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August 20-28, 2004

Volume n° 6 - from P55 to PW06

32nd INTERNATIONAL GEOLOGICAL CONGRESS

LOW-ANGLE NORMAL FAULTING... TWENTY YEARS AFTER

Leaders:
F. Brozzetti, L. Jolivet, R.E. Holdsworth

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Post-Congress **PW01**

The scientific content of this guide is under the total responsibility of the Authors

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Front Cover:
Outcrop of the Zuccale detachment fault (ZDF), near Punta di Zuccale. The sub-horizontal ZDF plane puts into contact Late Cretaceous Helminthoid Flysch, in the hanging-wall, with Triassic sediments of the Verrucano Fm on the foot-wall.

Leaders: F. Brozzetti, L. Jolivet, R.E. Holdsworth

Introduction

F. Brozzetti, L. Jolivet and R. Holdsworth

Preface

Field Workshop PW01 focuses on the geological and geophysical aspects of low-angle normal faults (hereafter referred to as LANFs), and will examine their geometry, kinematics, mechanics and seismogenic character, with special emphasis on field-based observations. The present workshop occurs some twenty years after LANFs became widely recognized by geologists as regionally significant structures in extensional geodynamic environments following the 1984 GSA meeting, held in Nevada.

At about that time, the tectonic setting and fault architecture of the Basin and Range Province was undergoing a revolutionary re-interpretation with widespread documentation of LANFs and associated deformation structures (e.g. Davis et al., 1980; Wernicke, 1981; Wernicke and Burchfiel, 1982; Davis 1980; Spencer, 1984, Wernicke, 1985).

In this revolution of ideas, new concepts were defined and characteristic structural assemblages recognized. New terms such as "metamorphic core complex", "breakaway zone", "extensional allochthon" were introduced into the literature (Wernicke, 1981; Wernicke & Burchfiel, 1982, Davis and Lister, 1988) and have since become widely accepted (e.g. Lister and Davis 1989). Two important discoveries emerged from this seminal period of research. The first was the recognition that LANFs have the potential to accommodate very large amounts of crustal extension (e.g. 100-200% in some cases; Davis et al. 1980). This allows such faults to act as major agents of tectonic exhumation, helping to bring mid-to lower crustal, or potentially even upper mantle rocks to the surface over relative short periods of geological time (Hoogerduijn Strating et al., 1993). The second was that major LANFs could act as a crustal- to lithosphere-scale basal detachment to an array of synthetic and antithetic, high-angle rotational and non-rotational normal faults in the hangingwall or "upper plate" (Wernicke, 1981; Wernicke & Burchfiel, 1982, Davis and Lister, 1988). In such models, the basal detachment was a brittle fault in the upper crust, broadening with increasing depth into a low-angle ductile shear zone in the deeper crust and upper mantle. It was proposed that increasing amounts of extension would progressively exhume the evolving fault rock assemblages of the regional LANF, leading to a characteristic sequence of ductile mylonites in the footwall – or "lower plate" regions - passing upwards into later, lower temperature cataclasites and brittle faults adjacent to the boundary with the hangingwall or upper plate.

Since they were first described by geologists working in the Basin and Range, the origin, evolution and existence of LANFs as seismically active structures has been hotly debated, especially within the geophysics community. At face value, LANFs present an apparent paradox from a mechanical viewpoint. Classical Anderson-Byerlee frictional fault mechanics suggests that as a first approximation, the Earth's surface can be assumed to correspond to a plane of principal stress (i.e. vertical $\Uparrow_1$ in extensional regimes), with laboratory-measured sliding friction coefficients typically falling in the range $0.6<\mu s<0.85$ (Anderson, 1951; Byerlee, 1978; Sibson, 1985). This predicts that normal faults should initiate at dips ~60° and then rotate (domino-like) down to frictional lockup angles (40°-30°). Thus slip upon severely misoriented LANFs with dips less than 30° seems mechanically unlikely. This prediction seems entirely consistent with the globally observed dip-ranges of moderate and large normal fault seismogenic ruptures where the rupture plane is unambiguously identified in the focal mechanisms (e.g. Jackson and White, 1989; Collettini and Sibson, 2001).

Despite the very significant theoretical objections to the existence of LANFs as regionally significant structures, they have now been recognised in numerous extended terrains located all over the world (e.g. Burchfiel et al., 1987, Hamilton, 1988; Howard et al., 1987; Carmignani and Kligfield, 1990, Abers, 1991; Johnson and Loy, 1992; Baldwin et al., 1993; Spencer, 1993; Wernicke, 1995, Jolivet et al., 1990, 1991, 1998, 1999, Barchi et al., 1998; Boncio et al., 1998, 2000, Collettini and Barchi, 2002; Hayman et al. 2003; Cowan et al. 2003). More recently, LANFs have begun to be recognized in oceanic lithosphere forming adjacent to certain slow spreading ridge-transform fault junctions (e.g. Cann et al. 1997 plus other refs.).

As a result, three groups of models have been proposed to explain the paradoxical existence of LANFs. Models in the first group suggest that many LANFs initiated as typical 'Andersonian' structures with dips of ~60° and that they have subsequently been reoriented to lower dips during continued extension, becoming essentially inactive beyond frictional lock-up. Re-orientation has been envisaged in some cases as a passive process resulting from the domino-rotation of successive normal fault sets, following the model of Proffett (1977). Others invoke more dynamic models in which isostatic adjustments generate footwall flexure and uplift (Wernicke and Axen, 1988, Buck et al 1988), rotating initially steeply-dipping normal faults into low-angle orientations. A second group of models proposes that LANF movement can occur where dramatic departures from an Andersonian state of stress are induced by severe topography (Abers et al., 1997) or high shear stresses at the base of the brittle crust (Westaway, 1999). A third group of models suggests that low-angle normal faulting is possible in an Andersonian stress field if the LANF is weak, either due the presence of high fluid pressures (refs.) and/or weak fault rocks in which the sliding friction coefficients are significantly lower than the standard laboratory values, i.e. μs less than 0.6, perhaps as low as 0.2 (e.g. Holdsworth 2004). Plausible fault rock weakening mechanisms require fluid-rock interactions in the core regions of LANFs and include reaction softening producing phyllosilicate-rich fault rocks (e.g. Gueydan et al. 2003) and/or the onset of fluid-assisted, grain-size-sensitive dissolution-precipation deformation mechanisms (Holdsworth 2004; Collettini & Holdsworth 2004). Such weak faults would be unlikely to produce large magnitude earthquakes since the fault rocks in its core would probably be too weak to sustain sufficient stresses. Instead, they might be expected to move by a combination of aseismic creep and abundant microseismicity (Colletini & Barchi 2002, 2004; Collettini & Holdsworth 2004).

In recent years, it has been suggested that LANFs are common in the Northern Apennines and Tyrrhenian basin region, playing a major role in determining the geometry, kinematic and dynamic evolution of rifting (Lavecchia et al., 1984; Lavecchia, 1988; Jolivet et al., 1990, 1994, 1998, Keller and Pialli, 1990, Carmignani et al., 1994, Gautier and Brun, 1994; Pialli., 1998 and therein ref.). It is generally agreed that, during the last 25 Ma, the locus of extensional deformation has migrated progressively eastwards from Alpine Corsica to western Umbria, passing through the north Tyrrhenian basin and Tuscany. The main surface expression of extension are narrow and elongate Neogene-Quaternary grabens that young progressively eastwards bounded by high-angle conjugate normal faults that are thought to detach along E-dipping LANFs at depth (e.g. Elter et al., 1975; Boccaletti et al., 1987; Bartole, 1995; Jolivet et al., 1998). A number of older LANFs that accommodate substantial extension in the Apennine system are now exposed at the surface in the Tyrrhenian islands and Tuscany (e.g. Jolivet et al., 1998).

The Corsica - Elba -Perugia Transect

During a 2-day trip in the northern part of Alpine Corsica (figure 1) we shall visit key-outcrops of a complex structure which involves a first episode of crustal thickening (Late Cretaceous to Early Oligocene), and a second episode of crustal scale extension and basin formation (Late Oligocene to recent times). This second episode was associated with whole-crust extension and the formation of shallow-dipping extensional shear zones (figure 2), the most typical being the East Tenda Shear Zone (figure 3).

For many years Alpine Corsica was considered a piece of the Franco-Italian Alps rifted away from the European margin during the Oligocene without much internal deformation after the end of compression. The mountainous landscapes of Corsica are indeed often very similar to those of the Alps but one important physiographic aspect of the area had been forgotten: Corsica is an island. It turned out that most structures seen today in the field resulted from the extensional part of the tectonic history of Corsica. Nowadays extension is more precisely understood and the kinematics and dynamics of the early compressional stage is more enigmatic. There is still work to be done to unravel the crustal stacking episode and the associated HP-LT metamorphism.

The outcrops examined during this field-trip are distributed along an E-W transect from the autochthonous Eocene foreland basin, to the west to the exhumed high-pressure and low-temperature metamorphic rocks to the east.

We will observe the footwall and hanging wall of the East Tenda Shear Zone, the oldest of the shallow-dipping extensional shear zones along the Corsica-Apennines transect.

The island of Corsica is located between two Cenozoic extensional basins; the northern part of the Liguro-Provençal Basin formed during the Early Miocene (Westphal et al., 1976; Burrus, 1984; Gueguen, 1995; Chamot-Rooke et al., 1999; Gattacceca, 2000; Speranza et al., 2002) and the northern part of the Tyrrhenian Sea, formed from the Early to Middle Miocene on the backarc side of the Apennines (Malinverno and Ryan, 1986; Réhault

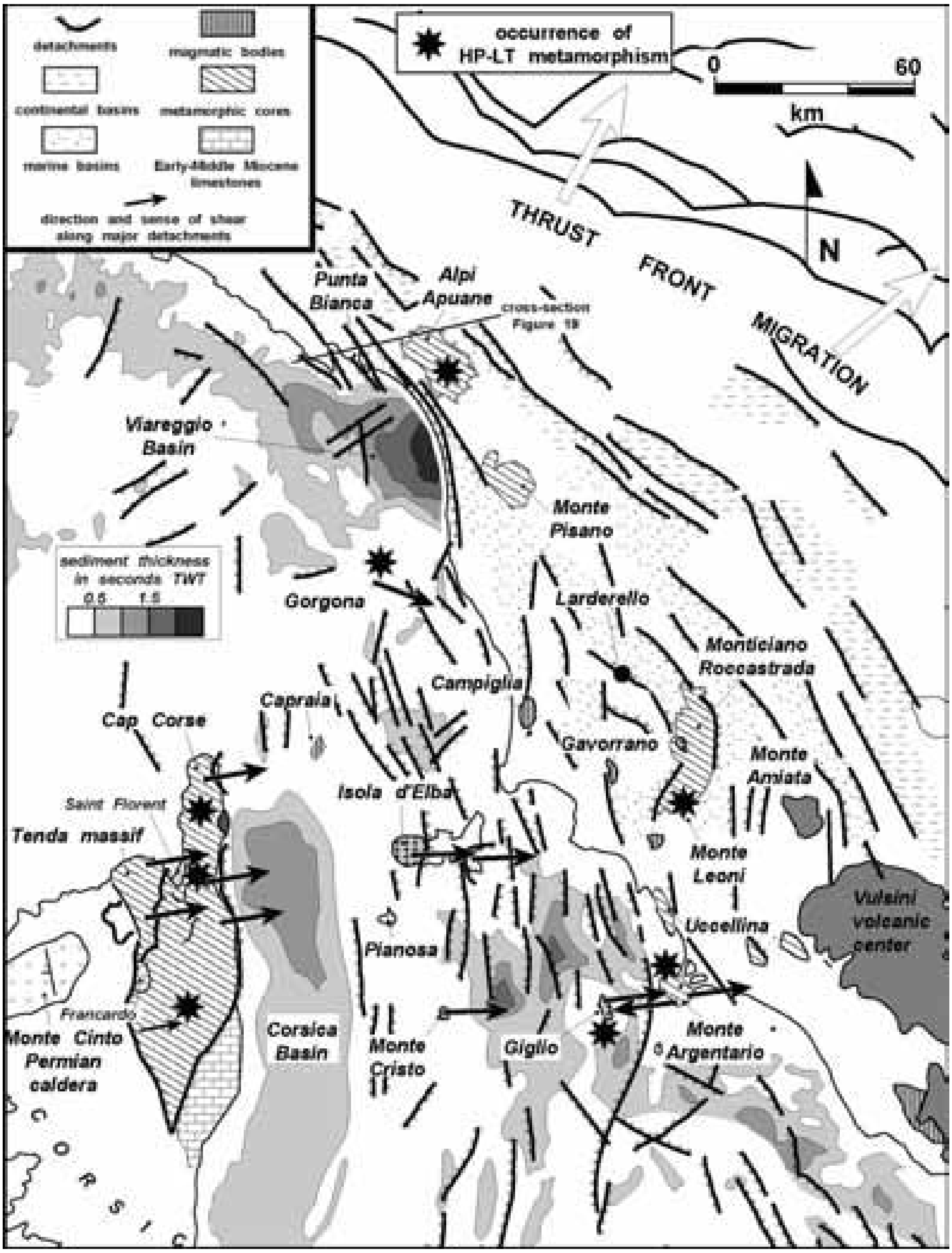

Figure 1 - Tectonic map of the northern Tyrrhenian Sea showing the position of Alpine Corsica in the western part of the studied transect (after Jolivet et al. 1998).

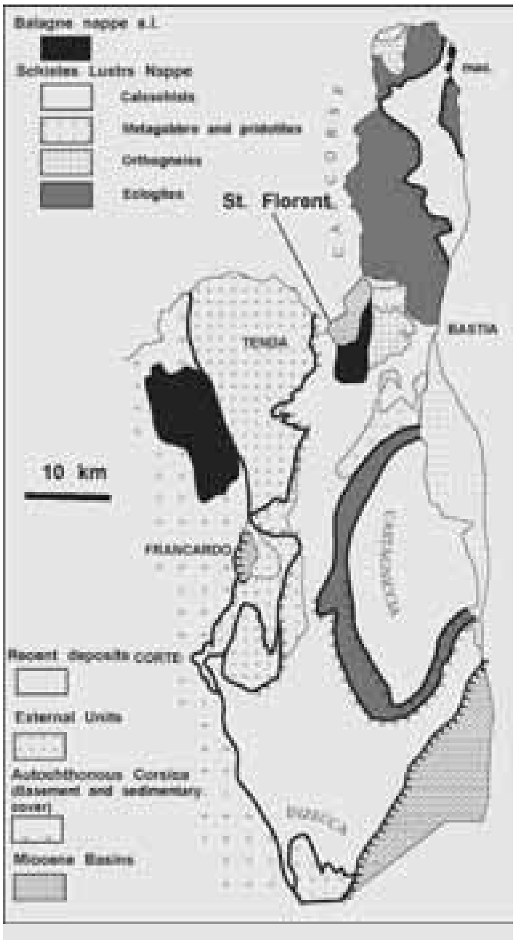

Figure 2 - Geological map of Alpine Corsica after Jolivet et al. (1991).

et al., 1987; Sartori et al., 1987; Lavecchia, 1988; Lavecchia and Stoppa, 1989; Mascle and Réhault, 1990; Sartori, 1990; Jolivet et al., 1998; Jolivet et al., 1999; Mauffret and Contrucci, 1999).

Compressional tectonics in Corsica formed a stack of nappes similar to the internal zones of the Alps (Mattauer et al., 1981; Caron, 1994) until the Late Eocene – Early Oligocene when extension started between Corsica and Provence (Jolivet et al., 1998).

The Eocene thrust pile was then reworked by post-orogenic shallow - east-dipping extensional shear zones such as the East Tenda Shear Zone (ETSZ) that unroofs the Tenda massif (Daniel et al., 1996).

The ETSZ footwall is the Tenda massif made of orthogneiss and minor metasediments and gabbros metamorphosed in high-pressure and low-temperature conditions (lower blueschists). The hangingwall is made of the Schistes Lustrés nappe that mostly comprised oceanic metamorphic rocks as well as minor continental orthogneiss and meta-sediments with high-pressure and low-temperature conditions locally reaching eclogite facies P-T conditions.

All contacts are reworked to various degrees by the late extensional event and the extensional deformation is penetrative within the nappe stack.

It is associated with severe retrogression of HP-LT parageneses in the greenschist facies, mostly along the contacts (Fournier et al., 1991).

The East Tenda Shear Zone and the contemporaneous easterner minor shear zones observed in Corsica developed mostly in the pressure and temperature conditions of the greenschist facies thus within the brittle-ductile transition.

During the Corsica part of the fieldtrip we shall observe the progressive deformation of the Tenda massif orthogneiss and the progressive formation of a new greenschist foliation and discuss the causes of strain localisation at the depth of the brittle-ductile transition (Gueydan, 2001; Gueydan et al., 2001, 2003a).

After reaching Tuscany we shall investigate the extensional structures originated by the Tyrrhenian rifting in the Elba Island and in southern Tuscany, with particular emphasis on LANFs-related ones. It is accepted that in this area extension is much greater than in the northern part of the region (that is in the Lunigiana – Garfagnana area).

The onset of extensional deformation in the inner part of the Tyrrhenian sea is generally referred to Middle - Late Miocene (Carmignani et al., 1994, Bertini et al., 1991; Decandia et al., 1998, Liotta, 2002). During this time-interval, the Orogen architecture consisted, from the top, of the Ligurides sheet (piled up over the insubric margin during a mid-Eocene phase), the sub-ligurian Unit, the Tuscan Nappe (non metamorphic) and the Tuscan metamorphic Unit (Boccaletti et al., 1971; Alvarez et al., 1976; Kligfield, 1979; Elter et al., 1993; Carmignani et al., 2001). Only the western part of the Umbria domain was undergoing compression whereas the main phase of chain building is generally referred to Messinian-Early Pliocene age (Lavecchia et al., 1987; Calamita et al., 1991; Brozzetti and Luchetti 2002).

Elba island and perityrrhenian Tuscany were involved in extensional deformation during the Late Miocene (Keller and Pialli, 1990; Daniel and Jolivet, 1995).

The third day on the Island of Elba will be dedicated

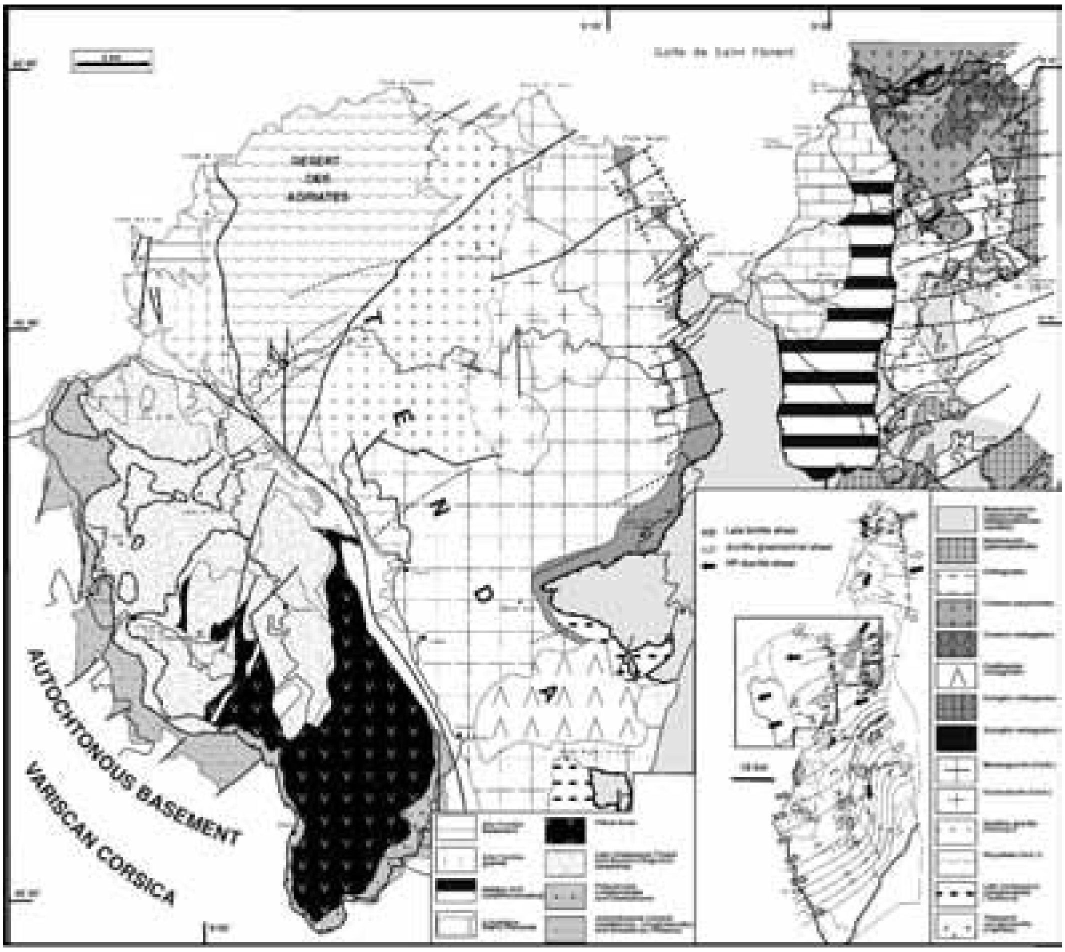

Figure 3 - Geological map around Saint Florent (Balagne after Jourdan, 1988)

to analyse the geometry and the kinematics of the Zuccale fault plane and shear zone which crop out on the eastern side of the Stella Gulf, south of Capoliveri (figure 4).

Without going into detail about Elba geology, we remark that the structure of the island consists of five imbricate complexes stacked during the Early Miocene compressional phase. The lowermost three complexes derive from the tectonization of the Tuscan domain and are characterised by different stratigraphy and metamorphic facies. The uppermost two complexes are ophiolite-bearing successions of Internal Ligurian pertinence.

The Zuccale fault dissects completely the thrust pile of the complexes 2 to 4 along the M. Orello eastern side and causes its complete repetition along the Cavo-Rio nell'Elba - M. Arco area. According to the CROP03 interpretation proposed in Pialli et al. (1998), this fault would be the innermost of an E dipping low-angle set which dissects the entire Tuscany and western Umbria pre-Apennine zone.

In Stop 22, Near P. di Zuccale the fault puts in contact the Cretaceous flisch of complex 5 in the hangingwall, with the Triassic Verrucano of complex 1, in the footwall. In this same area the kinematics of the fault and the fracturing history of the fault zone can be observed in detail.

On Stop 23 (Spiagge Nere) the time-space relationships between the low-angle fault and the aplitic dyke swarm of the Porto Azzurro Monzogranite are particularly well-exposed.

According to an extensive literature, in Tuscany different types of extensional structures can be observed (Fig. 4).

