carried out on samples belonging to the Marne a *Posidonomya* Fm., collected both within the shear zone and in the adjacent poorly deformed areas. The analysed samples were W-E oriented, normal with respect to the S_1 and S_0 foliations and F_1 hinge lines. A few deformed samples collected outside the shear zone are typified by a fine-grained wackstone containing very small-grain size phyllosilicates. Skeletal remains consisting of calcite filaments up to 100 μm in length are aligned along the bedding surface (Fig. 11a). Micritic pebbles with elliptical shape and variable dimensions are randomly distributed within the rocks.

Samples of marly rocks collected within the shear zone are mostly characterised by two different types of foliations: (1) in the calcareous rocks a spaced, disjunctive cleavage, with wriggly and/or stylolytic shape of cleavage domains (*sensu* Passchier & Trouw 1996 and references therein.) has developed (Fig. 11b); (2) in the pelitic dominant rocks, a

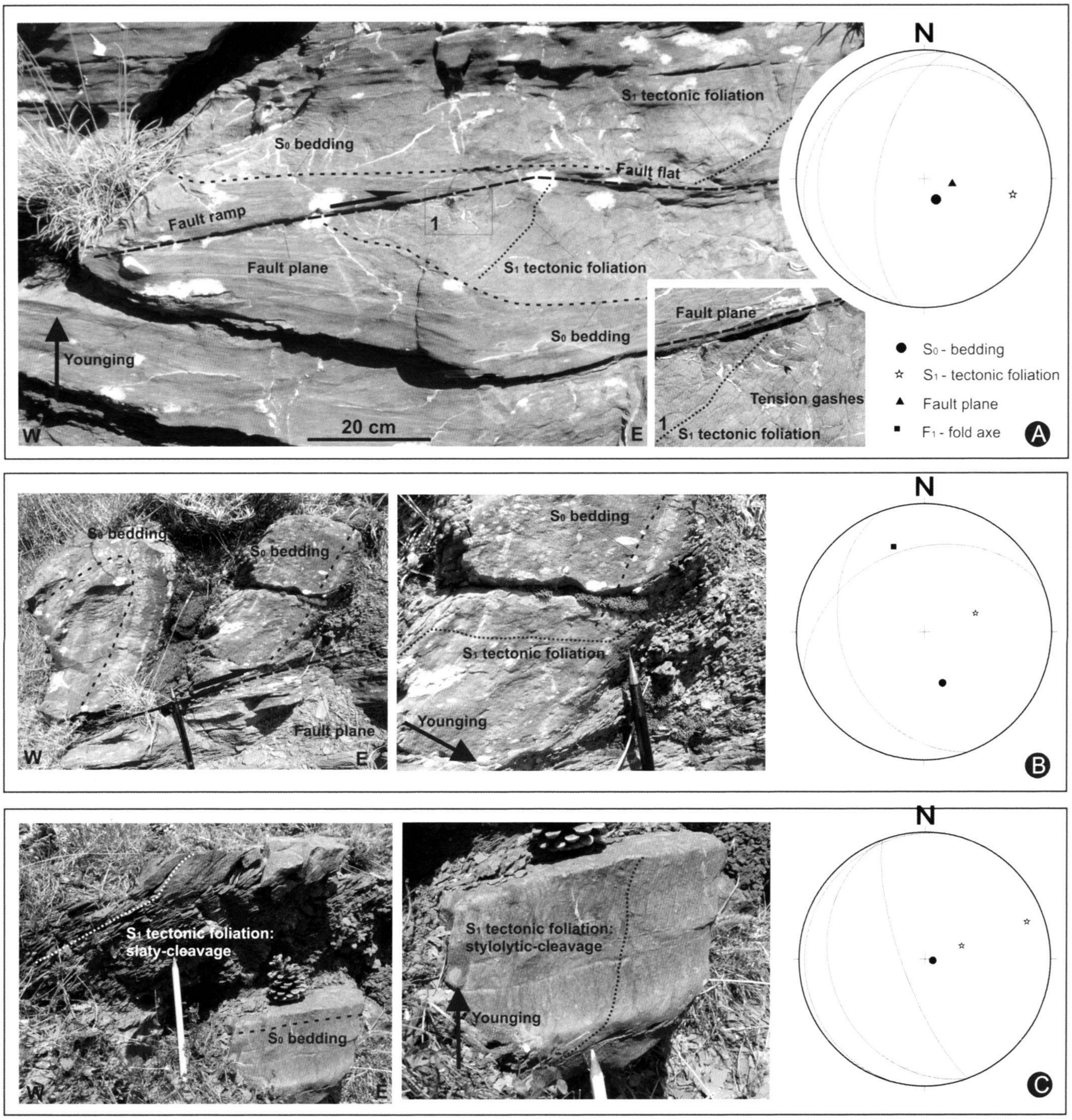

Fig. 10. Examples of east-verging meso-structures occurring within the Mt. Aquilaia shear zone (Marne a *Posidonomya* Fm.). a) Reverse fault with related S_1 tectonic foliation and tension gashes; b) Detached overturned F_1 folds and their axial planar foliation S_1; c) S_0/S_1 relationships. The angle between the tectonic foliation and the bedding surface in the fold limbs depends on the lithotype affected by the foliation: higher angles occur in marly limestones, lower angles in pelitic lithotypes.

spaced disjunctive cleavage, with rough and/or smooth cleavage domains (*sensu* Passchier & Trouw 1996 and references therein) is found (Fig. 11c). In both cases two domains characterise the deformed rocks: (a) the cleavage domains and (b) the microlithons, comprised between two cleavage domains (Figs 11b and 11c). The microlithons contain fabric elements oblique with respect to the cleavage domains and representative of a sedimentary fabric, such as skeletal re-

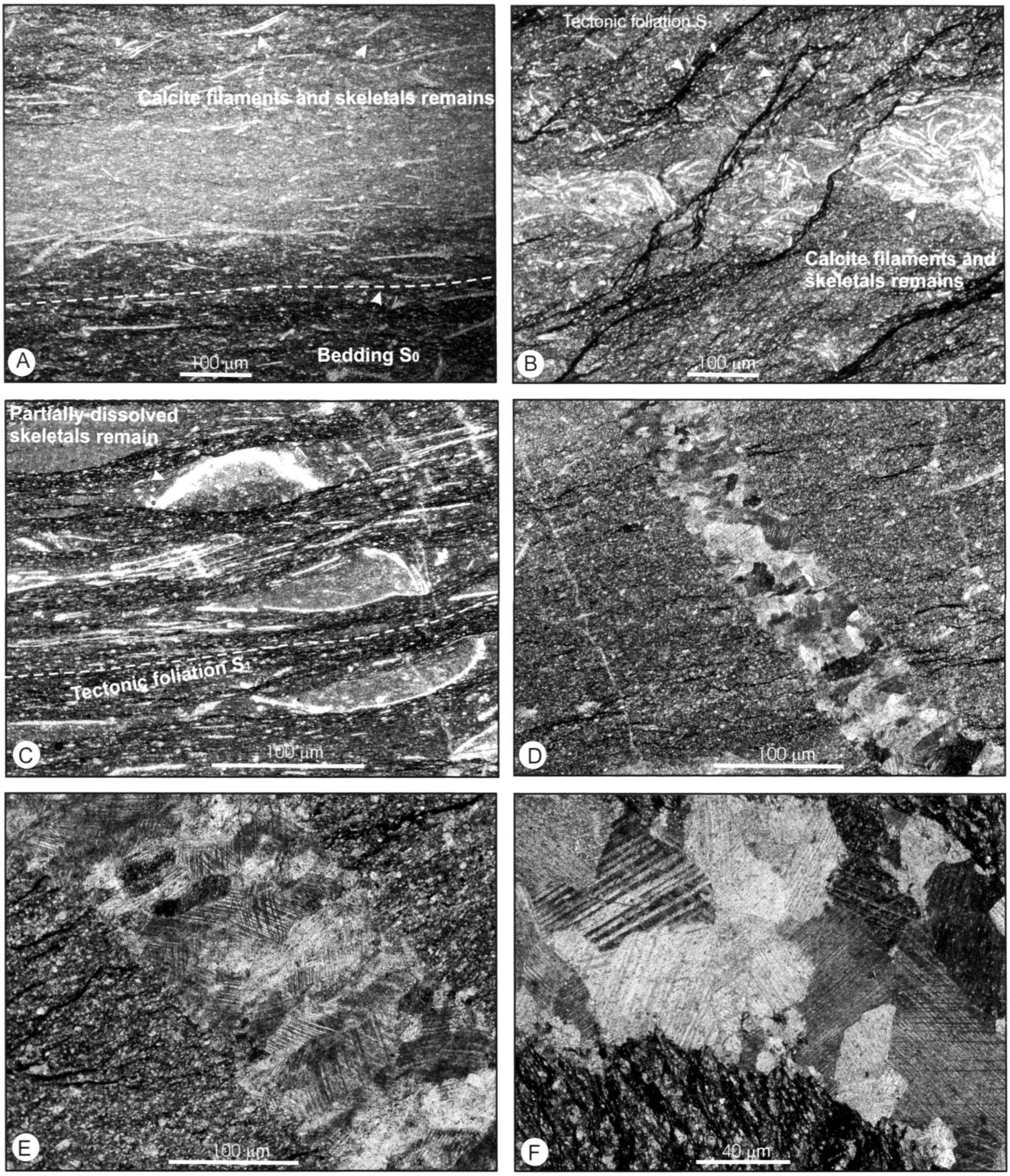

Fig. 11. Micrographs of Marne a *Posidonomya* marly samples. A) Poorly deformed sample where fossils remains are parallel to bedding. B) Tectonic foliation (S_1) developed at a high angle with respect to the bedding in a calcareous lithotype; S_1 is a stylolytic cleavage with very thin cleavage domains. C) Tectonic foliation (S_1) developed at a low angle with respect to the bedding in a marly (pelitic dominant) lithotype. D) Calcite vein perpendicular to the S_1 developed during cleavage formation. E) Calcite crystals within calcite veins showing twin-lamellae consistent with the typeII and III described by Burkhard (1993).

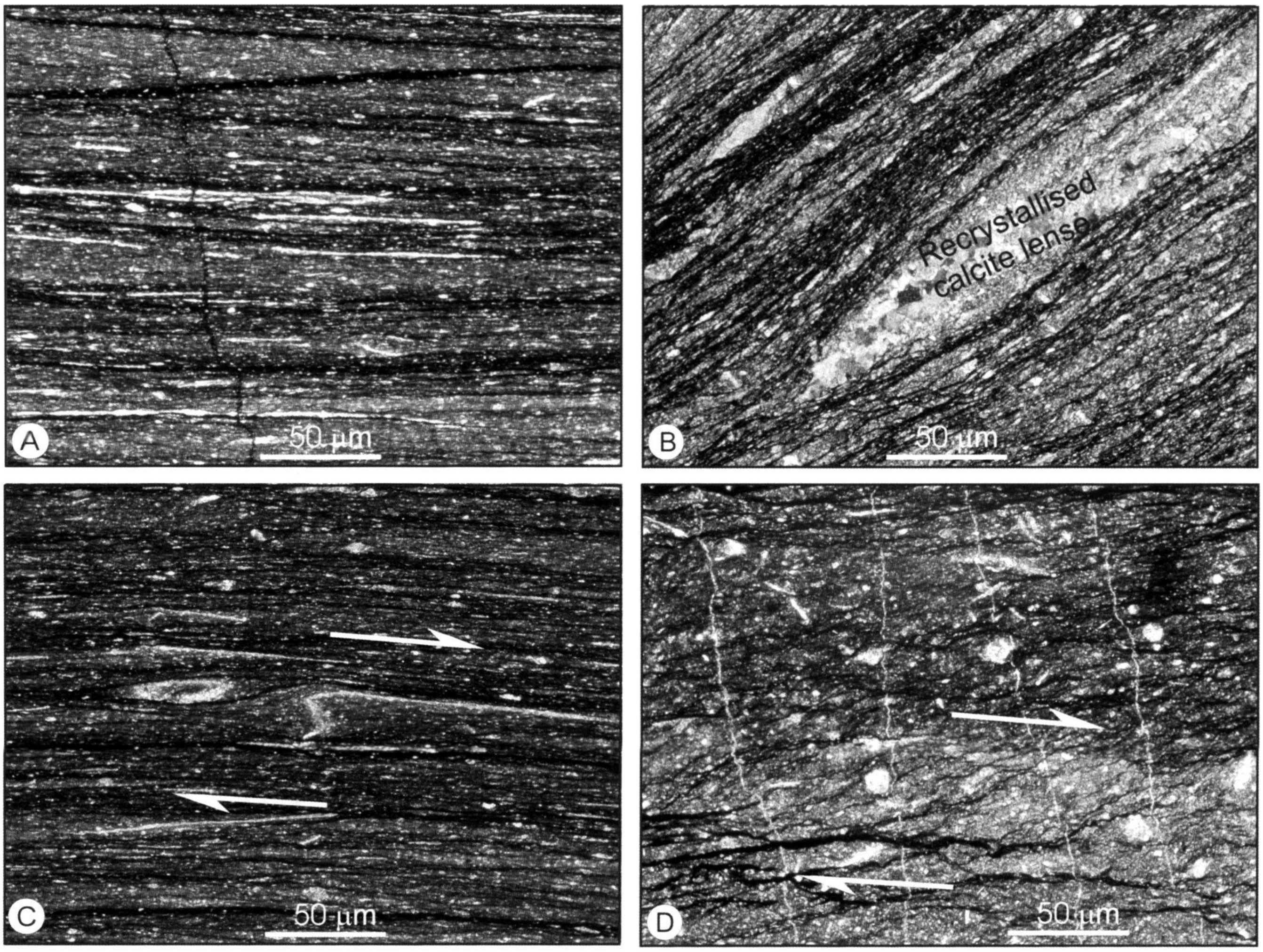

Fig. 12. Pelitic dominant levels and recristallysed calcite lenses defining the mylonitic foliation (A and B) developed at the base of the B-domain of the shear zone. Shear indicators (C and D) are in agreement with the general top-to-east shearing.

mains and lithological changes (Fig. 11b). The cleavage domains are planar elements ranging from 5–200 µm in thickness, defining the rock fabric. Clay minerals and oxides concentrated along the cleavage domains which formed by pressure solution and solution transfer mechanisms and which define the foliation both in the calcareous and the pelitic lithotypes. Pressure solution led to the accumulation of insoluble material along the dissolution surfaces which define the cleavage domains. The dissolved material mainly consists of calcite and minor quartz, transferred by fluids and crystallised within veins mainly orthogonal to the cleavage domains (Fig. 11d). Within the veins, the calcite crystals range from 10 to 60 µm and show *e*-twin-lamellae (Barber and Wenk 1979; Ferril 1991; Burkhard 1993) which define rhomboids (Fig. 11e). The thickness of a single lamella is, in general, greater than 1 µm (Fig. 11d). In some cases, the twins are gently curved mainly in the external border. These features characterising the twin lamellae are consistent with the types II and III as defined by Burkhard (1993).