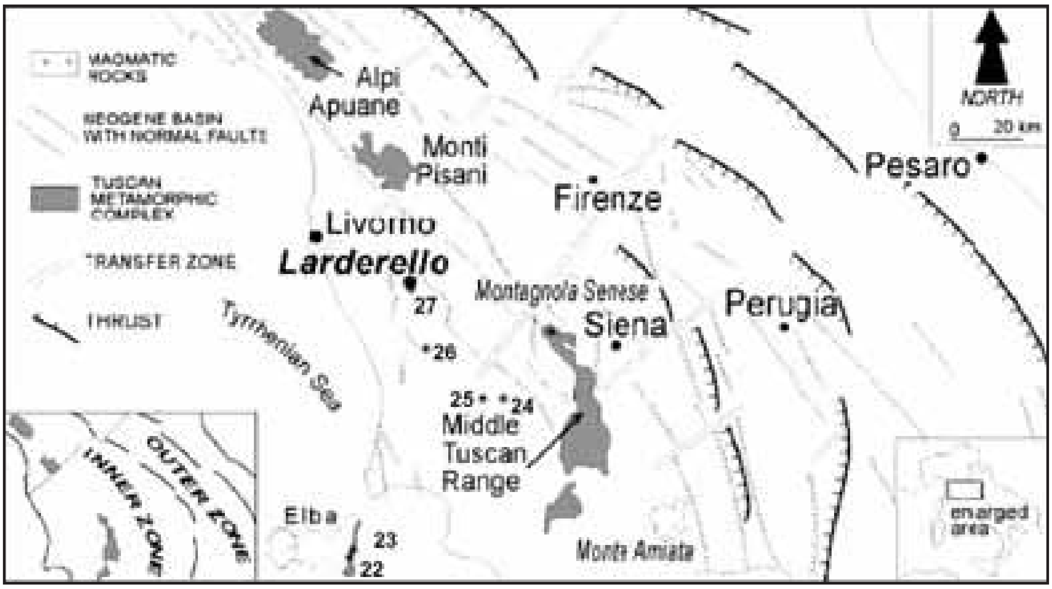

Figure 4 - Geological skecth of the Tuscany area in the sector crossed by the itinerary (numbered circles: location of stop sites)

A first one causes the Ligurides to overlay with low-angle contact, the Upper Triassic evaporites, or the Middle Triassic Verrucano, with the complete omission of the Tuscan succession of Jurassic to Oligocene age. Such a large stratigraphic omission is a common tectonic feature of southern Tuscany (expecially in the Montagnola Senese, Monticiano Roccastrada ridge) and has been known for a long time as "*Serie Ridotta toscana"*(Lazzarotto, 1967; Calamai et al., 1970). This latter has been recently explained hypothesising large displacement on extensional faults characterized by a staircase trajectory with two main flats inside both the Ligurides and the Upper Triassic Burano Anhidrites (Bertini et al., 1991; Carmignani et al., 2001; Liotta, 2002). The two flats are probably connected by ramps dissecting the carbonate and terrigeneous sequences of the Tuscan Nappe.

Upper Triassic evaporites would constitute the main detachment level which divided an upper plate characterized by brittle deformation from a lower plate characterized by coeval mainly ductile deformation. According to Carmignani & Kligfield (1990); Carmignani et al., 1994; 2001) In the Alpi Apuane core complex and in the Montagnola Senese area, the extensional ductile deformation consists of asymmetric folds and associated east-verging shear zones which deform the older Oligocene collisional structures.

K-Ar and ^{40}Ar/^{39}Ar age determinations, carried out on syn-extensional white micas provide a 12-14 Ma age (i.e. middle Miocene, Giglia and Radicati di Brozolo, 1970; Kligfield et al., 1986) for the exhumation of the Alpi Apuane core complex .

In southern Tuscany also unconformable deposition of marine sediments on the Ligurian Units stretched by the "Serie Ridotta" extensional process has been observed and dated to middle Miocene. These data, if confirmed, would suggest that the first extensional event occurred during the Langhian (?)- Serravallian interval, then after the onset of extension in the Island of Elba.

A second set of normal faults is recognized in southern Tuscany. These faults displace the older "serie ridotta" structures, flattening in the Palaeozoic phyllites and causing the stratigraphic omission of the overlying Verrucano. A typical structure due to these faults is the asymmetrical mega×boudinage of the rocks belonging to the Verrucano Group. Activity of these structures occurred in the Serravallian - Late Messinian time span as suggested by Lacustrine and marine sedimentation deposited in the associated tectonic depressions (Radicondoli and Volterra basins, Larderello area, (Baldi et al., 1994).

A further and more evident extensional feature in Tuscany is the formation of NNW-SSE and NS trending high-angle normal faults. These faults

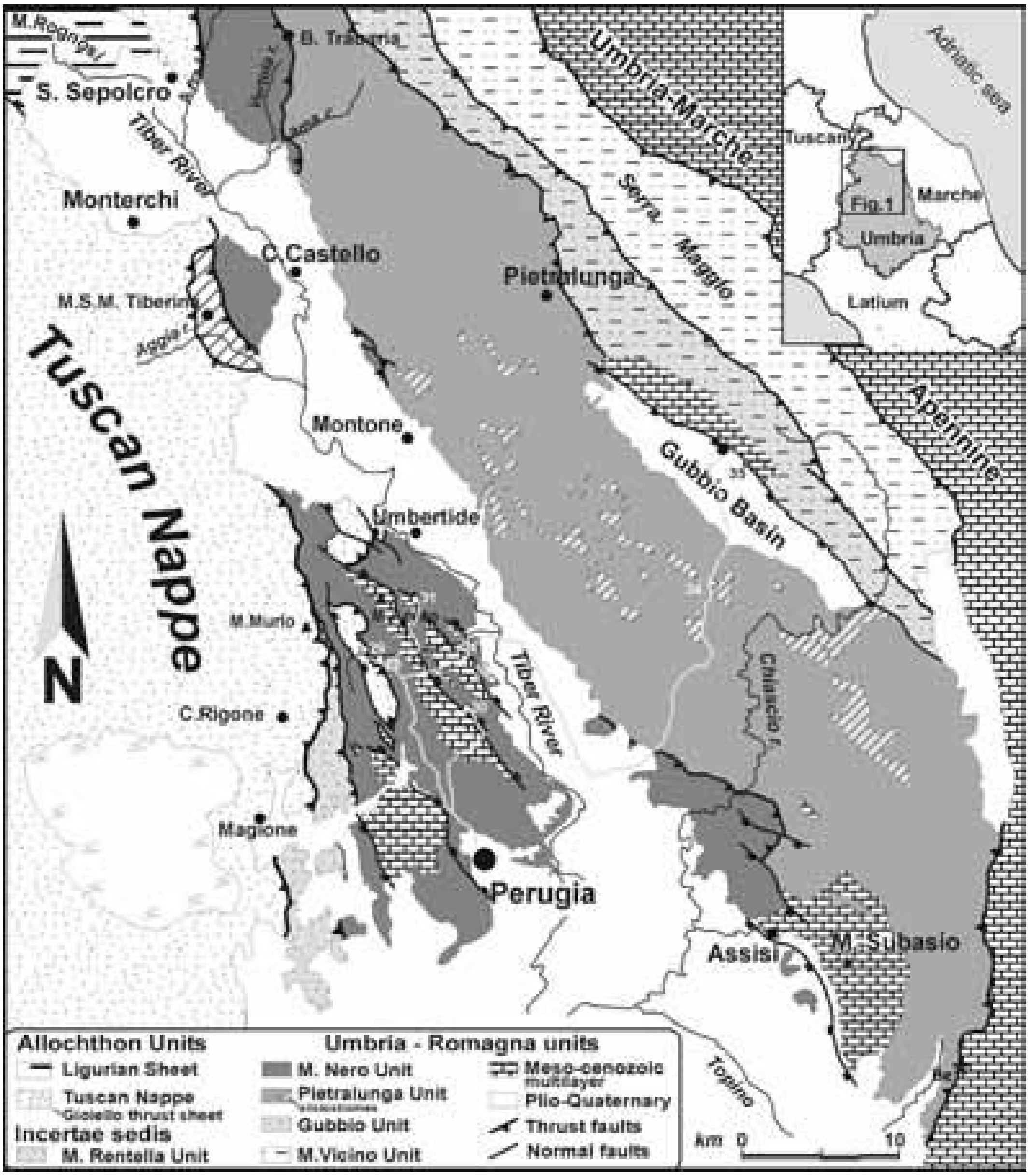

Figure 5 - Detailed geological sketch of the Study area showing the main compressional and extensional structures of northern Umbria (After Brozzetti and Luchetti, 2002)

appear to dissect all the previous structures, including the Upper Triassic evaporite and phyllite and generated several eastward younging grabens. Lower and Middle Pliocene marine sediments fill the tectonic depressions generated by these faults in western central Tuscany (Era - Elsa and Radicofani Grabens); Upper Pliocene to Lower Pleistocene lacustrine and fluvial deposits prevails in the basins located more to the east (Valdarno – Val di Chiana and Mugello – Casentino - Valtiberina grabens) (Bartolini et al., 1982, Ambrosetti et al., 1987; Bartole, 1995, Benvenuti, 1997).

One of the most evident and widespread extension-related geological processes in Tuscany, is the Late

Miocene to Middle Pleistocene Magmatism. Crustal and basic sub-crustal magmas are mixed because both the mantle and the crust, affected by extension, suffered partial melting.

Following the west to east migration of the extensional strain field, also magmatic processes rejuvenate moving from inner Tuscany to Umbria. Ages varying from 7.3 to 2.2 Ma characterises the Tuscan Archipelago magmatic bodies; 1.3 to 0.1 Ma have been ascertained for the easternmost Tuscany-latium volcanoes (Civetta et al., 1978; Lavecchia & Stoppa, 1989; Serri et al., 1993;) and for the small effusive centers of the Umbria-Latium ultra-alkaline district (Stoppa and Wolley, 1997).

The fourth day of the fieldtrip will be directed to observing some outcropping extensional faults in southern Tuscany and, in particular, attention will be drawn to low-angle structures of the "serie ridotta toscana". Stop 24, located along the Mersino Stream, shows a particularly clear exposure of the Pliocene Boccheggiano normal fault. This latter is a regional scale fault which appears as a low-angle contact between cretaceous Liguride pelagites and Palaeozoic phyllite.

Stop 25 focuses attention on the Upper Triassic Calcare Cavernoso fm, a vacuolar carbonatic breccia deriving from the alteration and tectonisation of the Burano Anhydrites, which acted as a major detachment horizon during both the compressional and the extensional tectonic phases.

Stop 26 and 27 deal with the Tuscan geothermal system whose characteristics are strongly influenced by the extensional structure of the area, involving low-angle and high-angle normal faults.

Near Monterotondo, the "Fumarole" site offers a panoramic view of a natural gas vent occurring in correspondence to the outcropping carbonatic reservoir. Stop 27 will be a guided trip around the Larderello field, the most ancient exploited geothermal field in the world. It will include a visit to the local Museum (in which the history of the Larderello exploitation and a complete collection of the tools used through the last two centuries can be examined), to a drill-hole and possibly to the geothermal supplied power station.

During the last day in the field, attention will be focused on the easterner and least known LANF along the transect. This fault, known as the Altotiberina fault (AF) is a Upper Pliocene-Quaternary east-dipping normal fault, bounding westward the Tiber Valley graben (Barchi et al.,1998; Boncio et al., 2000, figure 5). At the surface, the AF dissects and displaces eastward the leading edge of the Tuscan allochthon (Oligocene-Miocene shales and turbidites of the Falterona Nappe) and the frontal part of the Perugia Massifs thrust which is the innermost outcropping contractional structure of the Umbria Marche domain.

The importance of the AF fault, highlighted initially in Brozzetti (1995), mainly on the basis of field data, was stressed during the CROP 03 project (a deep-crust seismic reflection profile crossing the northern Apennines), led by G. Pialli (Pialli et al., 1998). An overview, synthesis, and updating of geologic, geophysical, and seismologic data available in literature on the AF have recently been published in Boncio et al. (2000) and in Collettini & Barchi (2002) in which the kinematic and seismogenic role of the AF has also been discussed.

Moving in a W to E direction, Stops 28 to 32 analyse the major tectonic features of the AF at the surface: the breakaway zone in the M. Malbe – M. Tezio area and the extensional allochthon (Perugia Massifs allochthon) along the Nese Valley. A number of both low-angle and high-angle faults of the Perugia Massifs domino-like structure will be observed in detail.

After a panoramic view of the eastern Tiber basin (Stop 33), an asymmetrical graben placed in the hanging wall of an active splay of the AF, we will move towards E to conclude the trip in the Gubbio Pleistocene half-graben (stops 34 and 35). The last site is located on a spectacular outcrop of the Gubbio fault plane, an antithetical Quaternary and presumably active structure, bounding eastward the zone affected by extensional tectonics in northern Umbria.

The workshop's final stage will be held in Perugia where the magnificent 13th century sala dei Notari, will give hospitality to the participants for a round table with discussion and planning of any future research projects and meetings on LANFs.

Regional geologic setting

Francesco Brozzetti and Laurent Jolivet

The Field Workshop PWO01 will move along the geotraverse Alpine Corsica – perityrrhenian Tuscany – Tuscan Umbria Appenines.

In all these regions surface and subsurface geology have been studied in depth for tens of years, thus a lot of stratigraphic, structural and geophysical data are available. In the last twenty years, seismic reflection lines and deep wells have provided additional information about the structure of the crust at depths less than 10 km whereas CROP profiles enlighten the deeper crustal features.

ALPINE CORSICA is a nappe stack formed in the Late Cretaceous and Eocene by underthrusting of the continental margin of Europe made of Hercynian plutonic rocks, under Ligurian oceanic units (Caron et al., 198; Mattauer et al., 1981; Amaudric du Chaffaut, 1982; Durand Delga, 1984; Caron, 1994).
The northern Apennines have developed since the Late Oligocene as a consequence of the closure of the Ligurian Ocean and the collision of the Adria.
The nappe stack can be described as follows, from base to top (figures 3 and 6) :
-**External units**. The Tenda Massif mainly consists of granitic rocks belonging to the European margin. It was deformed and metamorphosed during Late Cretaceous/Eocene thrusting of the Schistes Lustrés Nappe toward the west (Mattauer et al., 1981; Gibbons and Horak, 1984; Warburton, 1986; Jourdan, 1988). It is thought to be either the autochthonous basement (Mattauer et al., 1981; Gibbons and Horak, 1984; Waters, 1990) or a thrust sheet slightly displaced toward the west during shortening of the European margin (Warburton, 1986; Jourdan, 1988).
To the south, the Corte Units are imbricated in a nappe stack which involves the basement and its sedimentary cover (Amaudric du Chaffaut, 1982; Bézert and Caby, 1989; Egal, 1992), along the contact between Alpine Corsica and autochthonous Western Corsica. The latter is mainly made of Hercynian rocks, locally overlain by Eocene sediments.

-**The Schistes Lustrés Nappe** *sensu lato* rests upon the Tenda Massif by a shear zone initially formed in HP-LT conditions, with top-to-the-west sense of shear (Mattauer et al., 1981), later reworked as a ductile normal fault during retrogression in the greenschist facies (Waters, 1989; Jolivet et al., 1990; Fournier et al., 1991; Jolivet et al., 1991). This nappe comprises several oceanic and continental units. Oceanic units are similar to the Schistes Lustrés Nappe of the Western Alps (Caron, 1977) in terms of lithofacies and metamorphic grade. In the south of Alpine Corsica, the Schistes Lustrés Nappe is in contact with either the Corte Units, or the Hercynian basement and its Eocene cover.
From base to top it contains the following lithologies:
(1) A thick unit of calcschists with small lenses of metabasites which crops out in the Castagniccia antiform and in the core of Cap Corse. We shall name them the Castagniccia Schistes Lustrés. They display blueschist and eclogite parageneses.
(2) A complex pile of metabasalts and metagabbros which we shall see in the Lancone Gorges, near the northern tip of Cap Corse and its western coast. We will use the name of Lancone metabasites. The metamorphic grade reaches the eclogite facies.

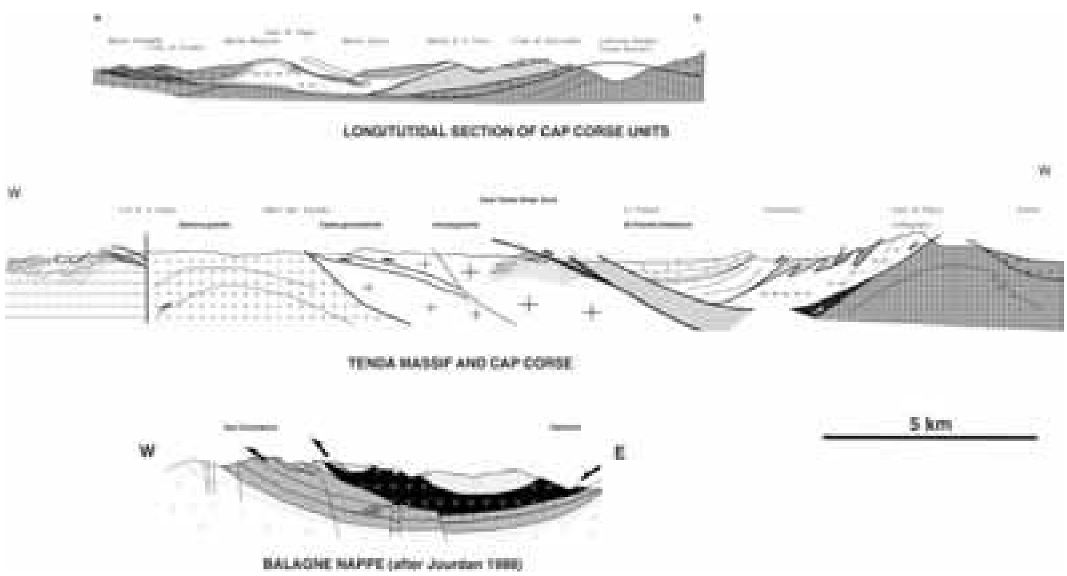

Figure 6 - Cross-sections of Alpine Corsica (Balagne after Jourdan 1988).

(3) Serpentinite lenses and metagabbros
(4) Calcschists and metabasites (Inzecca Schistes Lustrés) which crop out mostly in the south of Alpine Corsica until the latitude of Bastia. Their thickness is extremely variable. They display typical blueschist metamorphic associations with lower temperatures than in the underlying units and a colder retrograde path which has better preserved HP minerals.
(5) Lenses of continental basement and its sedimentary cover (marbles, quartzites and minor schists). They crop out as separates bodies with large thickness variations. The main outcrops are found between Sera di Pigno and Oletta, near Farinole, at Mt Zucarello, and near the village of Centuri at the northern end of Cap Corse. Metamorphic grade is variable from blueschist to eclogite (Lahondère, 1988; Fournier et al., 1991; Lahondère, 1996). Units displaying eclogite parageneses will be termed Farinole gneiss, and the blueschist one Sera di Pigno-Oletta gneiss.
(6) Large bodies of serpentinites, peridotites and gabbros which show only intermediate pressure associations locally reaching the blueschist facies. They crop out at Cima di Gratera, and at Monte Maggiore near the northern end of Cap Corse. We shall use the name of Gratera serpentinites and gabbros.

-**The upper unit** (called hereafter the **Balagne Nappe**) is characterized by a lack of metamorphism and ductile deformation. It is chiefly constituted of ophiolitic material and oceanic sediments of Upper Jurassic age as well as Upper Cretaceous flysch (Durand Delga, 1984). It rests upon all metamorphic units by a sharp contact subsequently folded into broad antiforms and synforms. It is distributed in three areas: (1) the Balagne Nappe *sensu stricto* rests upon non metamorphosed Eocene autochthonous sediments west of the Tenda Massif, (2) the Nebbio Klippe and (3) the Macinaggio Klippes in the Centuri region rest upon the Schistes Lustrés Nappe. This superficial nappe was emplaced in the Middle Eocene above Western Corsica, as attested by the presence of olistostrome deposits in the autochthonous Eocene basin below the Balagne Nappe s. s. (Durand Delga, 1984; Jourdan, 1988). Beneath the Nebbio Klippe, poorly metamorphosed acidic rocks crop out within the contact zone. They are similar to various facies of the Tenda Massif, though less deformed.
All these units are unconformably overlain by terrestrial to shallow marine deposits of Lower Miocene age (Francardo and St. Florent basins). The base of the basin is made of continental fluvial deposits attributed to the Upper Oligocene. A marine transgression is recorded from the Late Burdigalian as shallow water limestones and conglomerates. The Langhian episode is still calcareous and the Serravalian is more sandy and marly. The uppermost formation is a Tortonian continental conglomerate which erodes the underlying deposits. All formations are tilted toward the west. The lowermost formation is more tilted and separated from the upper ones by a slight angular unconformity (see Ferrandini et al., 1996) for more details).

A short note on radiometric ages of metamorphism:
$^{40}Ar/^{39}Ar$ dating on micas of the Schistes Lustrés nappe and Tenda massif was recently conducted along an E-W transect (Brunet et al., 2000). Ages range from the Late Cretaceous to the Early Miocene. Late Cretaceous age obtained for the eclogite in Monte Pinatelle are consistent with the ages obtained in the Castagniccia eclogite with the Sm/Nd method (Lahondère and Guerrot, 1997). Little is known about the kinematics of this stage but the existence of a Late Cretaceous HP-LT stage seems now granted.
In all units, from the Tenda massif to the Schistes Lustrés, the blueschists stage is associated with ages around 40-45 Ma. This age is consistent with previously published data. In this age MP-LT conditions probably continued until the Late Eocene (>37 Ma) as shown by the presence of crossite in the Bartonian flysch near Corte (Bézert and Caby, 1988). The most striking result is the progressive resetting of ages during the top-to-the-east shear. A clear younging is paralleled with the intensity of D_2. In the most highly sheared samples ages are totally reset at 25 Ma (Tenda massif) or 27-29 Ma, while others show only partial resetting with some discordant ages. This shows that the East Tenda Shear Zone had its main activity near the Oligocene-Miocene boundary around 25 Ma. Other contacts, such as those in Centuri or near Col de Teghime have slightly older ages around 27-29 Ma. The youngest blueschist age is at 32 Ma, and so is the oldest greenschist age. This could suggest that the top-to-the-east shear started at 32 Ma which is also the age of rifting in the Liguro-Provençal basin.
We thus can propose the following tectonic timing:
90?-60? Ma: HP-LT eclogite metamorphism in the Farinole-Morteda unit and the Castagniccia schist. The kinematics of this stage remains to be clarified.

50-40 Ma: exhumation of the Cap Corse eclogites up to the blueschist conditions and burial of other continental (Sera di Pigno-Oletta) or oceanic units in the blueschist facies. This stage corresponds to the formation of a thick accretionary complex at the expense of the oceanic domain and the thinned margin of Corsica. Top-to-the-west shear senses are predominant.
45-32 Ma: underthrusting of the Tenda massif and Corte slices below the Cap Corse Schistes Lustrés. This stage corresponds to the westward migration of the thrust front onto Western Corsica.
32-25 Ma: inversion of shear sense and reworking of thrust contacts as extensional shear zones. This stage corresponds to the rifting of the Liguro-Provençal basin and the Tyrrhenian Sea.
20 Ma to Recent: end of the extensional process and formation of sedimentary basins in asymmetric grabens.