High-strain zones are characterised by a mylonitic foliation (Fig. 12a) observed in the basal part of the B domain. In these cases, S_1 is defined by a spaced foliation characterised by alternating layers and lenses up to 50 µm thick, with different mineral composition, in which porphyroclasts, such as calcite skeletal remains and calcite aggregates, were strongly stretched (Figs 12a and 12b). These produced platy aggregates and elongate domains which are totally, or partly, dynamically recristallised (Fig. 12b), consisting of calcite and phyllosilicate dominant layers. Shear structures accompanied both the development of the mylonitic foliation (Fig. 12c) and the cleavage domains (Fig. 12d). In particular, σ-type porphyroclasts, both consisting of micritic aggregates and/or skeletal remains and indicating top-to-the-east shearing, are widespread (Fig. 12d). Pressure solution and solution transfer produced the rotation

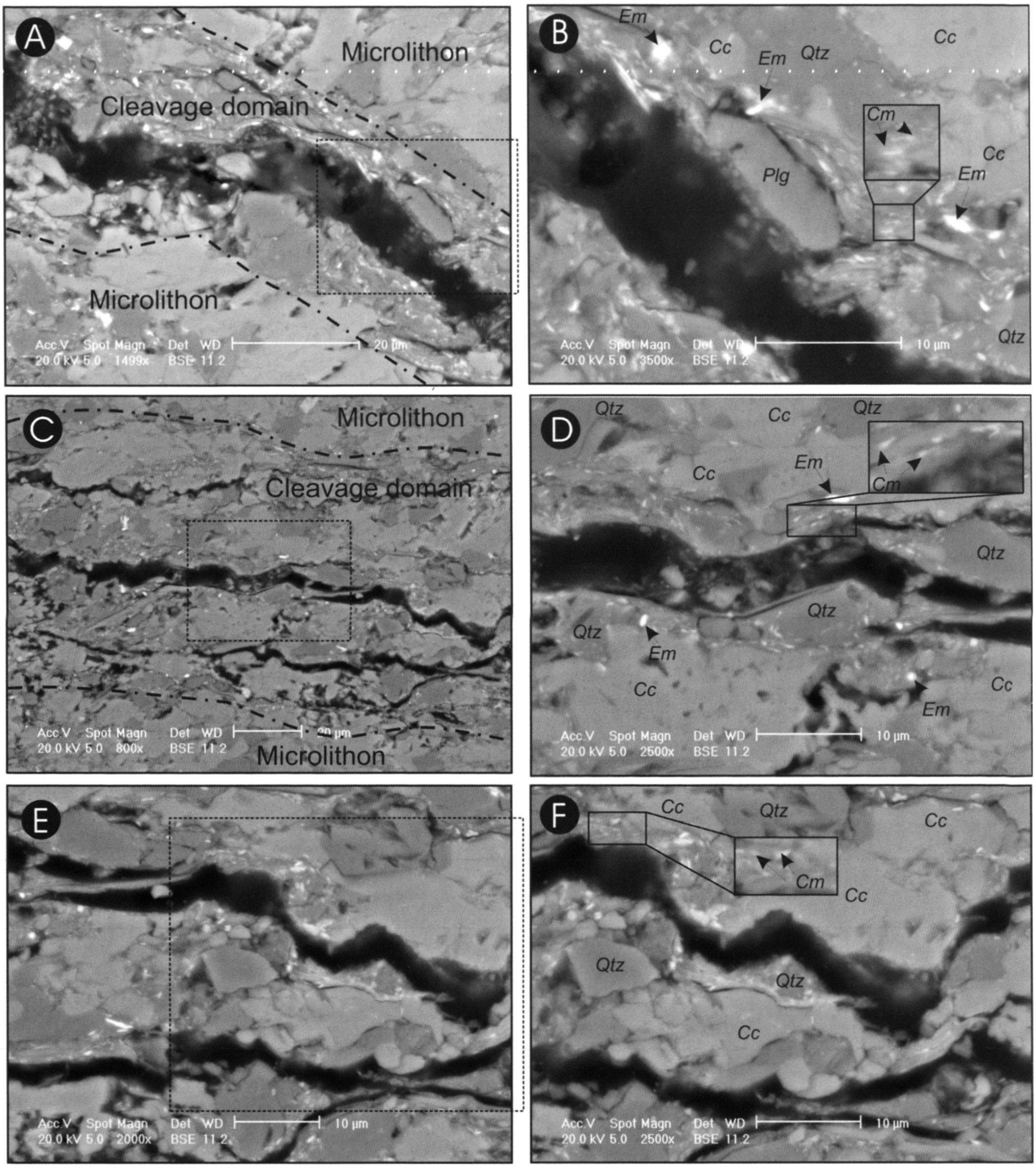

Fig. 13. SEM microphotographs of a marly sample belonging to the Marne a *Posidonomya* Fm. The dark layers correspond to the pressure-solution cleavage surfaces. New clay minerals (Cm) consisting of very small grains define the mineralogical assemblage within the cleavage domains. Grains with different size define the cleavage domains and the microlithons; small grains occur in the cleavage domains. The rectangles in figures A, C and E correspond to the enlarged photos in B, D and F. Keys: Qtz – Quartz; Cc – calcite; Plg – plagioclase; Em – hematite; Cm – white mica (illite-illite/smectite mixed layers).

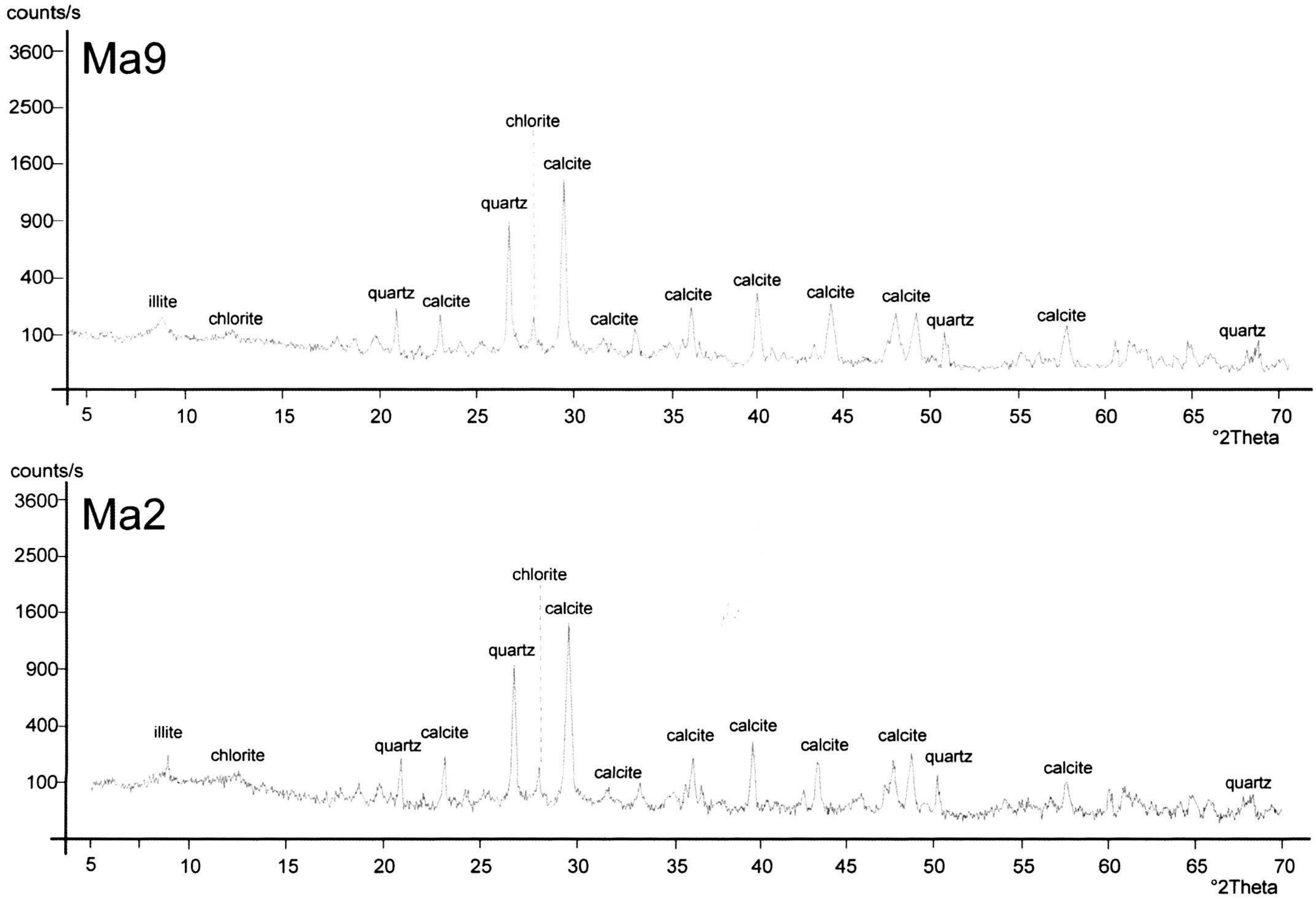

Fig. 14. X-ray diagrams of two representative marly samples (Ma2 and Ma9) belonging to the Marne a *Posidonomya* Fm. collected (a) within the B domain of the shear zone (Ma2), and (b) in the poorly deformed horizons not involved in the shearing (Ma9). Note the very similar mineralogical composition and the different amplitude and width of the illite peak, located at about 8° of incidence irradiation.

of elongate minerals, such as chlorite flakes and rare feldspars, due to selective solution and redeposition of surrounding material. SEM observations (Fig. 13) highlight that cleavage domains are mainly composed of small size crystals, consisting of: calcite, quartz, oxides and illite-illite/smectite mixed-layers (Reynold & Hower 1970), these latter being 0.5–2 μm in width. These phyllosilicates (Cm in Fig. 13) are localised only within the cleavage domains whereas they are absent within the microlithons. Illite-illite/smectite grains are elongate crystals showing a preferred orientation parallel or subparallel to the cleavage domains (Fig. 13). The microlithons are mainly composed of calcite, quartz, feldspar, oxides and phyllosilicates (mainly chlorite). On the whole the grain size within the microlithons is greater than within the cleavage domains. For example, calcite grains range between 10 to 40 μm within the microlithons but do not exceed 10 μm in the cleavage domains. Similar considerations apply for quartz grains which range from 20 to 50 μm in the microlithons and do not exceed 15 μm within the cleavage domains (Fig. 13).

Illite crystallinity analyses

X-ray analyses reveal that the analysed rocks are mainly composed of calcite, quartz and clay minerals, such as illite-illite/smectite and chlorite (Fig. 14). The clay fraction of the pelitic and calcareous rocks involved in the shearing and belonging to poorly deformed rocks from outside of the shear zone is always composed of illite-illite/smectite and chlorite. In order to reveal the possible presence of very low-grade metamorphism affecting the foliated lithotypes, illite crystallinity (IC) analyses have been conducted.

Fourteen representative samples have been collected close to and within the shear zone. Six of these consist of pelitic lithotypes (clayey marls, MA 1–6) collected in the hinges of the F_1 folds occurring within the B-domain, where a very pervasive S_1 tectonic foliation occurs (Fig. 15). Another six samples (MA 7–12) consist of apparently poorly deformed rocks (clayey marls) affected by the S_1 foliation and occurring within B and C-domains. Another two samples (MA 13–14), com-