The Northern Apennines

Two well-distinct structural provinces with different geological and geophysical characters can be distinguished in the northern Apennines s.l. (figures 7 and 8).
The Tyrrhenian province characterised by positive

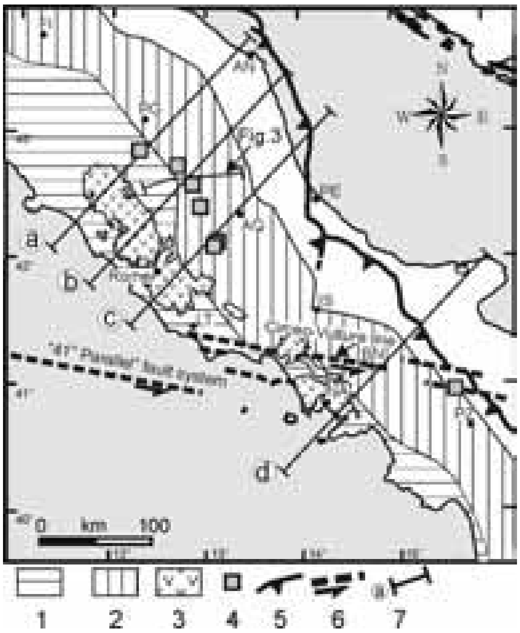

Figure 7 - Schematic tectonic zoning of the Northern-Central Apennines. 1) thinned crust zone (20-25 km); 2) thickened to normal crust zone (30-40 km)

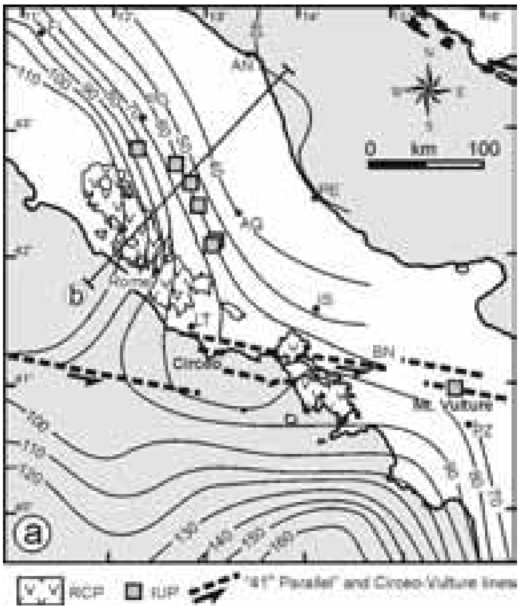

Figure 8 - Regional heat flow density map (After Mongelli and Zito, 1994)

bouguer anomalies and high heat flow values (exceeding 1000 mW /m^2 of geothermal fields) shows a depth of the Moho not exceeding 25 km with attenuated upper mantle velocities due to strong geothermic gradient. In this sector, the heat regime also explains the nature of the seismicity which shows rare, shallow and low-magnitude extensional events.
The Adriatic province, on the contrary, is characterised by low heat flow values, negative bouguer anomalies, crustal thickness of about 35 km and abundant shallow seismic activity. Most of the earthquakes occur at a depth shallower than 15 km and show extensional focal mechanisms in the chain area and compressional kinematics in the periadriatic region (outer Marche belt). Also some compressional events are recorded down to 80 km in the western sector.
The transition between these two domains, characterised by different tectonic regimes, is quite sharp and occurs, in the Umbria area, along the Tiber valley graben.
Across the inner part of the northern Apennines, the following major domains are usually distinguished from W to E (figures 4 and 9):
a) Ligurian domain: consisting of an oceanic basement covered by a pelagic Upper Jurassic-Lower Cretaceous sedimentary succession;
b) sub-Ligurian domain: whose only remnant is the

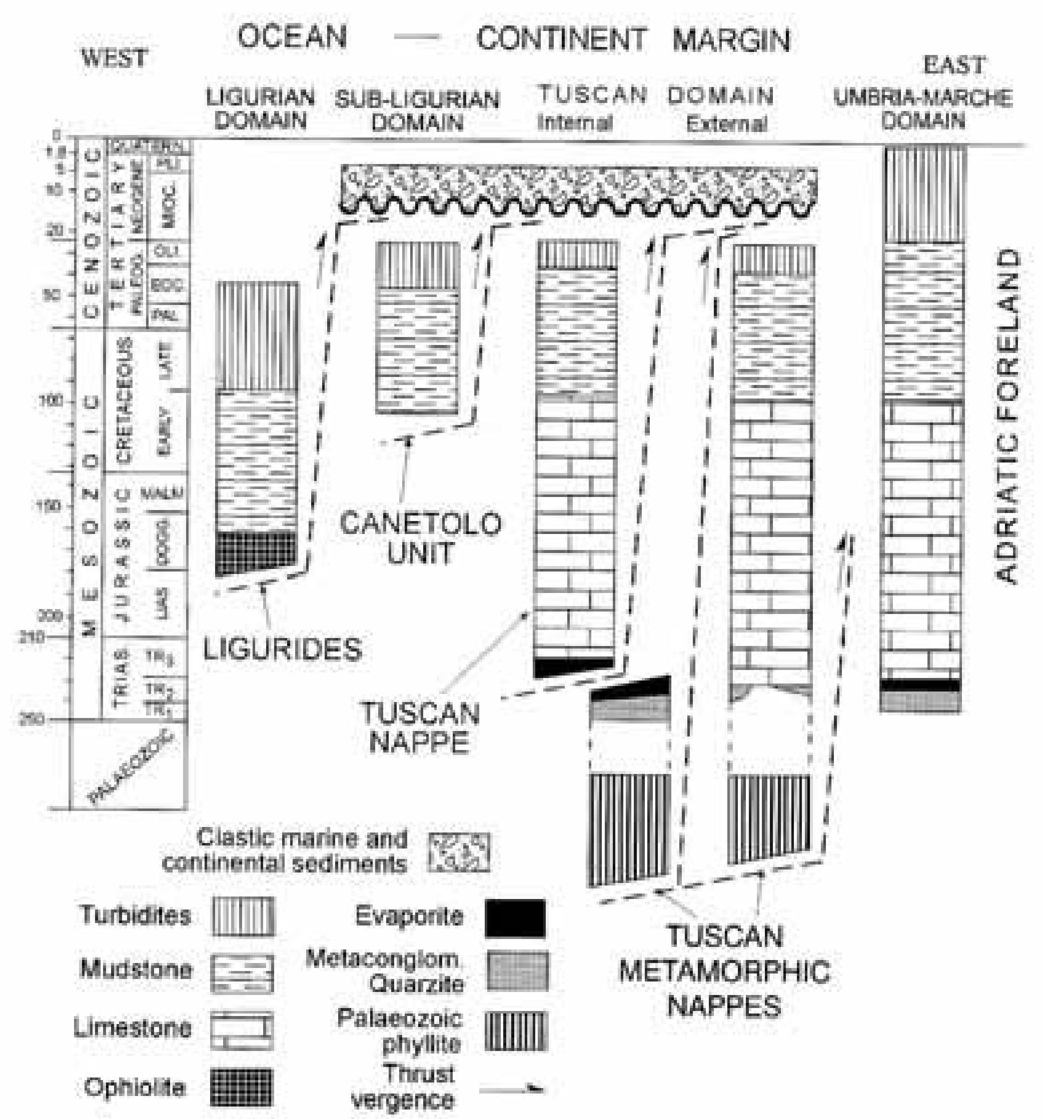

Figure 9 - Diagram showing the relations between the stratigraphic and tectonic units of the northern Apennines. The western tectono-stratigraphic units are progressively superimposed on the eastern ones (After Carmignani et al.,2001)

strongly deformed palaeogene sequence (Canetolo unit), referred to the transitional area placed between the Ligurian Oceanic domain and the continental Tuscan crust;

c) internal Tuscan Domain: made of a non metamorphic succession Upper Triassic–Lower Miocene in age; it is detached from its basement on the Burano Anhidrites and is deformed and translated in the east-verging Tuscan Nappe;

d) external Tuscan domain: forming metamorphic units in greenschist facies comprising Upper Carboniferous-Oligocene sedimentary succession and a Hercinyan basement.

e) Umbria domain: with a stratigraphy very similar to the internal Tuscan domain but characterised by a more recent turbiditic succession (Marnoso Arenacea Fm, Middle-Upper Miocene in age).

After the Late Oligocene, the onset of a compressional tectonic regime, due to the collision between the Sardinia-Corsica block and the Apulian plate margin, caused the following tectonic events: 1) emplacement of the Ligurian and Sub-Ligurian units on the Internal Tuscan Domain; 2) tectonization of the Internal Tuscan Domain, nucleation and thrusting of the Tuscan Nappe on the External Tuscan Domain; 3) deformation of this latter within an ensialic shear zone in low-grade metamorphic conditions (greenschist facies) (Carmignani et al., 2001 and therein ref.)

Within the area crossed by the fieldtrip, the collisional stage is documented in the structures of the metamorphic rocks outcropping in the Elba island and in the Montagnola Senese – Monticiano Roccastrada alignment (the so called Middle Tuscan Range) as well as in the sedimentary cover of Elba, Larderello and M. Amiata area. A pervasive schistosity, defined by sub-parallel orientation of micas, carbonate minerals and biotite, forms an axial plane foliation associated to E verging isoclinal folds. In all the area considered, Kinematic indicators show a consistent transport direction toward the NE.

Farther east the Tuscany-Umbria Apennines comprises the outermost and shallower compressional structures of the M. Falterona – M. Cervarola Nappe (Dallan Nardi and Nardi, 1972; Abbate et al., 1991; Damiani et al., 1991, Aruta et al., 1998; Brozzetti et al., 2002).

This latter consists of a strongly shortened imbricate fan, detached on the shallow incompetent layer of the Scaglia Toscana (Argille Varicolori o Scisti Policromi, Eocene – Lower Oligocene in age) and mainly composed by Upper Oligocene – Lower Miocene turbidites (Macigno Sandstones). During the Late Burdigalian, the Falterona Nappe overrode, for at least 40 km eastward, the inner part of the Umbria domain. Its leading edge, at present, is roughly located along the western side of the northern Val Tiberina.

Later this allochthon unit was folded and breached by the Upper Miocene compressional structures of the Umbria-Marche Apennine such as the Perugia Massifs thrust and the Gubbio anticline (Brozzetti et al., 2002; Brozzetti and Luchetti, 2002)

Tyrrhenian related extension in Tuscany and western Umbria

After the emplacement of the Elba stack and of the Middle Tuscan Range Unit (Upper Oligocene-Lower Miocene?), Tuscany was affected by extensional deformation characterised by both low-angle and high-angle normal faults.

For a long period, extensional tectonics in the northern Apennines had been underestimated but in the last two decades, new data have been collected suggesting a strong imprint of normal faults on present structure of the Tuscany and western Umbria area. Different interpretative models have been proposed to explain both the genesis of the neogene basin and the formation of the Tuscan “serie ridotta”.

According to some Authors (Bertini et al., 1991; Carmignani et al., 1994, 2001; Baldi, 1994;) at least two well-distinct extensional events and related normal fault systems would be recognisable in Tuscany: i) a first event, developed soon after the end of the compressional deformations (very early in the Middle Miocene), would have induced simultaneous ductile extension at mid-crustal level and brittle extension at shallow crustal depth; the generation of a core complexes (Apuane Alps, montagnola senese area) and the differentiation into a non metamorphic upper plate and a metamorphic lower plate would be favoured by the presence of the Burano Anhidrite-Calcare Cavernoso fm which, also during compression, acted as main decollement level; ii) a second event generated, during Tortonian and subsequent times, the high angle faults bounding the neogene basins, causing a minor amount of deformation.

Other Authors; supported also by the CROP03 interpretation (Lavecchia et al., 1984; 1988; Brozzetti, 1995; Barchi et al., 1998; Boncio et al., 2000; Collettini et al., 2002), suggested that extension induced by the Tyrrhenian rifting would originate different kinds of structures in response to a progressive deformation history but in the context of a single extensional phase.

These authors highlighted the presence of a set of east dipping LANFs, whose breakaway zones are roughly located along the western boundaries of the main neogene-quaternary grabens.

The westernmost of them, the Zuccale Fault on the Island of Elba, was active during Messinian because it cuts the Porto Azzurro Monogranite and the related dykes (6.0-5.4 Ma) but is displaced by high angle faults along which 5.3 Ma mineralisation is hosted (Keller and Pialli, 1990; Bortolotti et al., 2001). On the CROP03 line, its seismic reflection merge at a depth of about 8 sec twt into an extensional shear zone involving and stretching the deepest lower crust (Barchi et al., 1998).

On the contrary, the east-dipping LANF bounding westward the Siena-Radicofani basins is not well-defined at a depth greater than 3 sec twt (Barchi et al., 1998; Decandia et al., 1998).

Following this interpretation, the “serie ridotta” observed in the Monticiano-Roccastrada ridge and in the Montagnola Senese are, can be interpreted, as a stratigraphic omission caused by low angle faults splaying westward from the main plane (figure 10).

The eastern LANF, named Altotiberina Fault (AF) is well known in both its field evidence and seismic expression at least in the western Umbria segment which develops for about 70 km along the Val Tiberina - Valle Umbra basin (Brozzetti, 1995; Barchi et al., 2000; Boncio et al., 1998, 2000; Collettini et al., 2002).

West of the Tiber Valley basin, within the AF footwall, the triassic Burano Anhidrite is exhumed in the M. Malbe ridge whereas, in the hangingwall, several low-angle and high-angle splays severely stretch the Meso-Cenozoic carbonates. The Perugia Mesozoic massifs, which show the highest elevations of the Umbria pre-Apennines are in fact interpreted as a domino-like system tilted westward upon the AF plane and would correspond to the AF extensional allochthon. This latter gives rise, in the seismic lines, to an evident reflector which deepens beneath the Umbria-Marche carbonate fold and thrust belt to a depth of 12-14 km.

Detailed geophysical studies lead to interpret the AF as the main detachment horizon of all the extensional structures cropping out in more external areas (figure 11). The east-dipping faults would be AF splays branching at depths progressively greater from W to E; the west dipping faults instead, would belong to the antithetical conjugate system.

Also the east Umbria active and seismogenic west-dipping normal faults, among which the most

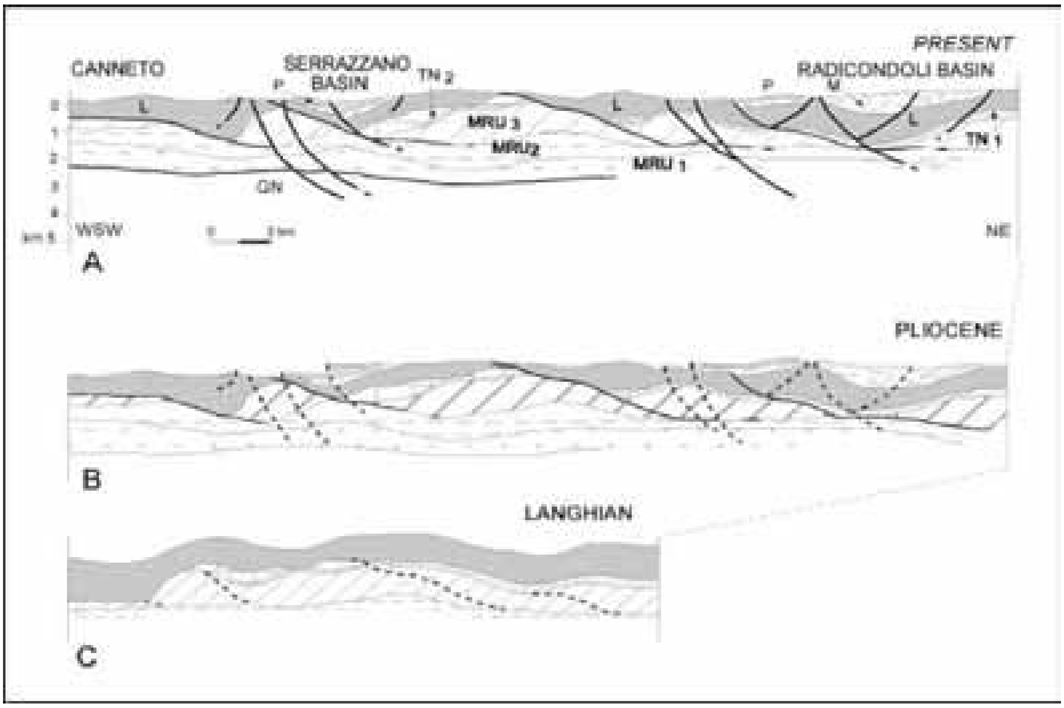

Figure 10 A - Geological cross sections through the Larderello geothermal field (constructed from borehole stratigraphy, seismic reflection lines and field mapping). Location indicated in Fig. 4. Pliocene normal faults cross-cut all earlier structures: P-Pliocene sediments; M -Late Miocene sediments; L - Ligurian Complex; Tuscan Nappe (TN), TN2 -carbonatic and terrigeneous sequence (Jurassic-Late Oligocene-?Early Miocene); TN1 -Late Triassic evaporites; Monticiano Roccastrada Unit (MRU), MRU3 - Mesozoic-Palaeozoic Group; MRU2 - phyllite-quartzite Group; MRU1 ×Hercynian Micaschist Group; GN - Gneiss Complex. B. Reconstructed Pliocene geological section. Note that Late Miocene sediments are preserved in tectonic depressions defined by synsedimentary normal faulting related to the second extensional event; Early and Middle Pliocene sediments rest unconformably on Late Miocene sediments which were deformed during their deposition. During the second extensional event, the deepest rocks were exhumed. C. Hipothetical reconstruction back to Langhian age (after Baldi et al., 1994b).

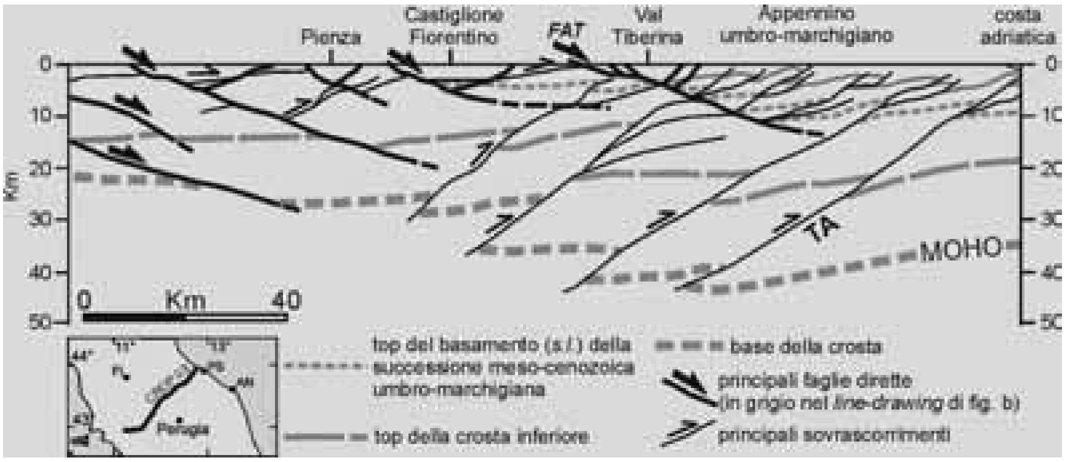

Figura 11 - geological interpretation of the Crop03 seismic profile showing a set of east dipping normal fault dissecting the compressional tectonic pile. The inner LANF corresponds to the Zuccale fault, the eastern one, to the Altotiberina fault.

important crop out along the Sansepolcro – Gubbio – Colfiorito alignment, detach on the AF which, furthermore, bounds downward the crustal volume affected by background seismicity.

An overview, synthesis and updating of geologic, geophysical and seismologic data available in literature suggest that the AF belongs to a regional east dipping fault system, named Etrurian Fault System which probably extends over 350 km along the strike and possibly controls the seismicity distribution in northern central Italy (Boncio and Lavecchia, 2000; Boncio et al., 2000).

Field Itinerary

DAY 1

From compression to extension, an E-W section from the East-Tenda extensional shear zone and the Schistes Lustrés
Laurent Jolivet

Stop 1:

Casta granodiorite, one magmatic protolith of the Tenda gneiss

One brief stop in the village of Casta shows a bad outcrop of a granodiorite with large magmatic amphiboles. This rock is totally undeformed. Rims of blue amphiboles around the hornblendes may locally occur either in the granodiorite or in intrusive basic dykes.

Stop 2:

Tenda orthogneiss with HP foliation and blue amphibole, F2 folds along the road to Santo Pietro (figure 12)

We shall leave St. Florent by road D81 toward the

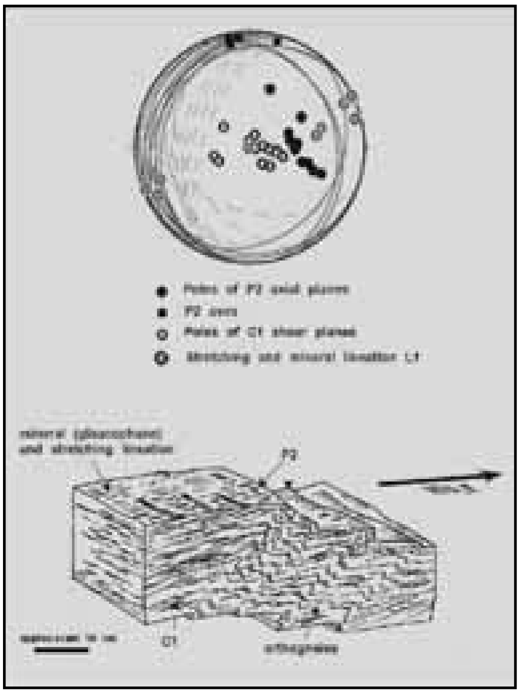

Figure 12 - 3D diagram of the structures observed at Fontain de Porraghia (stop 2).

west and then by D62 toward the south. This first stop of the day will show the high pressure foliation of the Tenda massif. Orthogneiss makes most of the massif. High pressure deformation is localized along narrow shallow dipping shear zones with a top-to the-W sense of shear contemporaneous with the crystallization of crossite. Most of those shear zones are not easily accessible with an excursion. On this outcrop we see the HP foliation folded by east-verging open folds similar to those seen on the eastern side of Cap Corse. Their axial planes dip slightly toward the W. These folds are found at various scales of the ridge of the Tenda massif. They will evolve toward sheath folds toward the ETSZ. They represent the first stages of the top-E shear.

Stop 3:

Sheared folds in the East Tenda Shear Zone (figure 13)

We go back to St. Florent and drive along a track that goes northward near the coast. We stop above Fornali. The high-pressure foliation here is more intensely folded and the folds have parallel limbs. The axial plane and the associated crenulation cleavage now dip eastward. Localized shear bands show a top-E sense of shear. The crenulation cleavage carries an E-W stretching lineation.

Stop 4:

Walk along the coast north of Fornali, ETSZ top-to-the-east-shear (figure13)

We then go down to the coast and walk northward until we reach an outcrop of intensely sheared gneiss (figure 14). The HP pressure foliation is still visible within small asymmetric sheared lenses. A late mylonitic foliation wraps around these lenses and is cut by east-dipping extensional shear bands. The mylonitic foliation contains a strong E-W stretching lineation. Curved fold axes and curved S1-S2 intersection lineation are the indication for a high shear strain. A walk further north would show the mylonitic foliation cut by steeper more brittle extensional shear bands (figure 15). The ETSZ thus shows a clear gradient of deformation.

Radiometric ages of micas evolve from ca. 45 Ma where the HP foliation is well preserved to 25 Ma within the ETSZ. The base of the St. Florent basin is dated to the Lower Miocene (approx. 20 Ma).

The role played by micas has recently been reassessed in F. Gueydan's thesis (Gueydan, 2001; Gueydan et al., 2001; Gueydan et al., 2003b) (figure 16).

Localizing shear zones at or near the brittle-ductile transition is quite frequent in post-orogenic extensional context during the formation of metamorphic core complexes (Mitra, 1978; Miller et al., 1983; Lister and Davis, 1989; Gautier and Brun, 1994; Rigo et al., 1996; Jolivet et al., 1998; Jolivet and Patriat, 1999). Yet, classical rheological envelopes (Goetze and Evans, 1979; Kuznir and Park, 1987; Le Pichon and Chamot-Rooke, 1991; Ranalli, 1995) suggest a maximum of resistance at the depth . Destabilizing mechanisms should thus be looked for to explain the localisation of shear zones. As suggested earlier by several authors (Mitra, 1978) the crystallization of weak phases such as micas can lead to localisation. The adjunction of a simple metamorphic reaction increasing the micas content of a quartz-feldspar-micas mixture supporting the same strain rate within a thermo-mechanical model indeed leads to the localisation of shallow-dipping extension shear zones within the brittle-ductile transition (Gueydan, 2001; Gueydan et al., 2003b). The transformation is linked to the fracturation of feldspar and fluid influx.

Stop 5:

St. Florent limestone and conglomerate near Farinole

We then reach the unconformable St. Florent Burdigalian limestone and its basal conglomerate (figure 17). The conglomerate is made of pebbles issued from Western Corsica and from the Alpine units, including the eclogites. Pebbles are coated with algal calcareous concretions. The limestone itself is bioclastic and contains shallow water fossils (Orszag-Sperber and Pilot, 1976; Ferrandini et al., 1996). A quite important component of detrital quartz is present throughout. Syn-sedimentary normal faults offset the beds of conglomerate and limestone (Jolivet et al., 1991).

Stop 6:

Section along the old road of Lancone defilé

Stretched pillows, pillow breccia, blueschist paragenesis top-to-the-west shear.

The Lancone défilé is located south of Saint Florent and Bastia and crosses a thick glaucophanite unit. The whole structure is folded in a broad antiform with a N-S axis. The deep incision of the Bevinco river shows several stacked oceanic and continental units. The regional foliation is affected by east-verging open folds on the eastern flank of the antiform. These F2 folds are similar to those seen on the eastern coast of

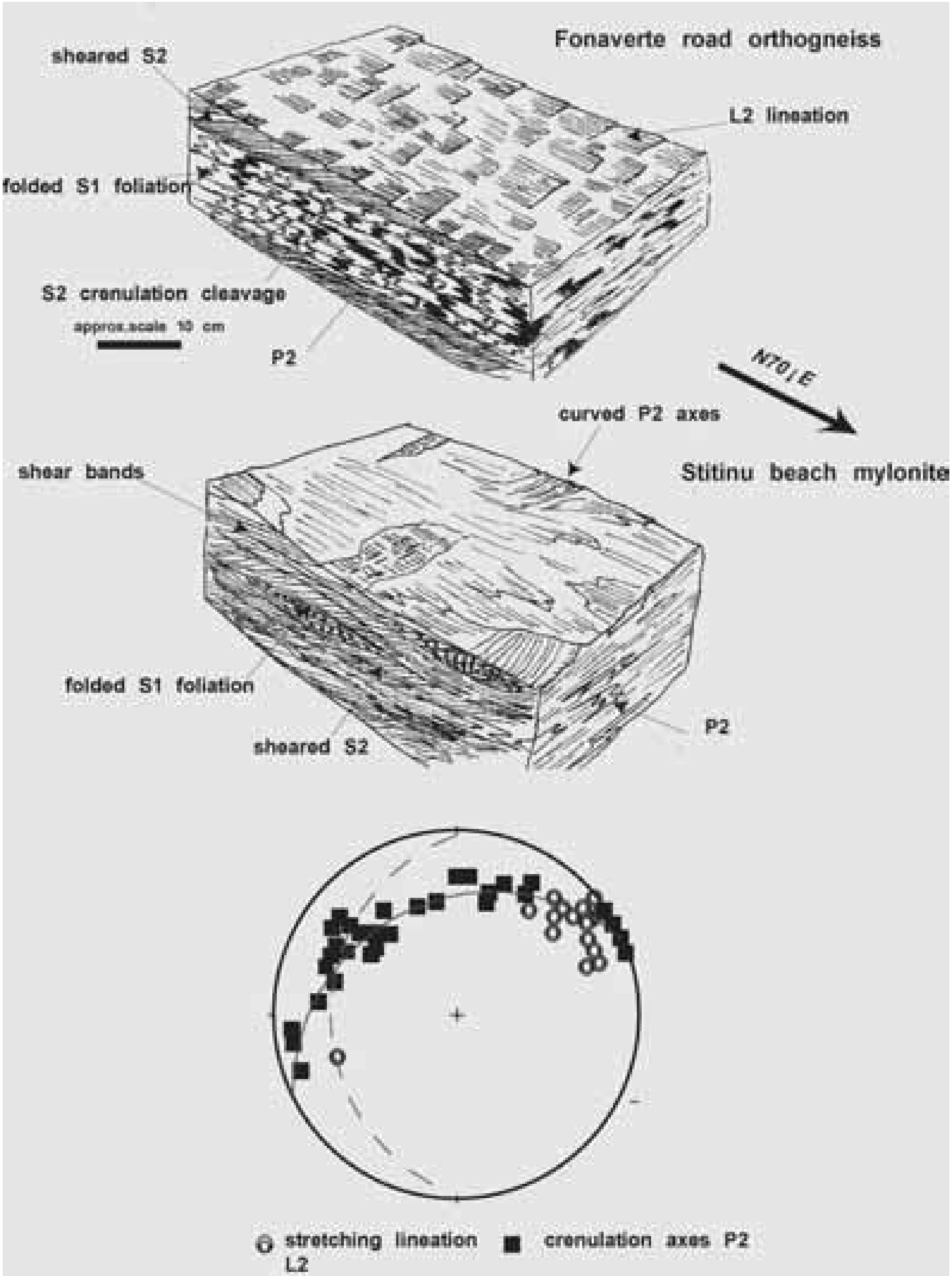

Figure 13 - 3d diagrams of structures observed in the East Tenda Shear Zone (stops 3 & 4).

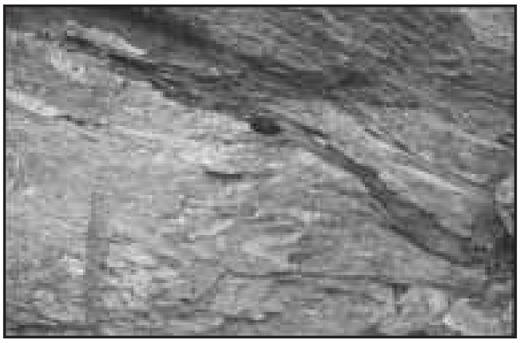

Figura 14 - Shear orthogneiss in the East tenda Shear Zone, top to the east asymmetric flods and east-dipping mylonitic foliation (camera cap for scale). East on the right.

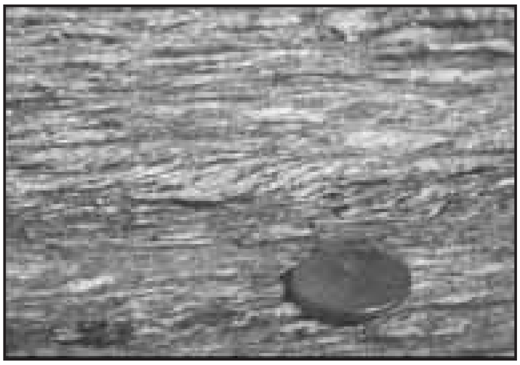

Figure 15 - Mylonitic foliation and top-to-the-east kinematic indicators in the most deformed part of the east tenda Shear Zone (East to the right)

Cap Corse during the second day.
From base to top we see the following tectonic units:
(1) The lowermost unit is made of a thick pile of metabasalts (basaltic flows, pillow breccias and pillows) showing characteristic blueschist parageneses (Lancone metabasites).
(2) It is covered with a metasedimentary unit (Inzecca Schistes Lustrés) made of calcschists.
(3) A thick unit of serpentinites covers underlying units and its basal contact truncates the lower ones.
(4) Then comes a lens of orthogneiss and their metasedimentary cover (mostly marbles and quartzites) which culminates at Mt Zucarello (Sera di Pigno - Oletta gneiss). This unit is the equivalent of the large gneiss unit which crops out further north between Sera di Pigno and Oletta. Here it is extremely thinned (100 m at most). Its eastern part is folded by the F2 folds together with the calcschists.
(5) The uppermost unit along this section is again made of glaucophanites.
The Lancone metabasites are continuously exposed along the old road on the northern side of

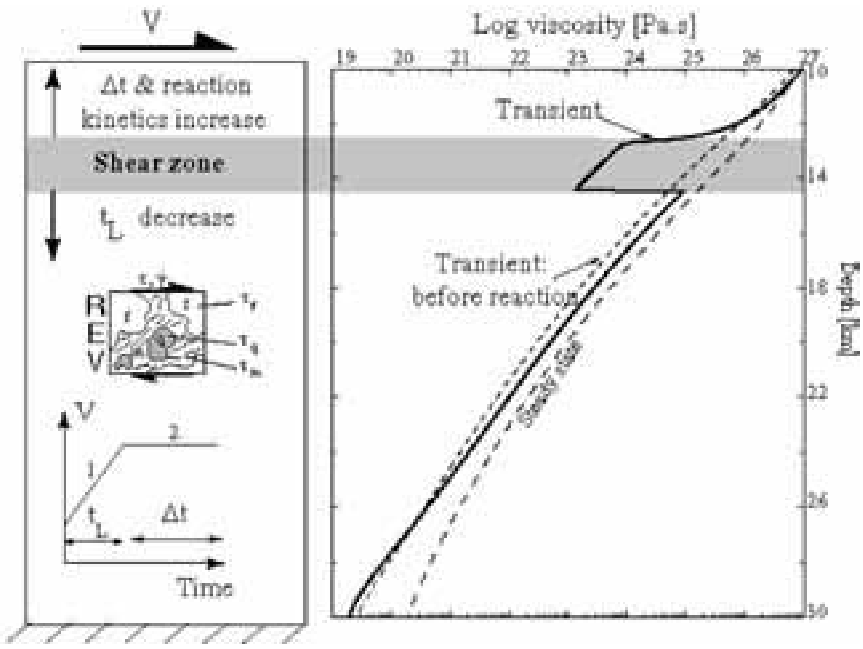

Figure 16 - Summary of the thermomechanical approach of Gueydan (2001) explaining the localisation of shallow-dipping shear zone at the brittle-ductile transition.

Figure 17 - Basal conglomerate of the Saint Florent limestone near Farinole.

the défilé. We shall stop in the serpentinites and walk down-section to the east.
The first metabasite outcrop shows highly retrograded metabasalts with mostly greenschist parageneses. Walking eastward the greenish colour of the rock is replaced by the deep blue of well-preserved glaucophanites. In this region a general rule is the retrogression of blueschist parageneses along major tectonic contacts or along minor shear zones inside blueschist units. We shall see one such narrow shear zone in the middle part of the section. The greenschist retrogression is associated with a N45°E trending stretching lineation and top-to-the-NE sense of shear.
After this contact a succession of massive basalts, pillow breccia and pillow lavas crop out. Massive basalts are strongly foliated and show a well preserved blueschist paragenesis with glaucophane, garnet and lawsonite. Pillow breccias and pillows are stretched along a NE-trending direction. Sections perpendicular to the stretching lineation show well preserved pillow shapes with polarity criteria. Pillow breccias show clasts of green fine grained basalts with a blue rim and a blue foliation. Sections perpendicular to the foliation and parallel to the stretching lineation show syn-blueschist top-to-the-SW shear bands

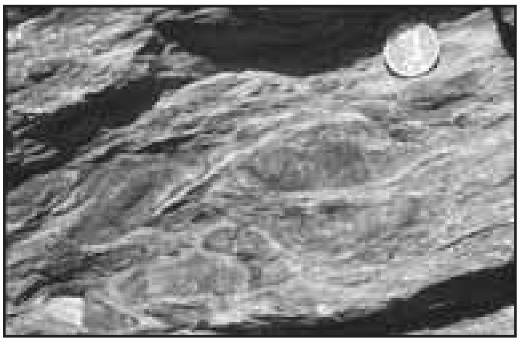

Figura 18 - Sheared pillow-breccia in the Lancone Gorges. Blueschist imprint is visible in the matrix and in the rim of sheared clasts.

and asymmetric clasts (figure 18). These are good examples of the fist stage deformation which is rarely so well preserved.

Stop 7:
Panorama on the col de Teghime, the Saint Florent basin and Tenda massif

A stop on road D38 will show us a complex panorama on the Sera di Pigno - Oletta gneiss unit, its metasedimentary cover and the underlying gabbros and peridotites to the NE, as well as on the St Florent basin and Tenda massif to the W (figure 19).
From East to West:
The metasedimentary cover of the Sera di Pigno - Oletta gneiss is made of marbles, schists and quartzites which crop out behind us on the crest line of Monte Secco. From Col de Teghime to the foot of Cima Orcaio to the north a thin band of marble and quartzites is squeezed between the main body of gneiss and the Cima Orcaio gneiss which correspond to the core of a large synfolial fold (figure 20). Sera di Pigno is made of orthogneiss recrystallized in the blueschist facies (glaucophane in metabasites and jadeite + quartz in acidic gneiss, remains of Fe-Mg carpholite in the metasediments) and variably deformed.
Gabbros and serpentinite rest above the gneiss unit and are folded together with it along the Malaspina section to the north. A complex imbrication of marble and serpentinite by sheath folds can be observed near the contact. In the background the summits of Cima di Gratera show massive serpentinites and gabbros.

Figure 19 - Panorama on the Saint Florent Basin and the East Tenda Shear Zone.

The village of Patrimonio and its vineyards are settled on the contact between the Schistes Lustrés nappe (*lato sensu*) and the Balagne nappe which crops out in the Nebbio depression. Olistoliths of basalts and Jurassic limestones are visible between the vines.
Monte San Angelo is made of the west-dipping St. Florent limestone of Lower Miocene age.
In the distance, west of the St. Florent gulf, the Tenda massif is visible. Its eastern side shows an east-dipping foliation which corresponds to the East-Tenda Shear Zone which we shall visit tomorrow. The core of the Tenda massif is less deformed. It is made of several Palaeozoic intrusives. Monte Genova which sticks out in the Désert des Agriates is a Permian alkaline granite. The summit of the Tenda massif is Monte Asto further to the south. If the weather is clear enough we will be able to see the E-dipping contact between the Tenda massif and the Schistes Lustrés nappe near Santo Pietro di Tenda.
The background shows the high summits of Western Corsica with its Palaeozoic intrusives and Permian volcanics.

Figure 20 - Spectacular synfolial fold in the metasedimentary cover of the Oletta orthogneiss near Col de Teghime (first described by Mattauer et al. (1981).

Stop 8:

Roadside outcrops along the road to the television antenna at Sera di Pigno, protolith of the Jadeite-quartz bearing orthogneiss

Along the narrow road D338 which climbs to the television antenna of Sera di Pigno we shall stop in the little deformed protolith of the Sera di Pigno gneiss. A mixture of basic and acidic lithologies shows the original aspect of the alpinized continental basement. Pegmatites, amphibolites and granitoids show only localized shear zones. Acidic lithologies contain jadeite + quartz associations and basic ones show rims of blue amphiboles on older hornblende.

Stop 9:

Deformed orthogneiss with glaucophane

A gradient of deformation is visible along the same road. We shall stop at the foot of Cima Orcaio below metasediments. A metamorphosed granodiorite shows a well developed foliation and a strong stretching lineation. The basic parts of the granodiorite are recrystallized in the blueschist facies. Hornblendes are totally replaced with glaucophane which makes large clasts and smaller needles of blue amphiboles (mostly crossite) grow within the pressure shadows.

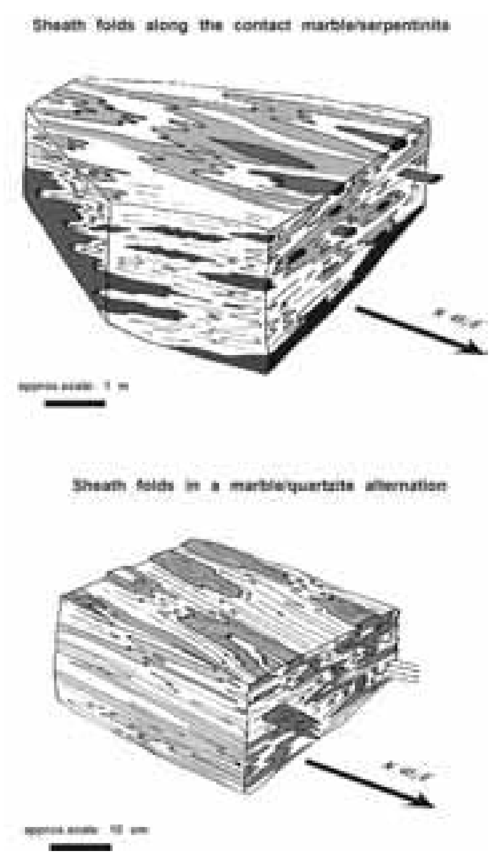

Figure 21- 3D diagrams of prograde structures seen along the contact between oceanic units and continental units in the Schistes Lustrés (stops 8, 9 & 10).

Top-to-the-west sense of shear is seen in thin section. A nearby section shows spectacular deformation along the contact between the metasediments and the overlying serpentinite (figure 21).

Stop 10:

Walk along the ridge from Sera di Pigno to Monte Pinatelle

We may skip this stop if we are short of time. This would be unfortunate because the walk offers very spectacular views on the St. Florent Gulf to the west and the Tyrrhenian Sea to the east, with several islands visible in the distance (from south to north: Monte Cristo, Pianosa, Elba and Capraia) and nice geology. We cross several contacts along the ridge from the Sera di Pigno - Oletta gneiss to the Farinole eclogites. E-W stretching is seen everywhere. Gradients of strain and of greenschist retrogression are observed. The section ends in the Farinole eclogitic gneiss at the summit of Monte Pinatelle where they were first described by D. Lahondère. We shall see equivalent outcrops near Bracolaccia but less spectacular. A complete succession of parageneses and deformations is seen there with eclogite boudins reworked in a blueschist foliation associated with top-to-the-SW sense of shear and sheath folds, as well as late greenschists recrystallization and top-to-the-NE sense of shear.

Stop 11:

Sheath folds in marbles and quartzites below the Cima Orcaio klippe

We shall stop the cars along road D81 east of Col de Teghime. Outcrops of marbles and quartzites make a cliff above the road. A short walk to this cliff will show spectacular sheath folds with closed eye structure in section perpendicular to the stretching lineation. The same kind of folds can be seen below the Malaspina ridge. Large scale synfolial folds crop out on the eastern side of Monte Secco.

Stop 12:

Panorama on the eclogite units near Bracolaccia

Two eclogite-facies units crop out near Bracolaccia. One is the Farinole gneiss which is continuous with the Monte Pinatelle one. The second is structurally below the gneiss with some intervening serpentinite; it is made of eclogitized iron-rich gabbros. The landscape from the village of Sparagaggio shows the superposition of several units dipping to the south (from N to S and base to top):

(1) foliated serpentinite
(2) eclogitized metagabbro
(3) a thin slice of serpentinite
(4) eclogitized orthogneiss
(5) serpentinite and gabbros of the Gratera unit and slices of metasediments within the basal contact.

Stop 13:

Farinole eclogite below Bracolaccia

Below the village of Bracolaccia we shall touch both eclogitized lithologies.

Along the road a succession of acidic gneiss and glaucophanites represent the Farinole gneiss. Leucocratic gneiss contain the association jadeite + quartz partially retrograded into albite. Meta-granodiorites show the replacement of magmatic amphiboles by garnet and omphacite. A NE-trending stretching lineation is visible in the glaucophanite with top-to-the SW sense of shear. Lenses and boudins of eclogite are included within the acidic matrix of the gneiss. The boudins contain a folded foliation (with sheath folds at Monte Pinatelle) made of glaucophane, omphacite and small garnets. A blueschist foliation wraps around the boudins and is cut by south-dipping extensional shear bands which contain mostly quartz and glaucophane. This illustrates the decrease of pressure and temperature during the top-to-the-SW shear. Along a narrow path we will then touch the eclogitized gabbros. The rock is iron-rich and dark. Characteristic associations contain glaucophane, garnet, omphacite and lawsonite. They represent low temperature eclogites. They can be usefully compared with the Giotani metagabbros which are Mg-rich and which did not develop a blueschist paragenesis though they belong to the same tectonic unit.

DAY 2

A trip around Cap Corse

Today's outcrops will illustrate the first deformation stage, contemporaneous with the obduction of oceanic units onto the continental basement of Corsica, and the high pressure and low temperature metamorphism, as well as the retrogression and top-to-the-east reworking during post-orogenic extension.

Stop 14:

Erbalunga, calcschists (Castagniccia Schistes Lustrés) and gabbros, retrograde metamorphism and top-to-the-east shear, gabbros, blueschist metamorphism and top-to-the-west shear

North of Bastia the village of Erbalunga offers two spectacular outcrops of the Schistes Lustrés nappe.

We will park the cars in the village and walk down to the shore. The first outcrop is used for sunbathing and the second is below a ruined tower, the village as a whole is a frequent sight-seeing stop for tourists. So please do not used your hammer here.

The first outcrop is made of calcschists showing a clear shallow east-dipping foliation and a stretching lineation trending N140°E (figure 22). Stretching is shown by elongated quartz and calcite lenses, foliation boudinage, tension gashes (figures 23 and 24). Fold axes are parallel to the stretching direction, and folds are flattened in the foliation plane. Curved fold axes can be seen in the foliation plane. Top-to-the-NE sense of shear is shown by en-echelon tension gashes and extensional shear bands.

The main foliation is an S2 and made of retrograde metamorphic minerals, such as white micas, chlorite.