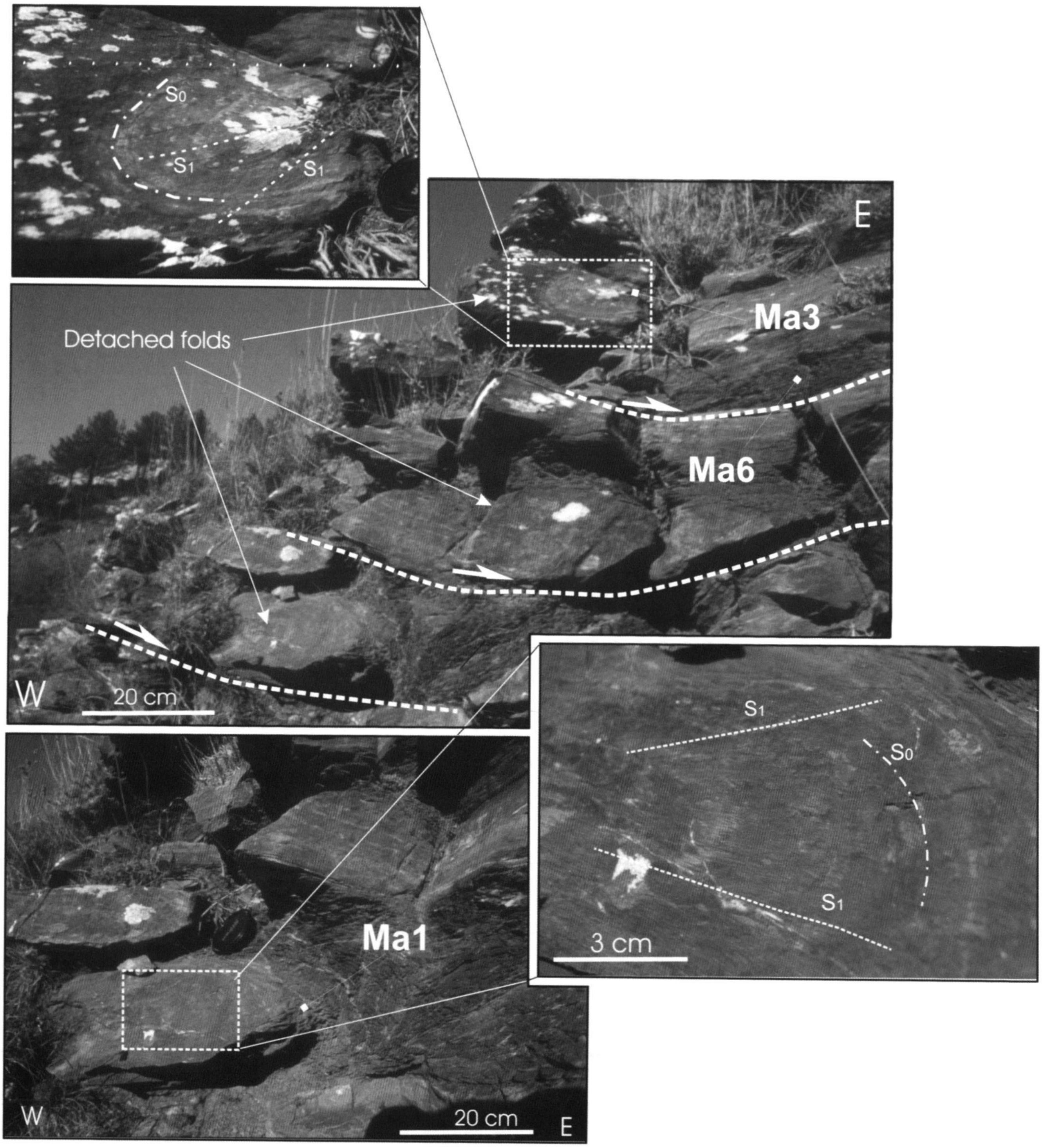

Fig. 15. Location of some samples collected for the IC analyses. Several samples have been collected in the hinge zones of detached and sub-isoclinal folds pervasively affected by S_1 foliation, as it is the case of the MA1 and MA3 shown in this figure. Other samples have been collected within minor shear zones and detachments, as it is the case of the MA6 sample given in this figure.

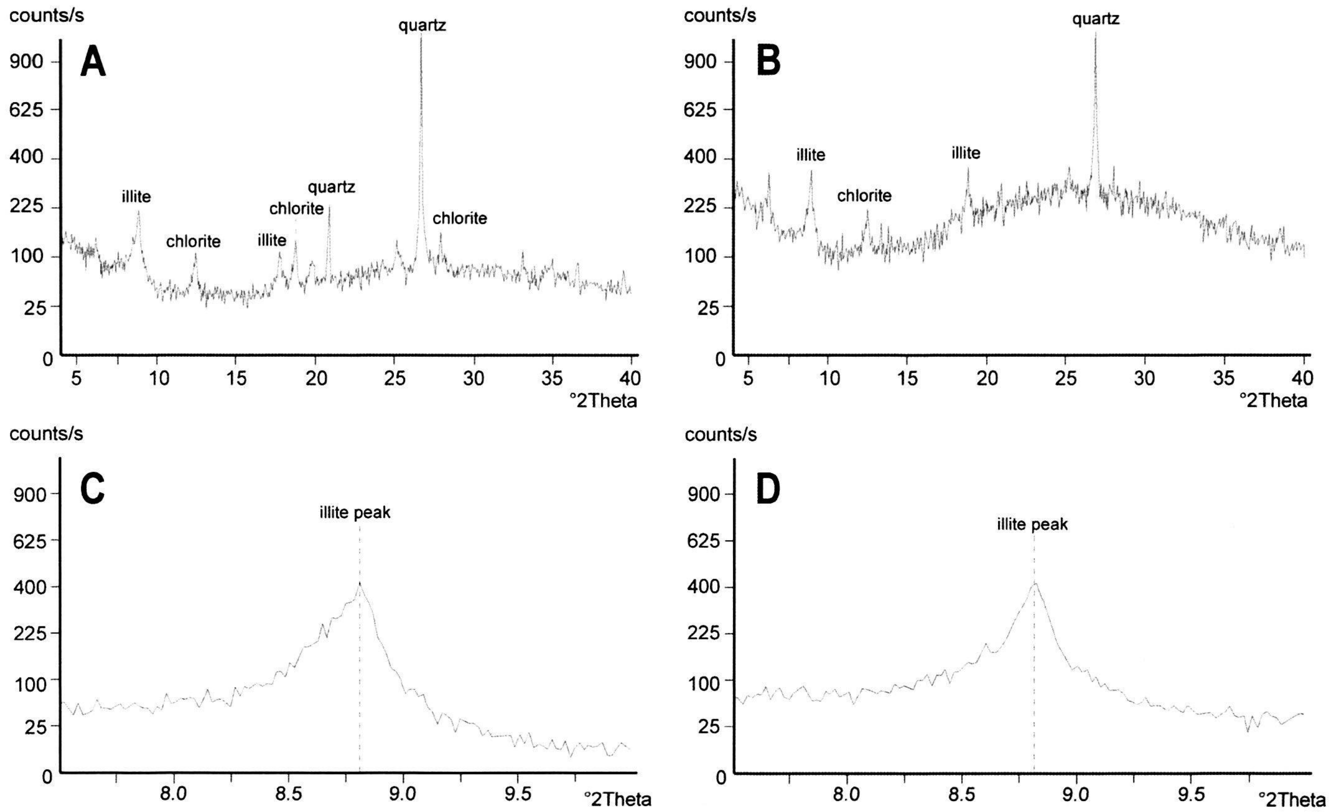

Fig. 16. X-ray analyses performed on the MA1 sample collected within the B domain of the Mt. Aquilaia shear zone. A) diagram showing the composition of the <64 μm fraction obtained after decarbonatation of the sample. B) diagram showing the composition of the <2 μm fraction. C) Detail of the illite peak of the B diagram. D) Illite peak after glycolation, utilised for IC analysis.

posed of the same lithotypes, have been collected from outside the shear zone. All samples belong to the Marne a *Posidonomya* Fm.

The analytical procedures for the IC analyses were those of Kubler (1984) and Kisch (1991). The samples were powdered and the powder was decarbonatated through a gentle HCl solution (15%). At this point, the carbonate matrix was removed and the insoluble minerals were concentrated (Fig. 16). Several washing procedures using deionised water permitted to get back to a neutral pH. The obtained decarbonatated powder was successively suspended in deionised water in order to make a grain size separation. In fact, as suggested by SEM analyses (Fig. 13), illite grains in the sampled rocks are mainly <2μm in size and hence, this grain fraction is suitable for IC analysis. These grain sizes were obtained by letting the shaken suspension settle for 6 hours and pipetting the topmost 5 cm. The aqueous suspension containing the <2μm grain-size fraction was dried on a glass slide at 40°C before diffractometric analyses were performed. These were repeated after glycolation (Fig. 16).