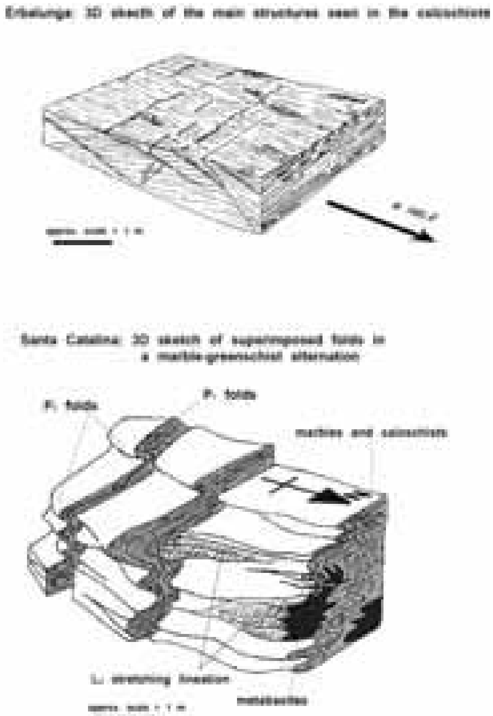

Figure 22 - 3D diagrams of structures seen along the eastern coast of Cap Corse (Erbalunga stop 14 and Sisco Stop 15).

Figure 23 - Stretching lineation in the Erbalunga calcschists.

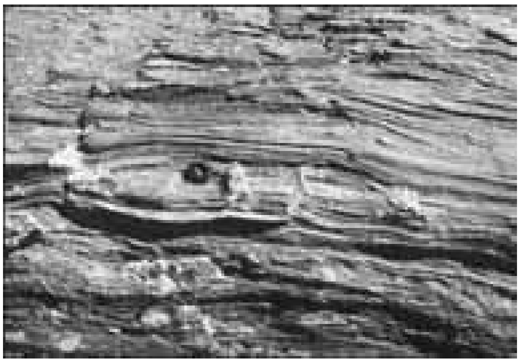

Figure 24 - Stretching and boudinage in the Erbalunga calcschists.

No remains of high pressure minerals are left here. The lithology is otherwise similar to the main body of the Castagniccia Schistes Lustrés which crop out in the core of the Cap Corse and Castagniccia antiforms. Retrograde metagabbros can be seen at the southern end of the outcrop.

The second outcrop shows metagabbros and metadolerites with well preserved HP-LT parageneses, especially here a light-blue glaucophane which grows within the main foliation and between the clasts of stretched pyroxenes and inside pressure shadows. The same direction of stretching as in the calcschists can be observed however with a reverse sense of shear, top-to-the-NW. Clear examples of asymmetric pressure-shadows filled with glaucophane are visible on the outcrop.

Stop 15:

Sisco: F1 and F2 folds in a marble-greenschist alternation (figures 25, 26 and 27)

The next outcrop, below the black virgin of Santa

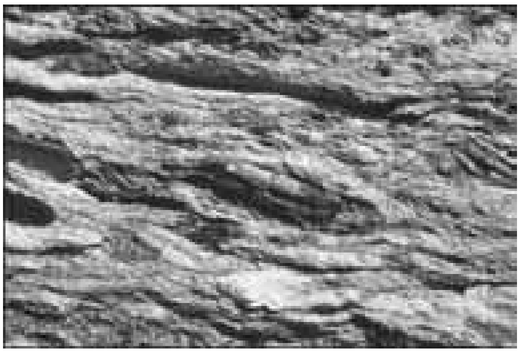

Figure 25 - Asymmetric pressure shadows filled with glaucophane around a pyroxene at Erbalunga (west to the left).

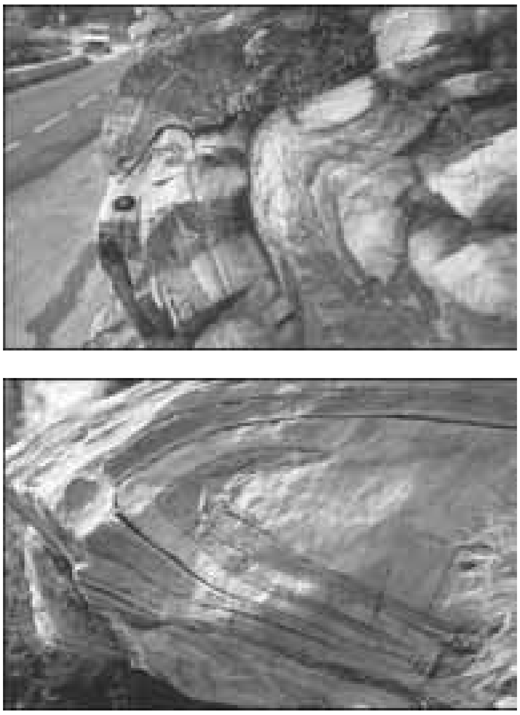

Figure 26a 26b - Superimposed folding near Sisco

Figure 27 - Post-orogenic folds reworking the HP foliation near Sisco (Stop15: see details in the two preceeding pictures (east to the right).

Catalina, shows the superposition of two stages of folding. A first episode of synfolial folds is reworked by a second stage of open folds with slightly curved axes and horizontal axial planes. F2 folds are asymmetric and verge eastward. They fold a stretching lineation shown by elongated albite porphyroblasts in the greenschists. Remains of pulled-apart glaucophane porphyroclasts will also be observed in the best preserved examples.

Stop 16:

Macinaggio, Tamarone beach, basal contact of the upper allochthon, east-dipping normal faults

We drive along the coast until we reached the village of Macinaggio. From there we drive along a dirt track which leads to Tamarone beach. We are now along the contact between the Schistes Lustrés, the deepest unit of Cap Corse and the uppermost unit of Alpine Corsica, the Balagne Nappe (figure 28). Three small klippens of the upper allochthon are preserved along the eastern coast south and north of Macinaggio. We shall see the basal contact south of the beach.

A shallow east-dipping fault with a fault gouge separates the folded Schistes Lustrés in the footwall from the basal breccia of the nappe.

East-dipping small scale normal faults root in the fault gouge and determine small tilted blocks. On the fault plane a badly preserved stria can be measured, its trend is N110°E. This kind of structure is often seen along the contact elsewhere in the same area. The last motion along the nappe contact was thus

Figure 28 - Extensional contact between the Schistes Lustrés (footwall) and the "Balagne nappe" (hangingwall) (east to the left).

toward the east along a shallow-dipping normal fault. This contact cuts down from the uppermost non metamorphosed unit to the deepest unit. The next stop will show that this geometry is characteristic of the whole Cap Corse.

Stop 17:

Panorama from Ersa toward the north, Centuri gneiss

This stop will show an E-W cross-section of Cap Corse near its northern tip. From the left to the right the following units can be seen:

(1) the Monte Maggiore peridotites and gabbros (Gratera serpentinites and gabbros) in the distance. Small klippes of serpentinized peridotites occur further to the right, for example near the summit of Monte San Antonino, in front of us.

(2) The Centuri gneiss (Sera di Pigno-Oletta gneiss) which form a large lens dipping westward. They end below the San Antonino klippe. The west-dipping foliation is curved and rotated parallel to the basal contact of the serpentinite. A strong stretching and top-to-the-east sense of shear are observed there.

(3) Retrograded metagabbros (Lancone metabasites) with some remains of high pressure and low temperature parageneses (garnet glaucophanites).

(4) The schistes lustrés (Castagniccia Schistes Lustrés)

(5) the Balagne nappe

All contacts cut down section toward the east and a predominant top-to-the-east sense of shear is seen almost everywhere. The Balagne nappe cuts down to the core of the structure as seen at stop 3. This extreme thinning and associated top-to-the-east sense of shear are coeval with an almost complete retrogression into the greenschist facies.

Stop 18:

Gabbros of the Monte Maggiore klippe near the Capo Grosso lighthouse (Gratera serpentinites and gabbros)

The Monte Maggiore peridotites and gabbros crop out near the light house. We shall see spectacular gabbros with pluricentimetric crystals of pyroxenes.

Stop 19:

Highly sheared gneiss near the base of the Centuri klippe (Sera di Pigno - Oletta gneiss)

A small quarry west of Orche shows highly sheared acidic gneiss and amphibolite near the base of the Centuri gneiss unit (figures 29, 30). The alternation

Figure 29 - Sheared gneiss near Centuri (east to the right).

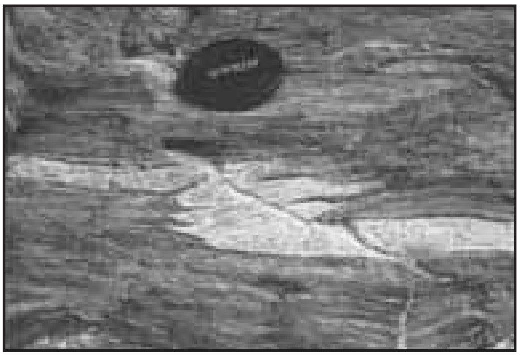

Figure 30 - Asymmetric fold in the sheared gneiss near Centuri.

of gneiss and amphibolite shows foliation boudinage, asymmetric boudins and asymmetric folds compatible with a top-to-the-east shear.

Stop 20:

Walk along the coast from Centuri to the north

The coastline north of Centuri shows a cross-section from the Schistes Lustrés to the Centuri gneiss. Starting from metasediments (Castagniccia Schistes Lustrés), highly stretched and flattened with symmetrical shear bands indicating an E-W stretching, the section goes through retrograded metagabbros with some remains of blue amphiboles. We then see metasediments again with clear top-to-the-east kinematic indicators. The section ends at the base of the Centuri gneiss where a more coaxial deformation is seen. The cliff would then show a gradient of deformation and metamorphism in the Centuri gneiss. The base is strongly retrograded, and blue amphiboles appears only as clasts within the foliation. The middle part of the cliff shows well preserved blue amphiboles. The HP metamorphism then decreases upward and some remains of high temperature variscan parageneses are found (garnet, biotite, amphiboles...). We are then in the middle of the Centuri gneiss unit which has been preserved from the prograde and retrograde alpine events.

Stop 21:

Giottani metagabbros (Lancone metabasites), spectacular green pyroxenes and N-S stretching

We then drive along a very scenic road on the western coast of Cap Corse until we reach white magnesian-gabbros with large (up to several tens of centimeters) crystals of green pyroxenes (smaragdite). A clear north-south stretching can be seen. Pressure shadows on pyroxenes contains large fibers of tremolite. The high magnesian content of the gabbros does not permit the crystallization of blue amphibole though this unit has seen eclogitic conditions as seen in some other lithologies.

DAY 3

The Zuccale East-dipping low angle normal fault.
Cristiano Collettini

Geologically, the structure of Elba (figure 31) consists of five thrust sheets stacked during an earlier Upper Cretaceous to Lower Miocene Apenninic compressional phase that are cross-cut by later Middle Miocene-Lower Pliocene extensional faults (Trevisan 1950; Keller & Pialli 1990). Two extensional faults, both of which are thought to have been active as easterly-dipping, low-angle structures are considered to be particularly significant with kilometer-scale displacements: the Capanne fault (Daniel & Jolivet 1995, Keller & Coward 1996; also known as the Central Elba fault, see Bortolotti *et al.*, 2001) and the Zuccale fault (Keller & Pialli 1990; Keller *et al.,* 1994; Keller & Coward 1996). On Elba, extension was broadly associated with intrusion of granite porphyry sheets (ca. 8 Ma; Rocchi *et al.,* 2002), the Monte Capanne granodiorite (ca. 7 Ma; Saupé *et al.,* 1982) and the Porto Azzurro monzogranite with associated dykes (ca. 5.1 Ma; Saupé *et al.,* 1982) (figure 31). The two main extensional detachments display different relative age relationships with these intrusions. The earlier Capanne fault is thought to have formed synchronously with emplacement of the Monte Capanne pluton at 7 Ma and to have been up-domed by the intrusion of the Port Azzuro monzogranite (Daniel & Jolivet 1995). By contrast, the later Zuccale fault appears to cross-cut all intrusions (Pertusati *et al.,* 1993).

The recognition and definition of the Zuccale fault by Keller & Pialli (1990) confirmed the existence of a regionally significant low-angle normal fault in Elba as first proposed by Trevisan (1950). Since then, the fault has been interpreted and discussed by many authors (e.g. see Bortolotti *et al.,* 2001 and references therein), although there appears to be some disagreement concerning the relative importance of the Capanne and Zuccale faults (e.g. contrast models in Daniel & Jolivet 1995 and Keller & Coward 1996). It is generally agreed that the Zuccale fault is a gently east-dipping fault that offsets part of the pre-existing thrust stack, previously intruded by 8 Ma porphyry sheets, eastwards (Trevisan 1950). According to Keller & Coward (1996) stratigraphic separations on

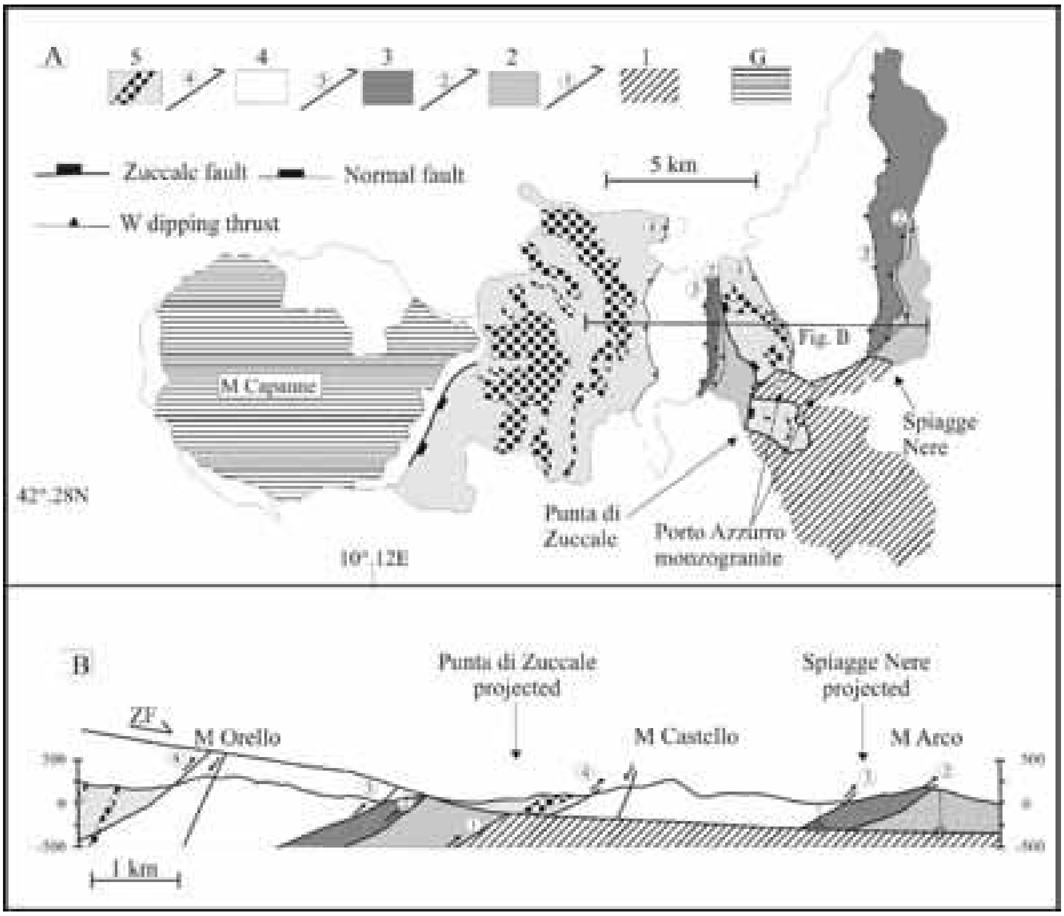

Figure 31 - A: Schematic geological and structural map of the Isle of Elba: G, Mt. Capanne granodiorite (ca. 7 Ma) and Porto Azzurro monzogranite (5.1 Ma); 1 = Complex I, basement schists; 2 = Complex II, Tuscan metamorphic sequence; 3 = Complex III, Tuscan carbonate sequence; 4 = Complex IV, Ligurian ophiolites; 5 = Complex V, Late Cretaceous flysch intruded by granite porphyry (ca. 8 Ma). B: Geological cross-section through central and eastern Elba. The section has been constructed along section 1 of the geological map of Elba (Trevisan et al. 1967).

Elba island suggest normal fault offsets of 7 – 8 km. These authors also suggest that the Zuccale fault can be recognised offshore in a seismic profile, located 20 km south of Elba. If this correlation is correct, the age range of associated syntectonic sediments in hanging-wall basins offshore implies that normal fault activity occurred from 13 to 4 Ma (Keller & Coward 1996; Pascucci *et al.,* 1999). This suggests an approximate long-term slip rate for the Zuccale fault in the range 0.8-0.9 mm/yr.

Stop 22:

Punta di Zuccale:

We will park at Punta di Zuccale car park and we will walk down to the Zuccale beach.

At Punta di Zuccale (figure 32A), the ZF separates a hangingwall sequence of Upper Cretaceous flysch (Complex V) from a footwall of Palaeozoic basement schists and Triassic sediments of the Verucano formation (Complex I). Fault-related frictional deformation is observed in both the hanging-wall block, affected by N-S trending thrusts and normal faults, and in the footwall block, represented by small displacement E-dipping normal faults and joints (figure 32B, stereoplots 1, 2). The basement foliation and lineation dip SW and pre-date extensional movements (figure 32, stereoplot 2).

The ZF is characterised by strongly foliated fault rocks (e.g. figure 33) derived from cataclasites and various protolith lithologies (carbonates, ultramafics) that were entrained from the pre-existing thrust stack (Complexes II-IV) into the detachment zone.

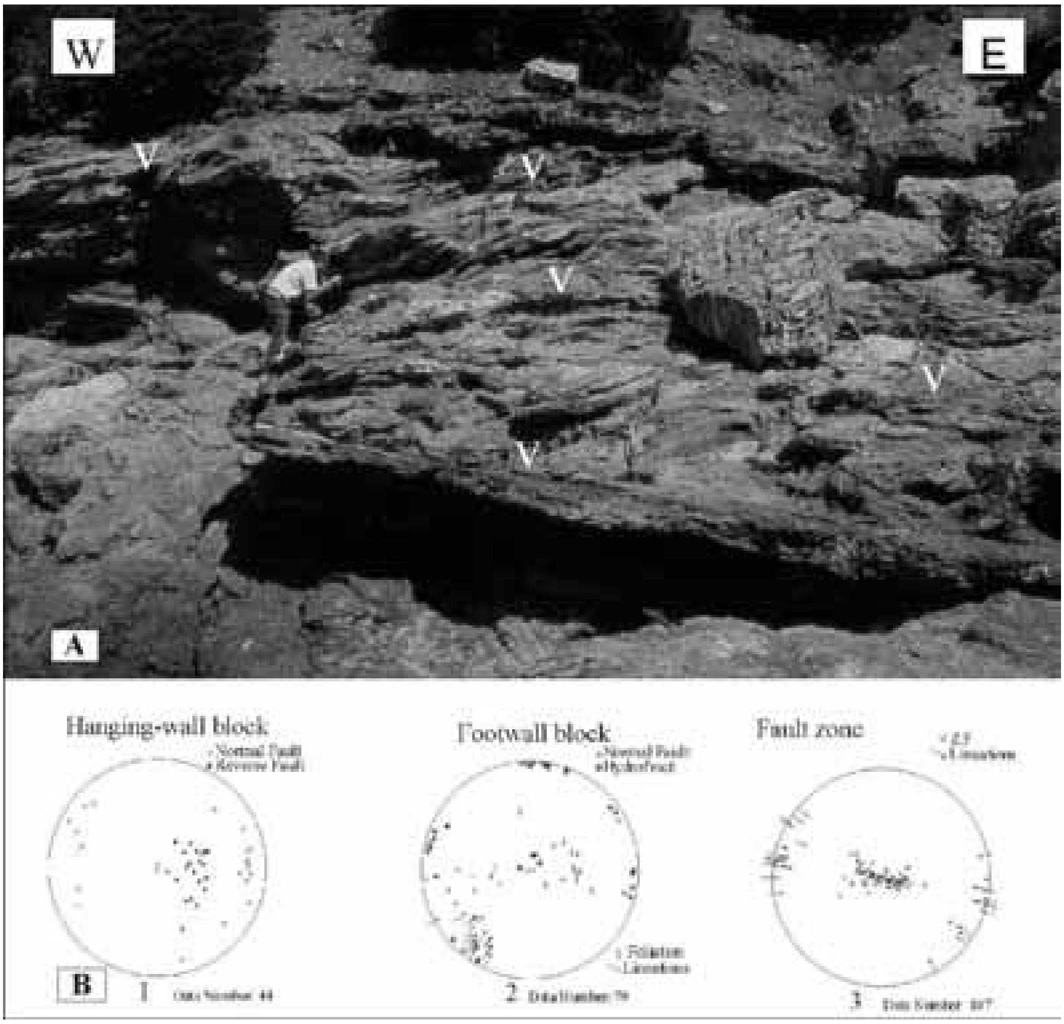

Figure 32 - A: The ZF at Punta di Zuccale showing a foliated nature of fault zone and sub-vertical carbonate veins (v). B: Equal area stereoplots (lower hemisphere projection) of structural data, with planes plotted as poles, collected at Punta di Zuccale: 1) normal faults and reverse faults in the hanging-wall block; 2) basement structures; 3) ZF fault plane and lineation. The ZF dips up to 15° east in offshore, but appears to be undulatory, so it is locally subhorizontal or gently west-dipping. The mean trend of the ZF at the surface is ~ N-S and the distribution of the slickenlines (striae and aligned calcite fibers) gives a direction of movement N110°E showing a kinematics close topure dip-slip (Figs. 2B steroplot 3).

Carbonate veins preserve spectacular crack-seal textures, demonstrating that cyclic build-ups in fluid overpressure followed by hydrofracturing occurred during fault activity (Collettini and Barchi, 2004). Two main vein systems can be recognised (figure 33A, stereoplot 1): an older system, which lies parallel to the foliation, cross cut by a younger system. This later system consists of three different sets: two sets are subvertical, one parallel (i. e. ~E-W trending) and one perpendicular (i. e. ~N-S trending) to the slip-direction marked by the slickenlines; the third set is subhorizontal and parallel to the base of the fault-related shear zone. The N-S trending veins are locally deflected into west-dipping segments (figure 33A). Microstructural studies show that earlier veins are overprinted by ductile deformation and are progressively sheared into concordance with the foliation. Kinematically, the fault zone (figure 33B) is characterised by abundant C-type and C'-type shear bands bounding S foliated domains (figure 33B,

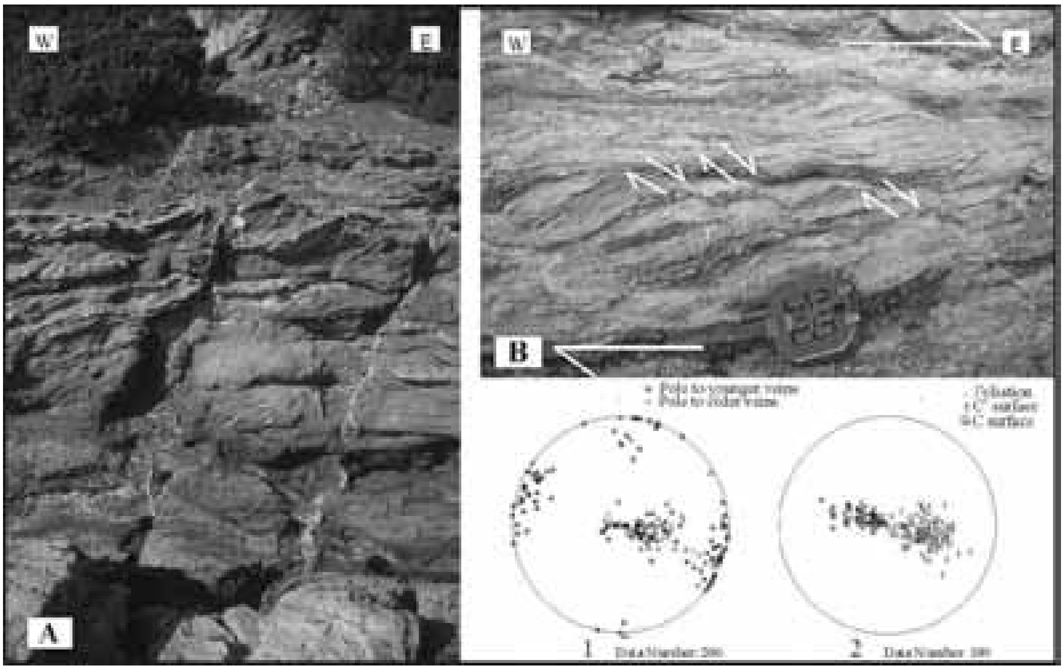

Figure 33 - A: Hydrofractures within the fault zone (stereoplot 1). B: Top-to-the-East C-type and C'-type shear bands and west-dipping foliation within the fault zone at Punta di Zuccale (stereoplot 2) .

stereoplot 2) that display consistent top-to-the-east senses of shear.

Microstructural studies have been used for the identification of a five-layer tectonic stratigraphy of the fault zone (from base to top): L1) cataclasite with basement clasts set in a carbonate-chlorite-quartz-matrix (figure 34A); L2-L3) highly foliated units of, respectively, green serpentine (chrysotile)-tremolite-talc-chlorite 'schist' (L2, figure 34B) and a highly heterogeneous unit of chlorite phyllonites (L3, figure 34 C) with lenses of carbonate mylonite; L4) carbonate vein-rich domain in cataclasite, incorporating sigmoidal pods and lenses of carbonate, calc-schists and ultramafic material; L5) foliated fault gouge and fault breccia. Individual fault rock units vary in thickness. Locally this occurs due to the presence of generally E-verging m-scale ductile folds in L2-L3 and local excision of fault rock units along low-angle extensional detachments, e.g. in the coastal outcrops immediately N of Punta di Zuccale.

The turbidites (Complex V) resting above the fault zone are almost undeformed. L1 is clearly derived from cataclasis of the immediately underlying basement (Complex I), whilst L2-5 appear to be derived from highly dismembered and modified parts of the intervening thrust nappes (Complexes II-IV) (e.g. figure 31B).

The preservation of highly strained foliated, apparently ductile fault rocks in L2 and L3 sandwiched between a footwall and hangingwall in which fault-related deformation is exclusively brittle is a striking feature of the ZF and requires explanation. The widespread preservation of cataclastic textures – often as early relict features overprinted by a foliation within the Zuccale fault rocks suggests that grain-scale deformation processes were initially frictional and brittle throughout most, or all of the fault (cf. figure 34A). Grain-scale brittle dilatancy would increase permeability, readily promoting the influx of CO_2-rich hydrous fluids into the fault zone soon after its initiation. These fluids would likely react with the fine-grained, crushed cataclasites, leading to alteration, reaction softening and the onset of fluid-assisted dissolution and precipitation processes such as pressure solution in the narrow, highly deformed core region of the fault. The observed localisation of strain into the foliated fault core suggests that these processes led to significant weakening of the fault zone.

Features to be discussed and studied:

1) Fault zone architecture

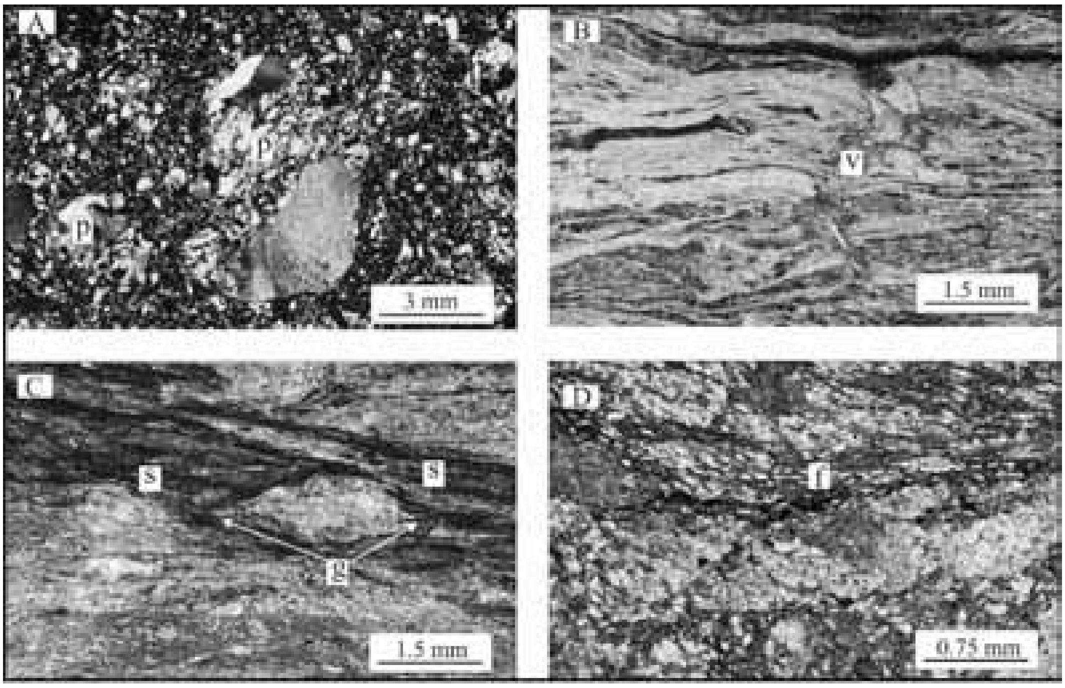

Figure 34 - Photomicrographs of fault rocks, from Punta di Zuccale. A) Cataclasite with basement clasts set in a carbonate-chlorite-quartz matrix from L1; (p) basement fabric in clasts thought to pre-date ZF. Crossed polar view. B) Highly foliated serpentine, tremolite, talc and chlorite rock derived from ultramafic protolith in L2. Partially sheared and folded carbonate vein (v) cross-cuts foliation in centre of view. Crossed polar view. C) Dark dissolution seams (s) and fibrous overgrowths (g) of tremolite, talc and calcite wrapping a carbonate clast in L3. Crossed polar view. D) Alteration and collapse of fine-grained cataclasite (L1-L2 boundary) showing smearing out of phyllosilicate-rich matrix to form foliation (f) in upper part of picture. Viewed in plane polarized light.

2) C-type and C'-type shear bands
3) Hydrofracture systems
4) Fault-zone weakening processes

Stop 23:

Spiagge Nere

From Punta di Zuccale we will drive to Spiagge Nere.

At Spiagge Nere the ZF separates a hangingwall sequence of different rock types pertaining to Complexes II and III from a footwall of Palaeozoic, muscovite-biotite, basement schists (Complex I).

Small displacement N-S trending, high angle normal faults crop out in the hanging-wall block: these faults sole into the ZF and show the same kinematics (figure 35). The fault zone is undulose (figures. 35 and 36) with E-W trending lineations (figure. 35, stereoplot 1), the fault rock is characterised by strongly foliated cataclasites with various protolith lithologies (carbonates and ultramafics). Within the fault zone the kinematic indicators are represented by C-type and C'-type shear bands bounding west-dipping S foliated domains (figure 35). The foliation surfaces are west-dipping but in some cases they show different orientations that are possibly due to bending phenomena.

The exposure in Spiagge Nere is spectacular for observing the deformation in the footwall block (figure 36). In this outcrop gently west-dipping segments of the ZF are associated in the footwall with a complex fault system (figure 36 stereoplot 1) that in the majority of the cases cut and displaces the aplitic dykes intruded during the emplacement of the Porto Azzurro granodiorite (see location in figure 31). Small-displacement listric normal faults show in their hanging-wall variably back-rotated arrays of normal faults.

Features to be discussed and studied:

1) Fault zone architecture
2) High-angle synthetic normal faults in the hanging-wall that sole into the ZF

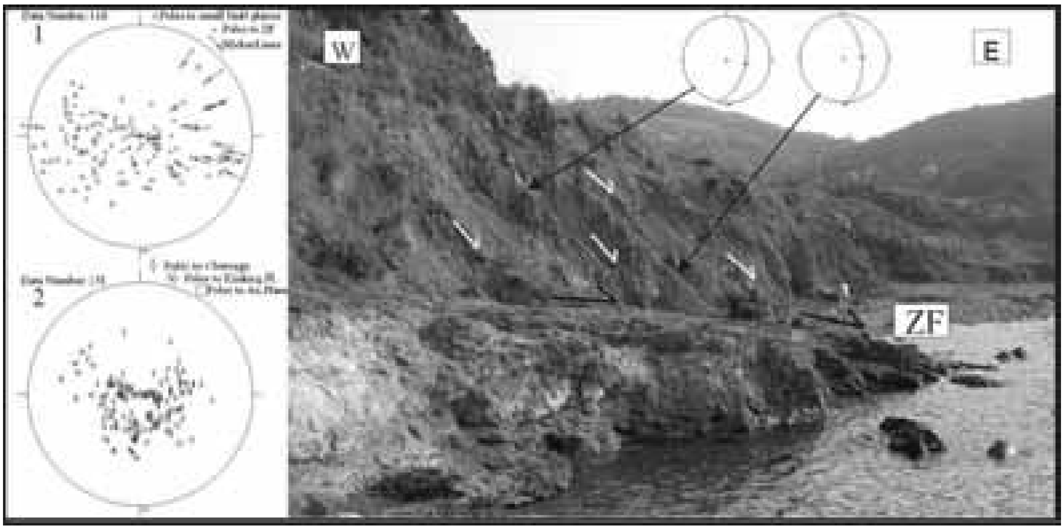

Figure 35 - A: Architecture of the fault zone at Spiagge Nere. High angle normal faults that sole into the ZF here represented by the black arrows. Fault zone structures equal area stereoplots 1 and 2 (lower hemisphere projection) of structural data, with planes plotted as poles.

Figure 36 - A: Back-rotated and listric low-angle normal faults within the basement (ZF footwall) at Spiagge Nere. Basement structures equal area: stereoplot 1 poles to fault planes, stereoplot 2 poles to extensional fractures and dykes (lower hemisphere projection).

3) Fault-zone weakening processes
4) Normal fault system in the ZF footwall

DAY 4

A trip across the Middle Tuscan Range and the Larderello Geothermal field.

(Domenico Liotta, Antonio Lazzarotto and Andrea Brogi)

Stop 24:

The mineralised Boccheggiano normal Fault (figures 36 and 37)

This stop is located in the Mining District of Boccheggiano. The outcrop is characterised by a regionally extended normal fault, less than 45° in dip, whose activity can be referred to Pliocene. Within the footwall, the Palaeozoic phyllites (“Filladi del Torrente Mersino” Fm.), belonging to the Monticiano-

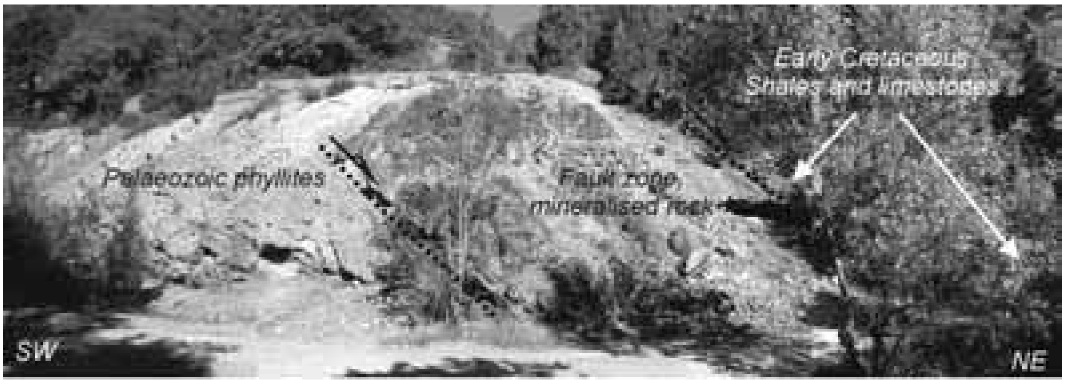

Figure 37 - Panoramic view of the Boccheggiano normal fault along the Mersino Stream.

Roccastrada Unit crop out; in the hanging wall, the Lower Cretaceous shales and limestones ("Argille a Palombini" Fm.), belonging to the Ophiolitic Unit, occur (figure 38).
The fault zone, about 10 metres thick, is characterised by a mineralised fault rock consisting of quartzite associated with sulphides, mainly represented by pyrite.
Sulphur, derived from alteration of sulphides also occur. Along the Boccheggiano fault, which can be mapped for more than 10 km, is located the most inportant Pyrite and mixed sulphides mines of the "Colline Metallifere". The last exploited mine (Campiano mine) was opened in the 1960s and definitively abandoned in 1993.

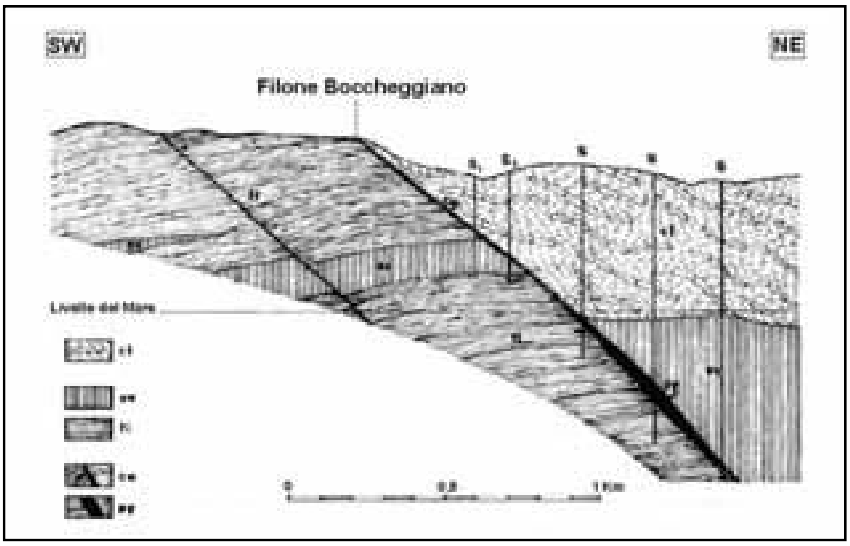

Figure 38 - Geological cross-section through the Boccheggiano fault area (after Arisi Rota et al., 1970). Key: cf: Ligurides; ev: evaporites; fi: phillites; cu:calcopyrite; Py: Pyrite.

Stop 25:

The Calcare Cavernoso Formation.

This stop is located along the Massa M.ma – Larderello road. The "Calcare Cavernoso" represents an important stratigraphical horizon deriving from the alteration of the Upper Triassic evaporites. Thus it belong to the "Formazione Anidritica di Burano" (Norian-Rhaetian), located at the bottom of the Tuscan Nappe. The Triassic evaporites represented the main detachment level during the collisional and post-collisional stages of the Northern Apennines. Tectonic activity and chemicals processes, also induced by hydration and dehydration of the anhydrites, transformed the Triassic evaporites (composed of dolostones and anhydrites) in a vacuolar carbonatic breccias, named "Calcare Cavernoso" Fm. The "Calcare Cavernoso" Fm., because of its brecciated texture (the estimated permeability is of 10^{-15} or 10^{-14}/m^2), is the upper geothermal reservoir both in the Larderello and Monte Amiata geothermal areas and represents the main mineralised (mixed sulphides) horizon occurring in the Colline Metallifere region.

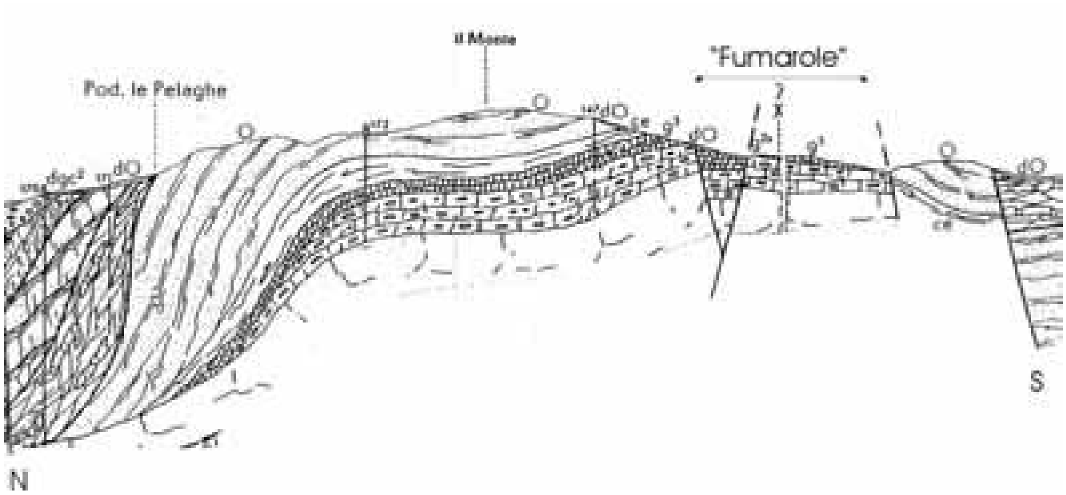

Figure 39 - Geological cross section through the Monterotondo M.mo area.The Macigno (O) and Scaglia Toscana (ce) fms represent the impervious cover of the carbonatic reservoir indicated by g3, g2 and the below formations

Figure 40 - Panoramic view of the fumarole in the "il Monte" area. The gas vents occur in correspondence of the calcareous reservoir.

Figure 41 - Panoramic view of the contact between the carbonatic reservoir and the impermeable shaly cover in the il Monterotondo M.mo area

Stop 26:

The "fumarole" of Monterotondo M.mo natural gas vent (figures 39, 40 and 41)

In this stop the first reservoir of the Larderello geothermal area crops out. It consists of fractured rocks related to the Upper Triassic, Jurassic and Cretaceous succession of the Tuscan Nappe outcropping in the southern side of the "Il Monte" hill. At the top of the hill, the "Scaglia Toscana" and "Macigno" Fms. form the impermeable cover of this reservoir.

The reservoir lithotypes, belonging to the Jurassic succession, are intensely altered by gas circulation. The gas emanation characterises only the calcareous reservoir and interrupts at the contact with the impervious cover.

Stop 27:

The Larderello Geothermal Field

(figure 42 and 43)

The Larderello geothermal field is the most ancient exploited geothermal field in the world. Particularly, the Larderello field has been industrially exploited since 1818, when the Count of Montecerboli, Francesco de Larderel, extracted boric acid from the geothermal vents. In 1904 the Larderello geothermal fluids were used to produce electricity by Prince Piero Ginori Conti. In contrast, the Mt. Amiata area has been exploited since the early 1960s, when the first electrical power plant was activated.

Today the endogenous fluids, intercepted in depth by

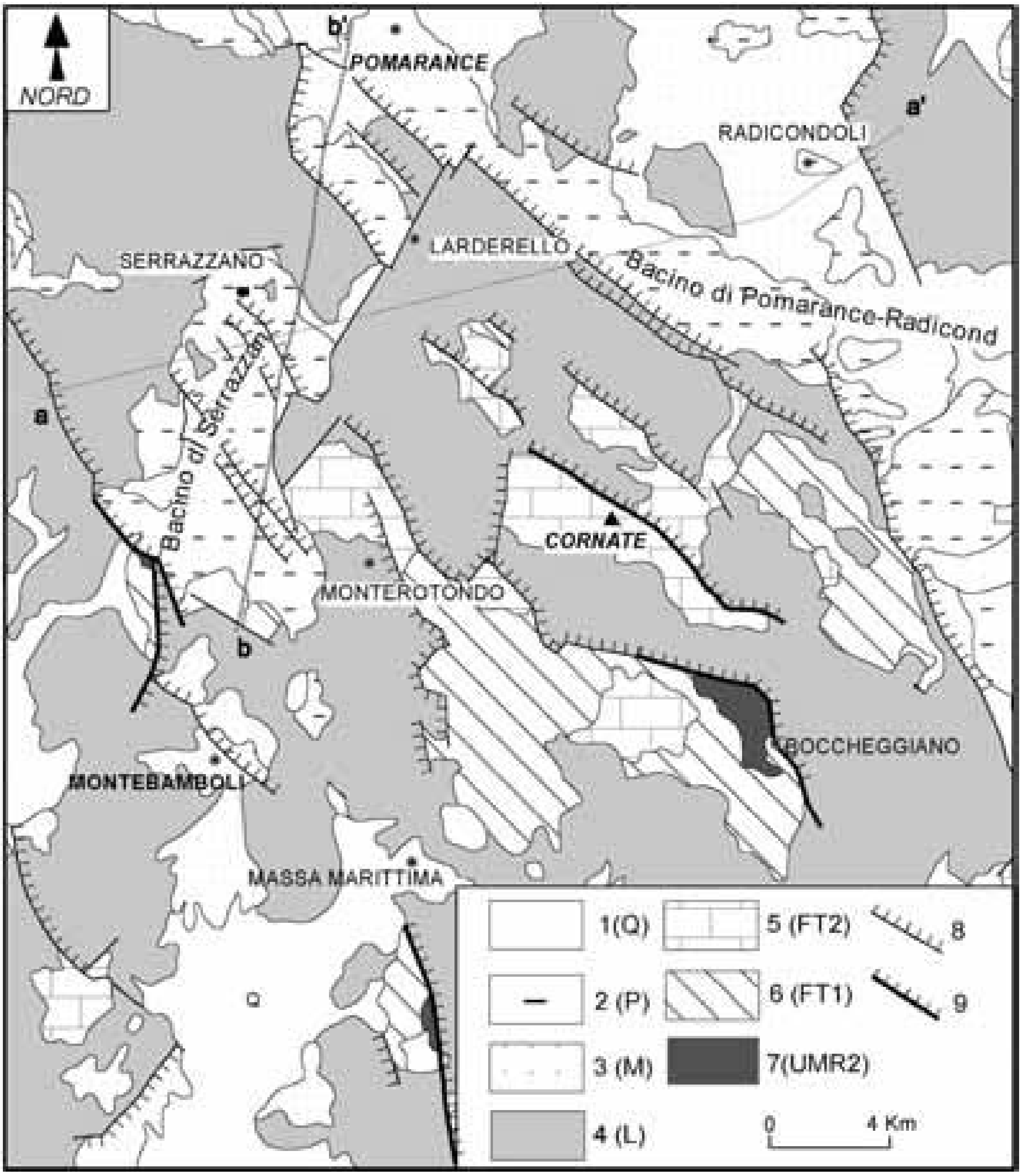

Figure 42 - Structural sketch of the Larderello area. Key: 1: Quaternary continental deposits; 2: Pliocene marine succession; 3: Late Miocene marine succession; 4: Ligurian Complex (Jurassic -Oligocene); 5: -Tuscan Complex (Late Trias -Early Miocene sedimentary sequence); 6: Reservoir: Tuscan Complex (Late Triassic evaporite); 7: Palaeozoic phyllites; 8: normal faults; : normal faults mineralized by mixed sulphides of hydrothermal origin, linked to the Pliocene-Quaternary Tuscan magmatism (Arisi Rota et al., 1971). b-b' : trace of section shown in fig 43.

boreholes, feed the Larderello and the Mt. Amiata power plants belonging to the ENEL Greenpower Electric Company. Present production is more than 700 MW, corresponding to about 2% of the total electricity production in Italy.