The mineralogical association of the decarbonatated samples is mainly characterised by quartz, illite, illite/smectite and chlorite (Fig. 16). The presence of illite/smectite was detected by the asymmetry of the 10 Å peak (Fig. 16). Glycolation was used to differentiate between illite and illite/smectite.

The obtained $\Delta 2\theta$ illite crystallinity (Kubler Index, hereafter KI) includes values ranging from 0.6 to 0.22 (Table 1). Pervasively foliated samples mainly show lower values of KI. The statistical error of a single KI measurement was estimated to be ± 0.03. The results obtained by the IC analyses are summarised in Table 1 which highlights that the KI of samples MA 1–6 ranges between 0.22 and 0.39; for samples MA 7–12 between 0.49 and 0.53; and finally for samples MA13–14 between 0.57 and 0.60. According to the relationships between KI and very-low grade metamorphic facies, as proposed by Kubler (1984) and also given in table 1, these values correspond to the anchimetamorphic zone (samples MA 2–6) and to the highest grade diagenetic zone (other samples), respectively. The sample Ma1 belongs to the epizone.

Discussion and conclusions

The Mt. Aquilaia shear zone affected sedimentary rocks and developed at a shallow crustal level. This structure experi-

Table 1. Illite crystallinity values and Kubler Index characterising the metamorphic zones.

Sample	*Density mg/cm2*	$\Delta 2\theta$
MA 1	1,29	0,22
MA 2	0,95	0,3
MA 3	2,1	0,35
MA 4	0,7	0,33
MA 5	0,91	0,28
MA 6	1,47	0,39
MA 7	1,53	0,49
MA 8	1,78	0,53
MA 9	0,82	0,51
MA 10	1,03	-
MA 11	0,9	0,6
MA 12	0,77	0,51
MA 13	2,3	0,57
MA 14	1,12	0,6

Zones	**Kubler Index (KI)**
High grade diagenetic zone	$\Delta 2\theta > 0.60$
Highest grade diagenetic zone	$0.60 > \Delta 2\theta > 0.42$
Anchimetamorphic zone	$0.42 > \Delta 2\theta > 0.25$
Epimetamorphic zone	$\Delta 2\theta < 0.25$

enced intense ductile and brittle/ductile shear deformation during the stacking of the Northern Apennines. In particular, shearing took place within the Tuscan Nappe during the thrusting of the Ligurian Units (Late Oligocene-Early Miocene). At present, the studied shear zone corresponds to a thrust zone separating two Tuscan Nappe subunits described in Brogi & Lazzarotto (2002). Fortunately the structures related to the shearing have been preserved, allowing to better investigate the structural features of this shear zone. The kinematic indicators occurring within the sheared rocks agree with the eastward emplacement of the tectonic units during the collisional stage (Carmignani et al. 2001 and references therein). In particular, the inferred thrust vergence is about N°85 (Figs 8 and 10). Consideration on the growth of the structures occurring in the three different shear zone domains coupled with their interference patterns, indicate that the evolution of the shear zone was characterised by two main uninterrupted stages of a single episode of deformation (D_1 in Brogi & Lazzarotto 2002). Two different hypotheses can be proposed (Fig. 16):

Hypothesis a) According to hypothesis (a) the main detachment horizon developed during a first stage (Fig. 17) along the contact between the clayey rocks of the Scaglia Toscana Fm. and the calcareous and marly lithotypes of the Marne a *Posidonomya* Fm. This gave rise to the highly damaged zone of the A-domain within the clayey rocks (Fig. 7). The detachment developed contemporaneously with asymmetric folds which affected the Marne a *Posidonomya* calcareous rock, described for the B-domain and occur in the hangingwall block. The progressive deformation (second stage) (1) amplified the damage within the A-domain; (2) caused the detachment of the previously developed asymmetric folds within the B-domain; (3) produced the development of reverse faults within the B-domain; and (4) favoured the advance of the S-C structures within the C-domain. This implies that an increasingly greater rock-volume was progressively involved in the deformation leading increasing thickness of the shear zone. The progressive expansion of the sheared rock-volume affected mostly the overthrusting calcareous rocks.

Hypothesis b) According to the hypothesis (b), asymmetric folds and S-C structures formed in the B- and C-domains, respectively during a first stage. The progressive deformation (second stage) produced (1) the detachment of the previously developed asymmetric folds, (2) the overthrusting and development of the mylonitic level and (C) the cataclasite of the A-domain. This implies a progressive shear zone which decreases in thickness.

At present one metre of pelitic rocks belonging to the footwall block and about 6–7 metres of calcareous rocks of the hangingwall block make up the Mt. Aquilaia shear zone.

Strain was partitioned during shearing. Partitioning was mainly controlled by (a) lithologic contrast, (b) lithologic anisotropy and (c) fluid-assisted deformation and related metamorphic reactions. The lithologic contrast between clayey (Scaglia Toscana Fm.) and calcareous rocks (Marne a *Posidonomya* Fm.), occurring in the footwall and hangingwall blocks respectively, favoured the development of the main thrust detachment taking place at the contact between these two formations. The rheological proprieties of the clayey rocks favoured the development of the high damage zone, presently coinciding with the clayey gouge of the A-domain. Furthermore, the alternation of pelitic and calcareous strata typifying the Marne a *Posidonomya* Fm. represented a very important lithological anisotropy during shearing which strongly influenced the deformation mechanism within the B- and C-domains. Here, lithological contrast and anisotropy coexisted inducing strain partitioning, which favoured strain concentration within the pelitic strata, often activated as minor detachments on the bedding surfaces.

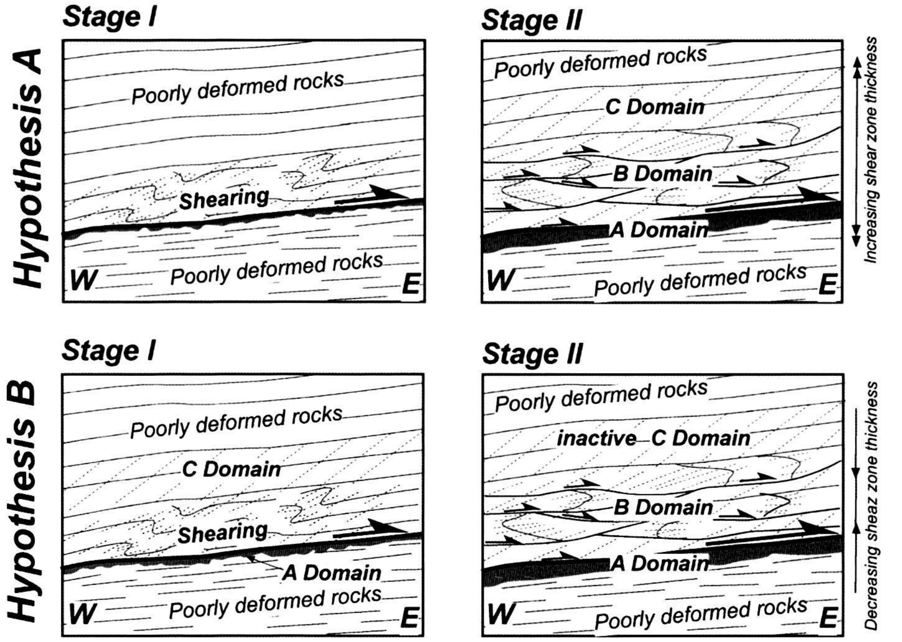

Fig. 17. Idealised and schematic evolution of the Mt. Aquilaia shear zone. See text for explanation.

The calcite veins within the sheared rocks suggest that the deformation was fluid-assisted (Bell & Cuff 1989, Musumeci 2002). Fluid-assisted deformation within shear zones is facilitated where the deformed rocks are composed of one ore more soluble minerals. In this case, solution transfer and pressure solution processes are the main mechanisms related to the deformation. Syn-deformational fluids play a fundamental role during metamorphic processes enhancing the development of replacement reactions (Musumeci 2002 and references therein) and favouring the increase of strongly deformed layers (e.g. mylonitic foliation) as the result of a preferential partitioning of the strain. This is demonstrated by the mylonite zone occurring at the base of the B-domain (Fig. 6). Metamorphism within the sheared rocks is documented by the illite crystallinity analysis which indicates that deformation took place under highest anchizonal facies conditions, at temperatures of about 200°–250°C. These temperature values are also confirmed by the calcite twin-lamellae developed in the calcite grains occurring within the veins, referable to type II and III as defined by Burkhard (1993) and indicating comparable temperatures. The anchizone implies very low-grade metamorphism in pelitic rocks (Kisch 1983, Frey 1987, Frey & Robinson 1999). The width of the 001 illite peak (illite crystallinity, IC) has been shown to be a sensitive indicator of tectono-metamorphic alteration in clays (Kubler 1968, Frey & Robinson 1999). The anchizonal facies, as defined by the Kubler Index (KI) of the illite (Kisch 1990, Kubler 1984) is documented in Table 1. They show lower KI values (some of these belonging to the anchizonal facies) with respect to poorly deformed equivalents (Table 2). These data suggest a localised alteration of clay minerals (Fig. 13), as is also supported by 0,5–2 μm crystals of illite-illite/smectite mixed-layers (Reynold & Hower, 1970) only developed within the cleavage domains (Fig. 13). These minerals, together with quartz and calcite, may be considered as a mineralogical association related to metamorphic reactions, which produced the local alteration of clay minerals only in the very deformed rocks (e.g. samples MA1–6). In fact, the same lithotypes involved in the shear zone, but poorly deformed, are characterised by the lowest KI values, referable to the *highest grade diagenetic zone*. Besides, the same lithotypes collected far from the shear zone did not experience any metamorphism.

All this implies that localised very-low grade metamorphism was probably induced by strain heating and syn-deformational fluid circulation within the sheared rocks. In this view, shear heating could be considered as the heat source of low-grade metamorphism, produced by friction, whereas the fluid circulation favoured the associated metamorphic reactions, as is largely documented and discussed in literature (Barton & England 1979, Molnar et al. 1983, Spray 1992, Leloup & Kienast 1993, Peackok et al. 1994, Lin 1999, Camacho et al. 2001). The microstructural analyses indicated that metamorphic reactions took place only in the cleavage domains, where mechanical and chemical processes acted mutually. The occurrence of illite-illite/smectite within the cleavage domains and of chlorite within the microlithons is recurrent features in the lithotypes composed of soluble minerals (Waldron & Sandifor 1988, Passchier & Trouw 1999). This is due to the ion exchange favoured by fluid-assisted deformation during the development of the cleavage domains under very low to low-grade metamorphic conditions (Passchier & Trouw 1999 and references therein). The activation of the metamorphic reactions needs adequate time. In fact, the shear zone must attain a higher temperature than the surrounding rocks for sufficiently long periods in order to produce the physical and mineralogical transformations. This indicates a relatively long period of activity for the Mt. Aquilaia shear zone.

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REFERENCES

BARBER D.J. & WENK H.R. 1979: Deformation twinning in calcite, dolomite and other rhombohedral carbonates. Phys. Chem. Minerals 5, 141–182.

BARTON C.M. & ENGLAND P.C. 1979: Shear heating at the Olympus (Greece) thrust and the deformation proprieties of carbonates at geological strain rates. Geological Society of American Bulletin 90, 483–492.

BATINI F., BROGI A., LAZZAROTTO A., LIOTTA D., PANDELI E. 2003: Geological features of Larderello-Travale and Mt. Amiata geothermal areas (southern Tuscany, Italy). Episodes 26, 239 – 244.

BELL, A.M., 1981: Vergence: an evaluation. Journal of Structural Geology 3, 197–202.

BELL T.H., CUFF C. 1989: Dissolution, solution transfer, diffusion versus fluid flow and volume loss during deformation/metamorphism. Journal Metamorphic Geology 7, 425–447.

BELL T. H., DUNCAN A. C. & SIMMONS J. V. 1989: Deformation partitioning, shear zone development and the role of undeformable objects. Tectonophysics 158, 153–161.

BERTINI G., CAMELI G.M., COSTANTINI A., DECANDIA F.A., DI FILIPPO M., DINI I., ELTER F.M., LAZZAROTTO A., LIOTTA D., PANDELI E., SANDRELLI F. & TORO B. 1991: Struttura geologica fra i monti di Campiglia e Rapolano Terme (Toscana meridionale): stato attuale delle conoscenze e problematiche. Studi Geologici Camerti, Vol. Spec. 1, 155–178.

BROGI, A. 2004a: Assetto geologico del nucleo di Falda Toscana affiorante nel settore occidentale del Monte Amiata (Appennino Settentrionale): strutture pre- e sin- collisionali relitte preservate durante lo sviluppo della tettonica distensiva post-collisionale. Boll. Soc. Geol. It. 123, 443–461.

BROGI A. 2004b: Seismic reflection and borehole logs as tools for tectonic and stratigraphic investigations: new geological data for the Tuscna Nappe exposed in the Northeastern Monte Amiata area (Northern Apennines, Italy). Boll. Soc. Geol. It. 123, 189–199.

BROGI, A. 2004c: Miocene low-angle detachments and upper crust megaboudinage in the Mt. Amiata geothermal area (Northern Apennines, Italy). Geodinamica Acta 17, 375–387.

BROGI, A. & LAZZAROTTO, A. 2002: Deformazioni sin-collisionali nella Falda Toscana a sudovest del Monte Amiata (Toscana meridionale): il sovrascorrimento di Monte Aquilaia. Boll. Soc. Geol. It. 121, 299–312.

BROWN M. & SOLAR G.S. 1998: Shear-zone systems and melts: feedback relations and self-organization in orogenic belts. Journal of Structural Geology 20, 211–227.

BRUNET C., MONIE P., JOLIVET L. & CADET J.P. 2000: Migration of compression and extension in the Tyrrhenian sea, insights from 40Ar/39Ar ages on micas along a transect from Corsica to Tuscany. Tectonophisics 321, 127–155.