The Larderello system is made up of: a) an impervious cover, constituted by the Miocene and Pliocene sediments, by the Ligurian Units and by

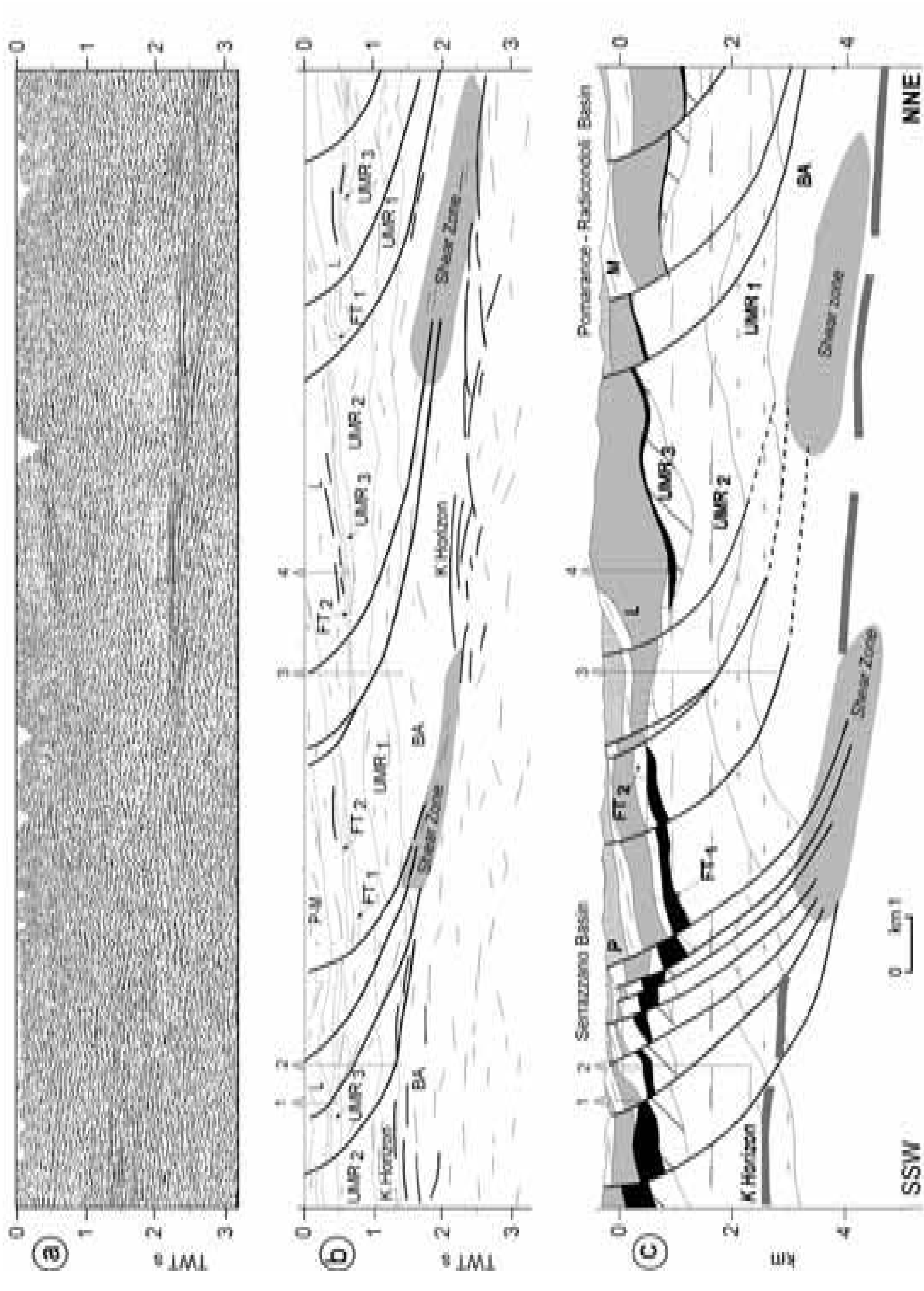
a
TWT s
b
c
km
SSW
NNE
K Horizon
Serrazzano Basin
Pomarance - Radicondoli Basin
Shear Zone
UMR 1
UMR 2
UMR 3
FT 1
FT 2
FT 4
BA
P-M
P
L
M
0 1 km

Figure 43 - A. Geological cross section through the Larderello geothermal field (constructed from borehole stratigraphy, seismic reflection lines and field mapping). Location indicated in Fig. 42. Pliocene normal faults cross cut all earlier structures: P- Pliocene sediments; M - Late Miocene sediments; L - Ligurian Complex; Tuscan Nappe (TN), TN2 - carbonatic and terrigeneous sequence (Jurassic-Late Oligocene-?Early Miocene); TN1 - Late Triassic evaporites; Monticiano Roccastrada Unit (MRU), MRU3 - Mesozoic-Paleozoic Group; MRU2 - phyllite-quartzite Group; MRU1 - Hercynian Micaschist Group; GN - Gneiss Complex. B. Reconstructed Pliocene geological section. Note that Late Miocene sediments are preserved in tectonic depressions defined by synsedimentary normal faulting related to the second extensional event; Early and Middle Pliocene sediments rest unconformably on Late Miocene sediments which were deformed during their deposition.

the Cretaceous-Miocene sequence of the Tuscan Nappe; b) two reservoirs placed at different depths: the shallower, located in the Upper Triassic evaporite level and in the brecciated Jurassic limenstone, at the base of the Tuscan Nappe (about 600 m in depth); the deeper, located in fractured horizons of the Metamorphic Basement and ranging in depth between 2000 and 3500 m; c) a recharge area located in the outcropping Jurassic and Cretaceous calcareous rocks belonging to the Tuscan Nappe.

The impervious cover

The Miocene and Pliocene deposits

These unconformably overlie previously deformed pre-Neogene rocks and fill the extensional tectonic depressions surrounding the Larderello area. The sedimentary sequence consists of Upper Tortonian-Messinian lacustrine, marine and brackish deposits and two different sequences of marine sediments (Lower and Middle Pliocene).

The Ligurides l.s.

These units are, from top to bottom :

The Ligurian Units

-*Ophiolitic Unit*, made up of Dogger-Malm ophiolites, Malm pelagic sequence (Radiolarites), and Lower Cretaceous pelagic deposits (*Calpionelle* Limestones and "Argille a Palombini")

-*Ophiolite-bearing Monteverdi M.mo-Lanciaia Unit*, entailing the following formations: *Lanciaia Formation*, made up of an ophiolite complex and an arenaceous flysch, with breccias and ophiolite bearing sandstones intercalated (Lower - Middle Eocene); *Monteverdi M.mo Fm.*, includes a clayey-calcareous complex of Cenomanian-Turonian (Poggio Rocchino Formation), and a marly-calcareous flysch of Upper Cretaceous - Lower Paleocene;

-*S. Fiora Unit*, in the Larderello area it is represented by a marly-arenaceous flysch (S. Fiora Fm.), with levels of Upper Cretaceous - Lower Paleocene quartz-calcareous sandstone (Pietraforte), and by a marly-calcareous flysch of Paleocene-Middle Eocene;

The Subligurian Unit

This is represented by: *"Argille e Calcari" Unit* made up of a clayey-calcarenitic flysch of Paleocene-Eocene (Canetolo Fm.);

The Tuscan Nappe Lower Miocene - Jurassic sequence

The Tuscan Nappe is made up of a sedimentary sequence from Upper Triassic to Lower Miocene formations whose sedimentary evolution reflects the transition from evaporitic (Burano anhydrites Formation, Upper Triassic) to carbonatic shelf environment (Rhaetic limestones with *Rhaetavicula Contorta*, and Lower-Lias limestones), to a progressively deeper pelagic basin (Middle-Upper Liassic chert Limestone, *Posidonomya* Marls of Dogger, Malm Jaspers, Lower Cretaceous - Oligocene calcareous marls named "*Scaglia Toscana"*) to, finally, a subsiding foreland basin where an arenaceous flysch, ("Macigno" Fm.) deposited during the Late Oligocene-Early Miocene.

The reservoirs

The first reservoir is mostly located in the cataclastic level corresponding to the Upper Triassic evaporites. Its depth ranges between 500 and 1000 m, roughly. This level was the main detachment surface during the collisional stage, favouring the stacking of the Tuscan Nappe on the eastern domains, and during the first event of the extensional tectonics. A permeability of about 10^{-15} or 10^{-14} m^2 are estimated for this reservoir. The bulk of permeability is attributed to tectonic fractures. However, the dissolution of the Upper Triassic anhydrite in the presence of hot saline solutions, with the consequence collapse and ropture of the overlying carbonate rocks can contribute to increased permeability. The second and deeper reservoir is located in fractured horizons within the Tuscan Metamorphic Basement.

The recharge area

The main recharge area is located in the Cornate area where fractured limestones and cherty limestones belonging to the Tuscan Nappe largely crop out. Other recharge areas are located at the borders of the Pliocene sedimentary basins where the exposed substratum Nappe largely crop out. Other recharge areas are located at the borders of the Pliocene sedimentary basins where the exposed substratum is affected by the cataclasis due to the normal faults delimiting the Pliocene-Quaternary tectonic depressions.

DAY 5

The Altotiberina Fault and the Plio-quaternary extensional strucures in Umbria

Francesco Brozzetti

In recent times, several authors (Brozzetti, 1995, Barchi et al., Boncio et al., 1998, 2000, Collettini and Barchi, 2002) highlighted an important low-angle normal fault (Alto Tiberina fault= AF) extending, for at least 70 km, in a NNW-SSE direction along western Umbria, from Sansepolcro to the south of Perugia and emerging some km to the west of the eastern Tiber basin (figure 44).

It is shown by field evidence (*Brozzetti*, 1995) and imaged by the CROP 03 near vertical reflection (NVR) seismic profile plus a number of commercial seismic lines (*Barchi et al.*, 1998; *Boncio et al.*, 1998). An average dip of 30° (with flats dipping 10° to 15° and ramps up to 45°) and a maximum depth of 12-14 km, reached by the AF beneath the Umbria-Marche Apennine, have been evaluated, mainly on the base of subsurface data.

To better understand the geology and kinematics of the AF, it is worth making a brief analysis of the deformation recorded in its hanging wall. On the basis of the structural style the AF hanging wall may be subdivided into three main structural blocks. From west to east they are (1) an inner zone, here named Perugia Massifs extensional allochthon (PMA) (in the sense of *Wernicke* (1985), where synthetic east-dipping normal faults prevail; (2) a central zone, where an asymmetric tectonic depression, the eastern Tiber basin, filled with fluvial and lacustrine deposits is developed; and (3) an outer zone, where high-angle west-dipping normal faults prevail, mainly distributed along the NNW-SSE Sansepolcro – Gubbio - Colfiorito alignment. These structural blocks will be analyzed in this last part of the field trip along a WSW-ENE transect which extends from the Perugia Massifs area to the Gubbio fault.

The line drawings of four seismic lines across the AF are reported in this paper (figure 45). They show an evident AF seismic image suggesting the presence of a thick, mature shear zone which slopes gradually eastward to a depth of 12-14 km (5 s two-way time (TWT)) beneath the Apennine mountain chain. The AF plane is well detectable also because a number of rather continuous seismic markers within the AF hanging wall abruptly stop on it. The seismic lines

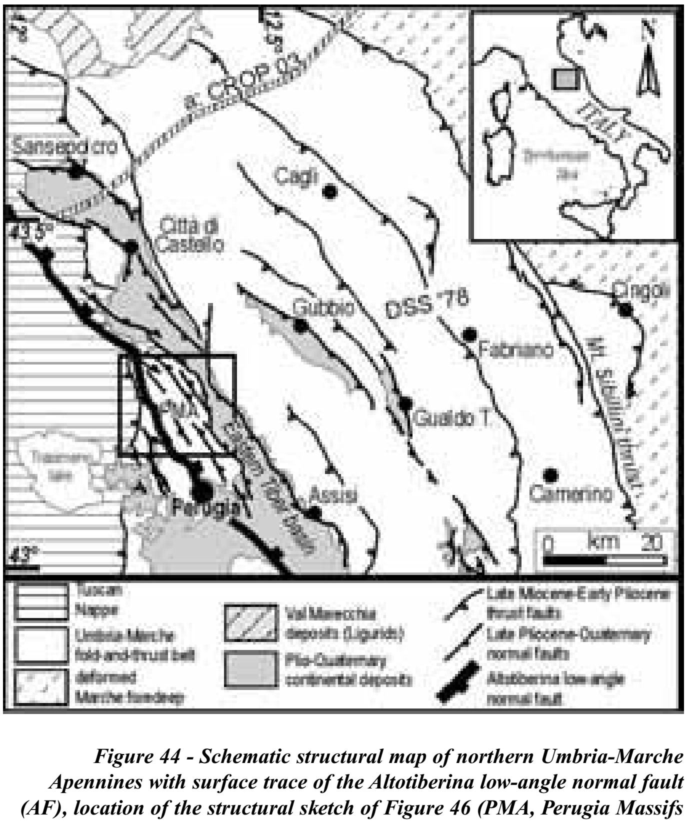

Figure 44 - Schematic structural map of northern Umbria-Marche Apennines with surface trace of the Altotiberina low-angle normal fault (AF), location of the structural sketch of Figure 46 (PMA, Perugia Massifs extensional allochthon) and traces of the CROP 03 deep-crust near vertical reflection (NVR) seismic profile.

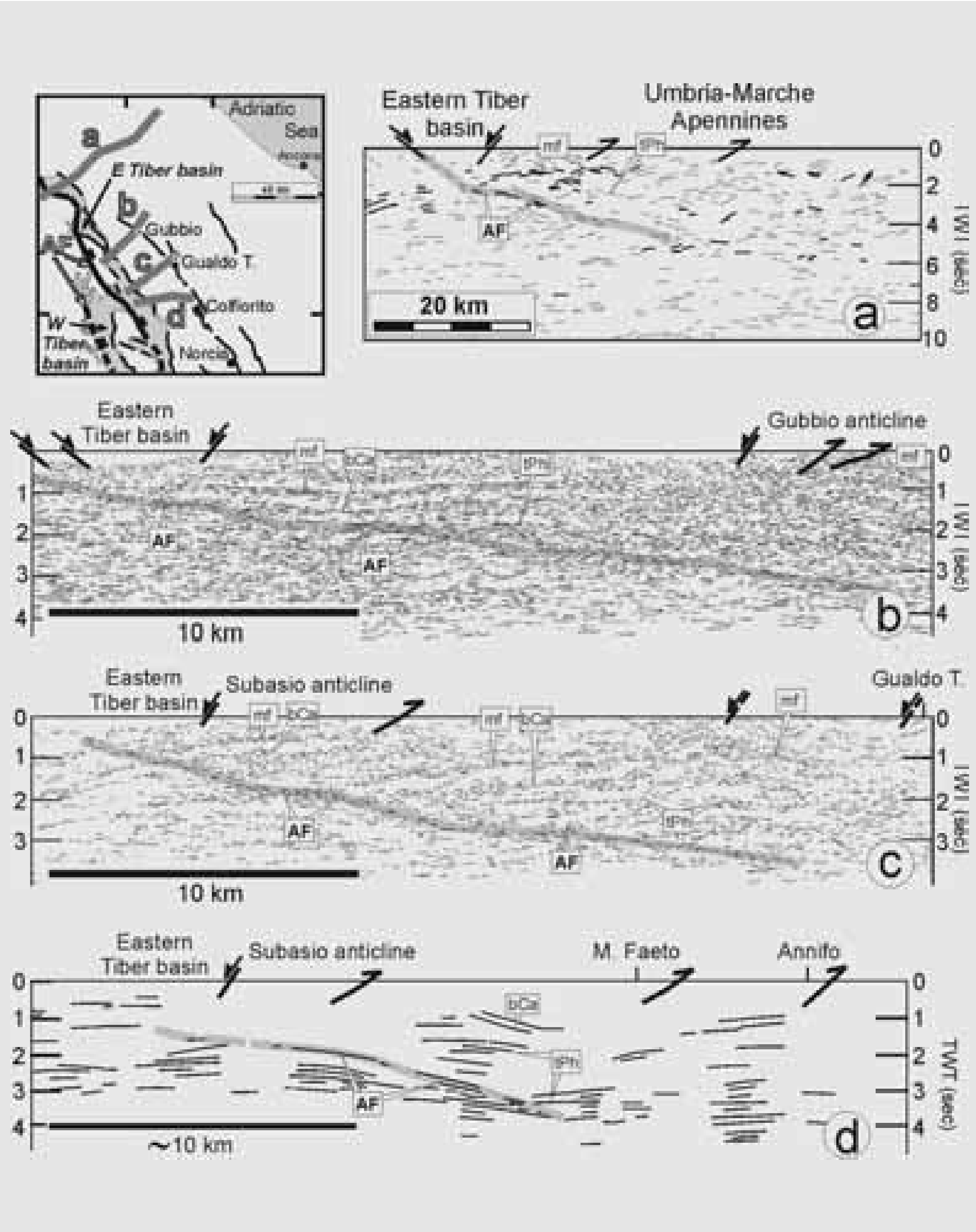

Figure 45 - Line drawings of NVR seismic lines showing the AF deep geometry. (a) CROP 03 deep-crust line [after Decandia et al., 1998], (b and c) eastern Tiber basin-Gubbio and Subasio-Gualdo Tadino lines [after Boncio et al., 1998], and (d) Subasio-Annifo line [after Bally et al., 1986]. AF, Altotiberina normal fault; mf, Marne a Fucoidi fm of the Umbria-Marche carbonatic sequence; bCa, bottom of carbonatic sequence; tPh, top of phyllitic basement. Location and geometry of thrust planes and normal faults are taken from surface geology. Note the sharp westward termination of the stratigraphic markers on the AF plane.

show west dipping high-angle normal faults antithetic to the AF as well. Although the seismic expression of the antithetic faults is not as good as the AF one, it is still evident that the AF trace is not offset at their intersection.

The normal offset along the AF plane, computed on these sections, can be evaluated altogether in 4-5 km.

Stop 28 : *(looking towards SE)*

Pantano village: The AF Breakaway zone

North of Perugia, the AF breakaway zone is clearly exposed along the eastern side of the M. Malbe-M. Torrazzo ridge where Lower Miocene turbidites, are justaposed on Triassic evaporites, with a local displacement of 1000 m causing the complete omission of the Meso-Cenozoic carbonate sequence. Looking in the S-SE direction a panoramic view of the AF breakaway zone along the M. Malbe - M. Tezio transect can be observed: on the west side of the view (right) the wooded hill zone of M. Malbe (mean elevation 600 m a.s.l.) consists of an exhumed and uplifted Triassic core at the footwall of the AF; on the east side the M. Tezio ridge (in the foreground) is a fault-bounded block whose structure can be described as a SW dipping monocline.

Stop 28 : *(looking towards NW) - Pantano* Village:

The Perugia Massifs extensional allochthon (PMA).

The PMA consists of east-dipping normal and normal-oblique faults which dissect and severely stretch a previously folded and thrusted carbonate and siliciclastic Meso-Cenozoic multilayer (Figure 46). The faults, spaced 500 m apart, originated a domino-like structure characterized by fault-bounded blocks having a progressive eastward lowering of the crest structural elevation (*Brozzetti*, 1995). Bedding inside the blocks constantly dip toward SW, with 20° to 45° dip angle. Taking in account the general observation that in a domino-like structure the ratio between the depth of the detachment and the fault spacing would not exceed 2.5 (*Le Pichon and Sibuet*, 1981; *Davison*, 1989), it may be predicted that the PMA structures detach onto an east dipping gently inclined horizon, ranging in depth from 1 to 4 km.

This prediction is supported by seismic lines showing that the PMA faults sole at an average depth of 2 km on a very low-angle (10°-20°dip) segment of the AF (see section a-a' in figure 46).

The stratigraphy of the S. Donato well that penetrates the AF 12 km W of the Tiber basin (location in figure 46) confirms the anomalous contact of Miocene Marnoso Arenacea over Upper Triassic anhydrides due to the AF displacement.

In the PMA, two coaxial sets of east dipping normal faults detaching on the AF may be recognized: a low-angle set (average 30° dip) crosscut by a high-angle one (45°-60° dip). This fault pattern has been interpreted in terms of progressive rotational extension of the AF hanging wall rocks (*Brozzetti*, 1995).

The amount of extension computed for the PMA, through balancing and restoring the geological cross section in figure 46 is 4 km, a value exceeding 40% of the preextensional length. Such a value is coherent with the total amount of displacement (4 -5 km), which has been estimated along the AF detachment plane by the interpretation of seismic profiles (*Barchi et al*., 1998a).

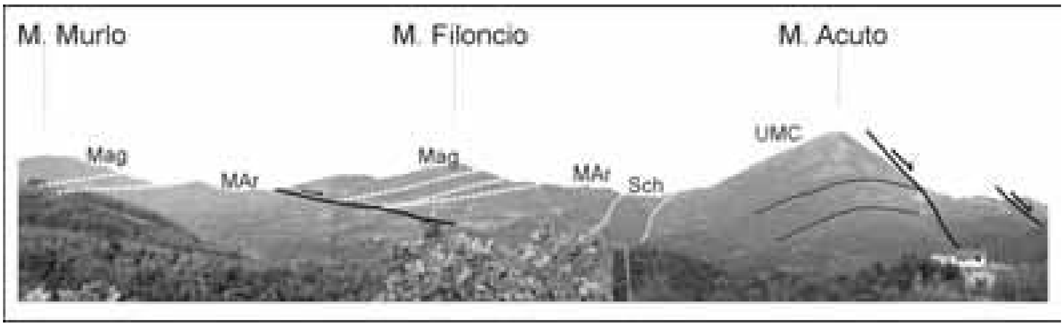

Figure 47 - Panoramic view of the northern side of the Nese Valley along the transect M. Murlo – M.. Acuto. (approximately 2 km north of the the trace of the section in figure 48). Mag: Macigno sandstones and underlying varicoloured shales; Mar: Marnoso Arenacea fm; Sch: Schlier fm; UMC: Umbria-Marche meso-cenozoic carbonate succession; white dotted lines: thrust faults; white heavy lines: stratigraphic boundaries; black heavy lines: normal faults.