BURKHARD M. 1993: Calcite twins, their geometry, appearance and significance as stress-strain markers and indicators of tectonic regime: a review. Journal of Structural Geology 15, 351–368.

CALAMAI A., CATALDI R., SQUARCI P. & TAFFI L. 1970: Geology, Geophysics and Hydrogeology of the Monte Amiata Geothermal Field. Geothermics, 1 Vol. Spec., 1–9.

CAMACHO A., MCDOUGALL I., ARMSTRONG R. & BRAUN J. 2001: Evidence for shear heating, Musgrave Block, central Australia. Journal of Structural Geology 23, 1007–1013.

CANUTI P., FOCARDI P., SESTINI G. 1965: Stratigrafia, correlazione e genesi degli Scisti Policromi dei Monti del Chianti. Boll. Soc. Geol. It. 6, 93–166.

CARMIGNANI L., DECANDIA F.A., DISPERATI L., FANTOZZI P.L., LAZZAROTTO A., LIOTTA D. & MECCHERI M. 1994: Tertiary extensional tectonics in Tuscany (Northern Apenines, Italy). Tectonophysics 238, 295–315.

CARMIGNANI L., DECANDIA F.A., DISPERATI L., FANTOZZI P.L., KLIGFIELD R., LAZZAROTTO A., LIOTTA D. & MECCHERI M. 2001: Inner Northern Apennines. In: Gian Battista Vai & I. Peter Martini (Eds.): anatomy of an Orogen: the Apennines and adjacent Mediterranean Basins, 197–213, Kluwer Academic publ.

FAZZINI P. 1978: Aspetti particolari della Scaglia Toscana nella regione amiatina: la successione di Monte Aquilaia. Atti Soc. Nat. Mat. Modena 108, 29–37.

FAZZUOLI M., PANDELI E. & SANDRELLI F. 1996: Nuovi dati litostratigrafici sulla Scaglia Toscana (scisti policromi) dei Monti del Chianti (Appennino settentrionale). Atti Soc. Tosc. Sci. Nat. Mem. Serie A, 103, 95–104.

FERRIL D.A. 1991: Calcite twin width and intensities as metamorphic indicators in natural low-temperature deformation of limestone. Journal of Structural Geology 13, 667–675.

FREY M. 1987: Very low-grade metamorphism of clastic sedimentary rocks. In: M. Frey (Ed.): low temparature metamorphism, 9–58, Blackie and Son Ltd, Glasgow.

FREY M., ROBINSON D. 1999: Low-grade metamorphism. Blackwell science ltd. Editorial office, 312 p.

JOLIVET L., FRIZON DE LAMOTTE D., MASCLE A. & SERANNE M. 1999: The Mediterranean basins: Tertiary extension within the Alpine orogen – an introduction. In: Durand B., Jolivet L., Horvath F. & Seranne M (Eds.): The Mediterranean basins: Tertiary extension within the Alpine Orogen. Geol. Soc. Lond. Spec. Publ. 156, 1–14.

KISCH H.J. 1983: Mineralogy and petrology of burial diagenesis (burial metamorphism) and incipient metamorphism in clastic rocks. In: G. Larsen & G.V. (eds.): Ghillingar diagenesis in sediments and sedimentary rocks. Vol. 2, 289–493, 513–541, Elsevier.

KISCH H.J. 1990: Calibration of the anchizone: a critical comparision of illite crystallinity scales used for definition. J. Metam. Geol. 8, 31–46.

KISCH H.J. 1991: Illite cristallinty: recommendations on samples prerating, X-ray diffraction settings and interlaboratory samples. J. Met. Geol. 9, 665–670.

KUBLER B. 1967: Anchimetamorphisme et schistosité. Bull. Centre Rech. Pan. S.N.P.A. 1, 259–278.

KUBLER B. 1968: Evaluation quantitative du metamorphisme par la cristallinité de l'illite. Bull. Centre Rech. Pan. S.N.P.A. 2, 385–397.

KUBLER B. 1984: Les indicateurs de trasformations physiques et chimiques dans la diagenèse, temperature et calorimétrie. In: Lagache (Ed.): Thermometrie et barométrie géologiques. Sez. Franc. Miner., Paris, 489–596.

LELOUP P.H. & KIENAST J.R. 1993: High-temperature metamorphism in a major strike-slip shear zone: the Ailao Shan-Red River. People's Republic of China. Earth and Planetary Sciences 39, 457–471.

LIN A. 1999: Roundness of clasts in pseudotachylytes and cataclastic rocks as an indicator of frictional melting. Journal of Structural Geology 21, 473–478.

MOLNAR P., PING C.W., PADOVANI E. 1983: Calculated temperatures in overthrusted terrains and possibile combinations of heat sources responsible for the Tertiary granites in the greater Himalaya. Journal of Geophysical Research 88, 4183–4214.

MONTÉSI L.G.J. & HIRTH G. 2003: Grain size evolution and the rheology of ductile shear zone: from laboratory experiments to postseismic creep. Earth and Planetary Science Letters 211, 97–110.

MUSUMECI G. 2002: Sillimanite-bearign shear zones in syntectonic leucogranite: fluid-assistend brittle-ductile deformation under amphibolite facies conditions. Journal of Structural Geology 24, 1491–1505.

PANDELI E., BERTINI G., CASTELLUCCI P., MORELLI M., MONECHI S. 2005: The Ligurian, Subligurian and Tuscan Units of the Monte Amiata geothermal region (South-eastern Tuscany): new stratigraphic and tectonic data. Boll. Soc. Geol. It., vol. Spec. 3, 55–72.

Passchier C.W. & Trouw R.A. 1996: Microtectonics. Springer-Verlag Berlin Heidelberg, 289 p.

Peacock S.M. 2002: Blueschist-facies metamorphism, shear heating, and P-T-t paths in subduction shear zones. Journal of Geophysical Research 97, 17693–17707.

Peacock S.M., Rushmer T. & Thompson A.B. 1994: Partial melting of subducting oceanic crust. Earth and Planetary Sciences Letters 121, 227–244.

Pertusati P.C., Plesi G. & Cerrina Feroni A. 1977: Alcuni esempi di tettonica polifasata nella Falda Toscana. Boll. Soc. Geol. It. 96, 587–603.

Ramsay J.G. 1967: Folding and fracturing of rocks: McGraw-Hill Book Company, New York, 560 p.

Ramsay, J.G., & Huber, M.I. 1987: The techniques of Modern Structural Geology. Volume 2: Folds and Fractures. Academic Press, London, 391 p.

Reynolds R.C., Hower J. 1970: The nature of interlayering in mixed-layer illite-montmorillonites. Clays and Clay Minerals 18, 25–36.

Spray J. 1992: A physical basis for the frictional melting of some rock-forming minerals. Tectonophysics 204, 205–221.

Waldron H.M. & Sandiford M. 1988: Deformation volume and cleavage development in metasedimentary rocks from the Ballarat slate belt. Journal of Structural Geology 10, 53–62.

Wang C. & Ludman A. 2004: Deformation conditions, kinematics, and displacement history of shallow crustal ductile shearing in the Norumberga fault system in the Northern Appalachians, eastern Maine. Tectonophysics 384, 129–148.

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