In Figure 47 a panoramic view of the northern side of the Nese Valley and a schematic line-drawing, showing the main outcropping geological structures, is shown. This WSW-ENE transect crosses the AF breakaway zone and the PMA along a direction which is very close to the b-b' section in Figure 46. Three main features corresponding to the more evident morphological high can be observed. From W to E they are: a) the Tuscan allochthon leading edge along the eastern side of the M. Murlo ridge; b) The M. Filoncio lowered block, in which the Tuscan succession (Argille Varicolori+Macigno fms) lies upon the Marnoso Arenacea fm along an east dipping low angle plane; c) The M. Acuto ridge, a SW-dipping monocline made of the Umbria

carbonatic succession bounded eastward by an high angle east dipping normal fault.

A detailed geological recostruction of the Nese Valley geology, interpreted at depth to some hundreds meters below the sea level, is shown in the geological section

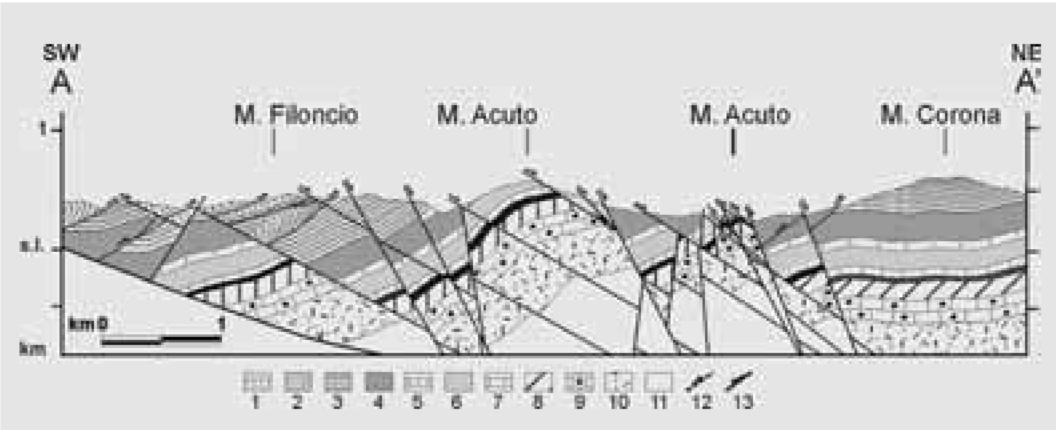

Figure 48 - Detailed geological section across the S. Donato well – M. Acuto transect (trace in Figure46).
1: Macigno fm (late Oligocene); 2: Argille Varicolori fm (early-late Oligocene); 3: Marnoso Arenacea fm (middle Burdigalian – early Langhian); 4: Schlier fm (middle Burdigalian); 5: Bisciaro and Scaglia Cinerea fms (Oligocene – early Burdigalian); 6: Scaglia Variegata and Scaglia Rossa fms (Turonian - late Eocene);
7: Scaglia Bianca fm (Cenomanian – Turonian); Marne a Fucoidi fm (Aptian-Albian);
8: Maiolica fm (late Tithonian – Barremian); 9: Corniola to Calcari Diasprini succession (middle Lias – early Tithonian); 10: Calcare Massiccio fm (early Lias);
11: Burano Anhidrites and Raethavicula contorta fm (late Triassic); 12 thrust fault; 13: normal fault.

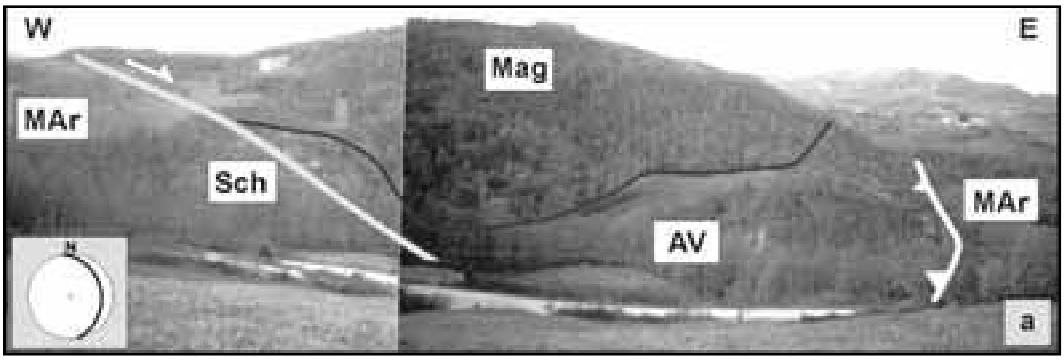

Figure 49 - Fosso Feriano Valley. View of a low-angle normal fault in the breakaway zone of the AF and stereographic projection of the fault plane (location in Figure 46). Mag, Macigno fm of the Tuscan Nappe (late Oligocene-early Miocene); AV, Argille Varicolori fm of the Tuscan Nappe (late Eocene-early Oligocene); Sch, Schlier fm (early Miocene); Mar, Marnoso-Arenacea fm (early Miocene).

of figure 48.

Stop 29:

Nese Valley: The Fosso Feriano-Rio dell'Acqua low-angle splay of the AF

In the north side of the Feriano Stream, the AF westernmost low-angle splay causes the contact between the western Umbria Miocenic turbidite succession and the overlying Tuscan Falterona Nappe (figure 49). The emplacement of the Tuscan allochthon occurred, in this part of the Apennines, during the Late Burdigalian.

The Feriano Stream Fault, mapped for about 8 km, shows a mean 25° dip-angle on a N60/70 dip direction. Another good exposure of this same fault plane can be observed 2 km northward in the Rio dell'Acqua stream (Figure 50) .

Another low-angle east dipping normal fault has been recognised, in spite of the low along-strike continuity, on the east side of the M. Acuto – M. Gudiolo ridge and in the Galera quarry (see also description of Stop 31)

Stop 30:

Migiana: The M. Tezio-M. Acuto boundary fault.

In the eastern part of the MPA, the east-dipping, low-angle normal faults are cut and displaced by high-angle sinthetic ones. The high-angle east-dipping boundary faults are the main extensional structures influencing the actual morphology of the area. Among these, the M. Tezio-M. Acuto and the M. Mussarello – M. Elceto faults are the most evident reaching offsets up to 1000m.

The M. Tezio-M. Acuto boundary fault, in particular, crops out for about 10 km lowering the Miocene Schlier and Marnoso Arenacea fms (within the Hanging wall block) on the Liassic Calcare Massiccio and Corniola fms (in the footwall block) (figure51). Its mean direction is N140E but sharp variations to NS or to N100 "oblique" attitude are frequent; oblique transfert segments do not exceed, in general, 500 m in length. The dip angle of the main fault plane varies from 35° to 60°; a 45° value is the most frequently surveyed. A number of lens-shaped extensional horses, consisting of strongly deformed cretaceous limestones bodies (Scaglia fm), align along the fault. Their size varies from some tens to some hundreds of meters along the strike. Dip slip movements prevail on NW-SE (N140°E-N160°E) oriented segments; normal oblique movements with right-lateral or left-lateral components are observed on N100°E or NS segments. The related stress tensor is triaxial-extensional, with a N60°E nearly horizontal $\upuparrows_3$ axis and a nearly vertical $\upuparrows_1$ axis.

Stop 31:

Low-angle east-dipping normal fault and rotated antithetic fault in the Galera Quarry.

In the Galera Quarry a SW dipping faulted monocline made by the meso-Cenozoic carbonates is exposed (Figure 52).

The oldest stratigraphic unit, the Marne a Fucoidi (Aptian-Albian) crops out on the right side of the section and passes upward to the Cenomanian - Turonian Scaglia Bianca, containing the Bonarelli bed (black layer), and to the Upper Cretaceous Scaglia Rossa. Two normal fault systems can be observed, consistent and opposite to the dip of bedding. The opposite system appears with a clear east-dipping low-angle plane displacing the Oligocene-Miocene successione (Scaglia Cinerea and Bisciaro fms) above the various members of the Scaglia fm. In the central part of the quarry a subvertical fault, coaxial to the low-angle one, lowers the Bisciaro beds towards SW. The entire extensional structure can be interpreted as a conjugate normal fault

AF Kinematics

The fault pattern described at the previous system, rotated above a NE dipping shallow detachment, identifiable with the AF plane points, considered to be representative of the entire PMA, has been interpreted in terms of progressive rotational extension of the AF hanging wall rocks (*Brozzetti*, 1995). As a consequence of the progressive extension, the first-generation fault set, originally dipping 45°-60°, would have rotated 20° counterclockwise.

Contemporaneously, the AF basal detachment, originally dipping 30°, gradually decreased in dip in response to the flexure and uplift of the footwall induced by hanging wall unloading.

The second-generation set of high-angle faults, produced by further extension, would have not undergone any significant rotation. The nucleation of the high-angle extensional splays and the related unloading would have migrated from west to east, and with the same polarity, the main fault plane would have gradually flattened consistently with the concept of a "rolling-hinge" model (e.g. *Buck*, 1988; *Wernicke and Axen*, 1988) .

Syntectonic extensional deposits are missing almost

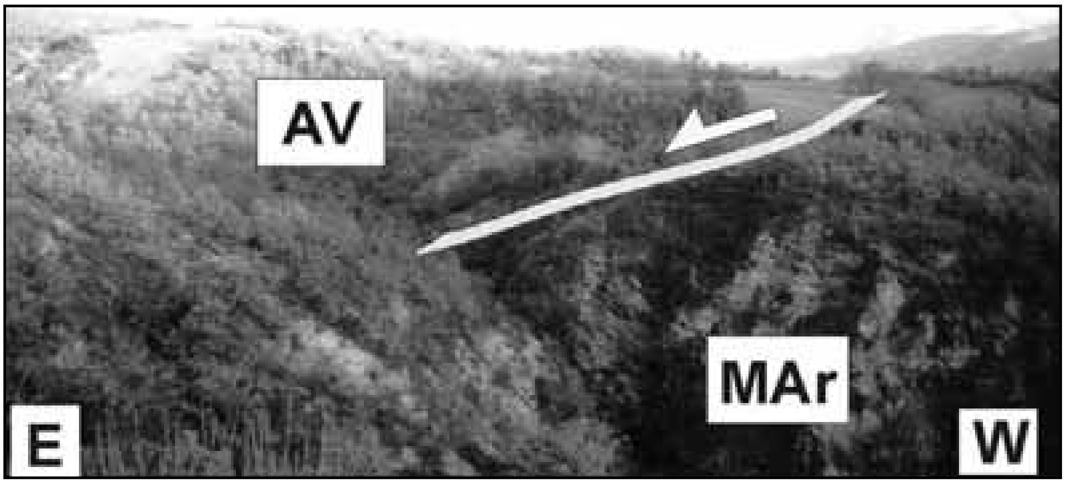

Figure 50 - Exposure of the previous fault plane along the Rio dell'Acqua Valley (2 km north of Fosso feriano. AV, Argille Varicolori fm of the Tuscan Nappe (late Eocene - early Oligocene); Mar, Marnoso-Arenacea fm (early Miocene).

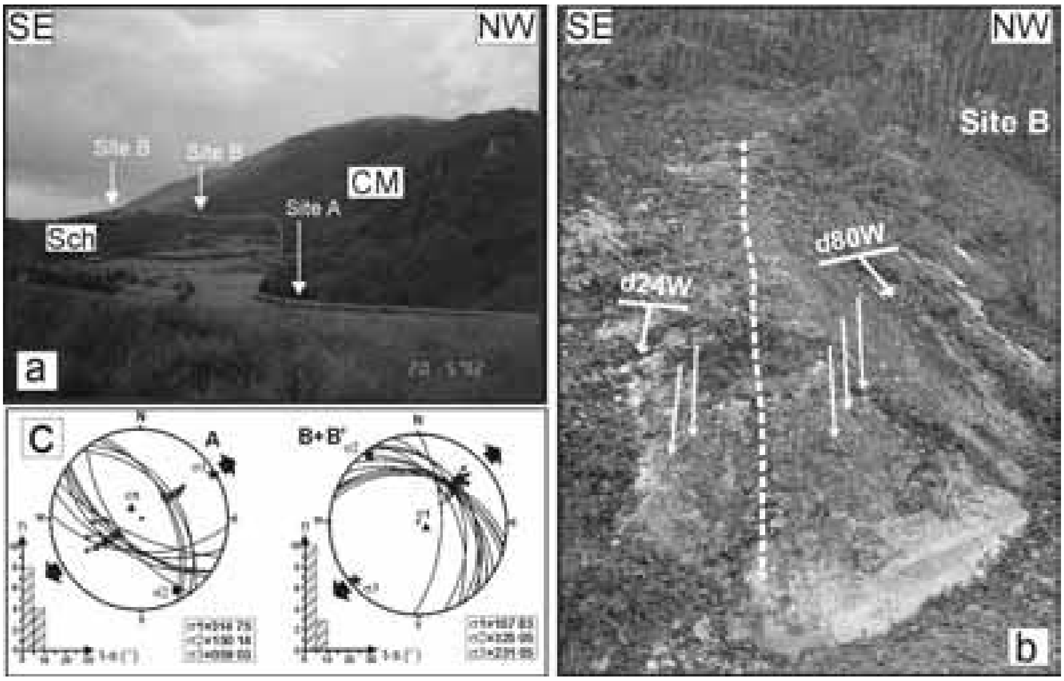

Figure 51 - a) panorama of the M.Tezio – M. Acuto boundary fault looking southward from the M. Mussarello ridge; CM Calcare Massiccio fm (early Lias) Sch: Schlier fm (middle Miocene) (b), detail of the fault plane in the site B, (stop point 3) showing a sharp attitude change but constant displacement vectors (NE dipping) c) kinematic and tensorial analysis on structural survey sites (location in figure 46) performed by the Carey's (1976) inversion method (see text for details)

completely in the PMA.

This may be related to an intense uplift which, from the lower Pleistocene onward, affected the area owing to both local (hanging-wall unloading) and regional processes (*Dramis*, 1999). These uplifts, on one hand, did not favour depositional processes; on the other hand, they led to intense erosion, which produced a removal of eventual syntectonic deposits.

Stop 32:

Valenzino Castle: The M. Mussarello – M. Elceto boundary fault.

Moving eastward in NE direction the itinerary crosses the eastern part of the PMA. In spite of the less marked topography, also in this part of the structure, the domino-like geometry, characterised by SW dipping monoclines bounded by an east dipping rotational normal fault system, can be observed. Near the Valenzino Castle it is possible to see a panorama of the M. Elceto - M. Mussarello east-dipping normal fault. Along the road, two of its splays causing respectively the displacement of the Cretaceous Scaglia Rossa on the Liassic calcare Massicio and of the Miocene schlier fm on The Scaglia Rossa can be observed.

Stop 33:

Castiglione Ugolino: The eastern Tiber Basin

The eastern Tiber basin is a tectonic depression ≈6 km wide, filled with lacustrine and fluvial deposits. In most of the basin the oldest outcropping sediments are dated to Lower Pleistocene. The basal sequence could reasonably be older (probably the end of late Pliocene). Indeed, at the southeastern border of the basin (near Spoleto) and in the Todi-Terni area (50-70 km south of Perugia), mammal assemblages (Triversa faunas) and palynological determinations allow us to date clayey lacustrine deposits of the lower part of the basal sequence to Upper Pliocene (*Esu and Girotti*, 1991, and references therein).

The basin is clearly asymmetric, bounded westward by major high-angle (average 60°) east-dipping normal faults and eastward by minor antithetic west-dipping faults, both splaying from the AF. A wedge-shaped geometry with westward thickening of the basin clastic fill is well shown by seismic images (Fig.2a). Strong along-strike variations of sediment thickness are also frequent (*Barchi et al.*, 1998a, 1999). The northern part of the basin, from the CROP 03 profile to Città di Castello, consists of ≈500 m of continental deposits. South of Città di Castello, the thickness increases to ≈1000 m (≈1 s TWT) and reaches its maximum value (≈1300 m) north of Umbertide. These along-strike variations may be due to differential subsidence ratios related to the extensional reactivation of preexisting oblique faults confined within the AF hanging-wall block.

Along the western shoulder of the basin, remnants of the synextensional deposits reach an elevation of 650 m above sea level, that is, ≈200 m above the mean elevation observed at the eastern border. This complies well with a progressive counterclockwise tilting and dip decrease of the upper AF segment in response to the PMA tectonic unloading The present configuration of the Tiber basin, with respect to that of the adjacent PMA, suggests that at any given time, only the easternmost high-angle synthetic splay of the AF would be active and induce significant subsidence; the westernmost and more deformed part of the hanging wall would be rapidly uplifted

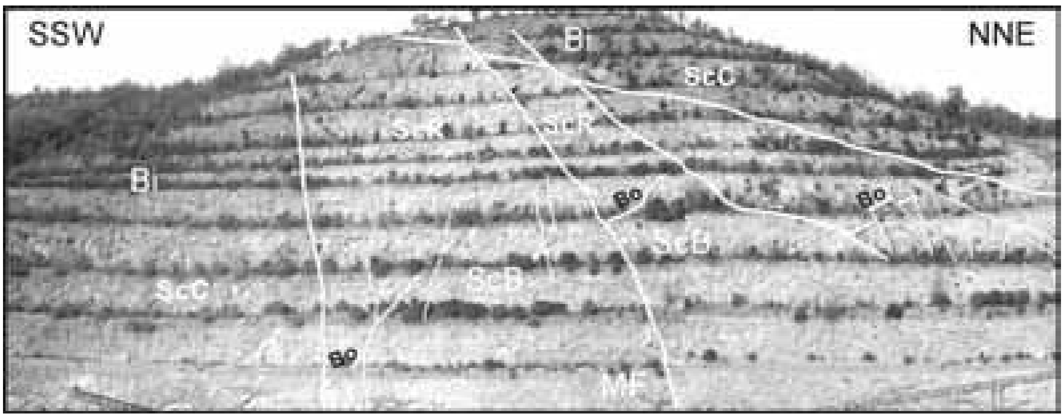

Figure 52 - View of the Galera quarry (see text for explanation); SCB: Scaglia Bianca fm (Cenomanian-Turonian p.p.); Bo: Bonarelli key bed (Turonian); ScR: Scaglia Rossa fm (late Turonian - early Eocene); ScC: Scaglia Cinerea (Oligocene – Aquitanian); Bi: Bisciaro fm (late Aquitanian – early Burdigalian)

and eroded after an initial stage of subsidence and sedimentation.

Stop 34:

Mengara: The Gubbio basin and the Gubbio normal fault *(figure 53)*

The Gubbio basin is a 20 km long, 4 km wide half graben filled with Lower (?)-Middle Pleistocene continental deposits. It extends in the NW-SE direction and originated from the subsidence induced by the Gubbio normal fault.

This latter bounds the basin to the east and dissects the inner limb of an east-verging anticline made by the Meso-cenozoic umbria marche succession. The Gubbio anticline generated by a contractional tectonic phase that affected the western part of the Umbria-Marche domain during the Tortonian-Messinian interval.

The Gubbio Normal fault can be surveied for at least 25 km, from Mocaiana to Pieve Compresseto. Its trace is particularly well exposed in the segment extending just above the old Gubbio town to the city cemetery (site of the following stop 8).

Seismic data and the stratigraphy of some water-research holes (Menichetti, 1992) suggest an half graben geometry with a marked depocenter located very close to the boundary fault.

The clastic fill consists mainly of lacustrine clays with minor lens-shaped fluvial sands and sandy gravels. Along the east border of the basin, the latter alternates with thick calcareous-clastic slope-debris supplied by the steep, fault-controlled side of the mountain ridge.

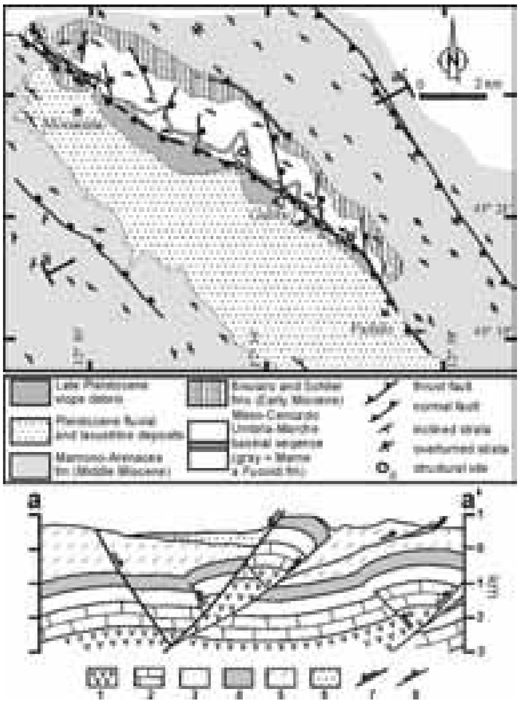

Figure 53 - Structural sketch of the Gubbio area (modified after Menichetti, 1992) and geological cross section through the Gubbio anticline.
1: Triassic evaporites; 2: Calcare Massiccio fm (Lower Lias); 3: Corniola to Maiolica fms (Middle Lias – Lower Cretaceous; 4: Marne a Fucoidi to Scaglia fms (Aptian-Oligocene); 5: Bisciaro to Marnoso Arenacea fms (Miocene); 6:Pleistocene continental deposits; 7: normal fault; 8: thrust fault.

Stop 35:

The Gubbio fault zone.

The Gubbio normal fault zone bounds eastward the Gubbio basin. It dips to the southwest 50°-70° and crosscuts a NE verging anticline with a vertical throw of >1000 m (*Menichetti and Pialli*, 1986) (figure 53 and 54). The available seismic lines show that at a depth of ≈6 km the Gubbio fault merges with the AF but does not penetrate and/or dislocate the basal detachment (figure 45).

A 20 km long and 6 km wide basin, filled with Pleistocene fluvio-lacustrine deposits (*Esu and Girotti*, 1991), is related to the fault. The fault plane is often covered by slope debris which probably accumulated during the Late Pleistocene cold climatic conditions. Locally, the Pleistocene fluvial and lacustrine sediments within the basin as well as the slope debris on the fault plane are tectonically deformed, suggesting a Quaternary activity of the Gubbio normal fault (*Boncio et al.*, 1996). Certainly, the fault is presently active, as indicated by the related seismicity (*Haessler et al.*, 1988; *Menichetti and Minelli*, 1991).

Minor and major normal and normal-oblique striated faults have been measured in survey sites along the Gubbio normal fault zone and in some quarries near the main fault.

Most of the slip data show a NE-SW extension. The minor faults show a relatively more complex kinematics compared to the major faults. Within a single stress regime, minor faults may register and reflect local stress complexities. The computed stress tensor is triaxial (*R*=0.6) and extensional, with a N50°E trending, nearly horizontal $\Uparrow_3$ axis and a nearly vertical $\Uparrow_1$ axis. The consistency of the stress ellipsoids calculated in the Gubbio sites with those in the PMA and in the eastern Tiber basin sites is indicated by the similarity in the direction of the main stress axes.

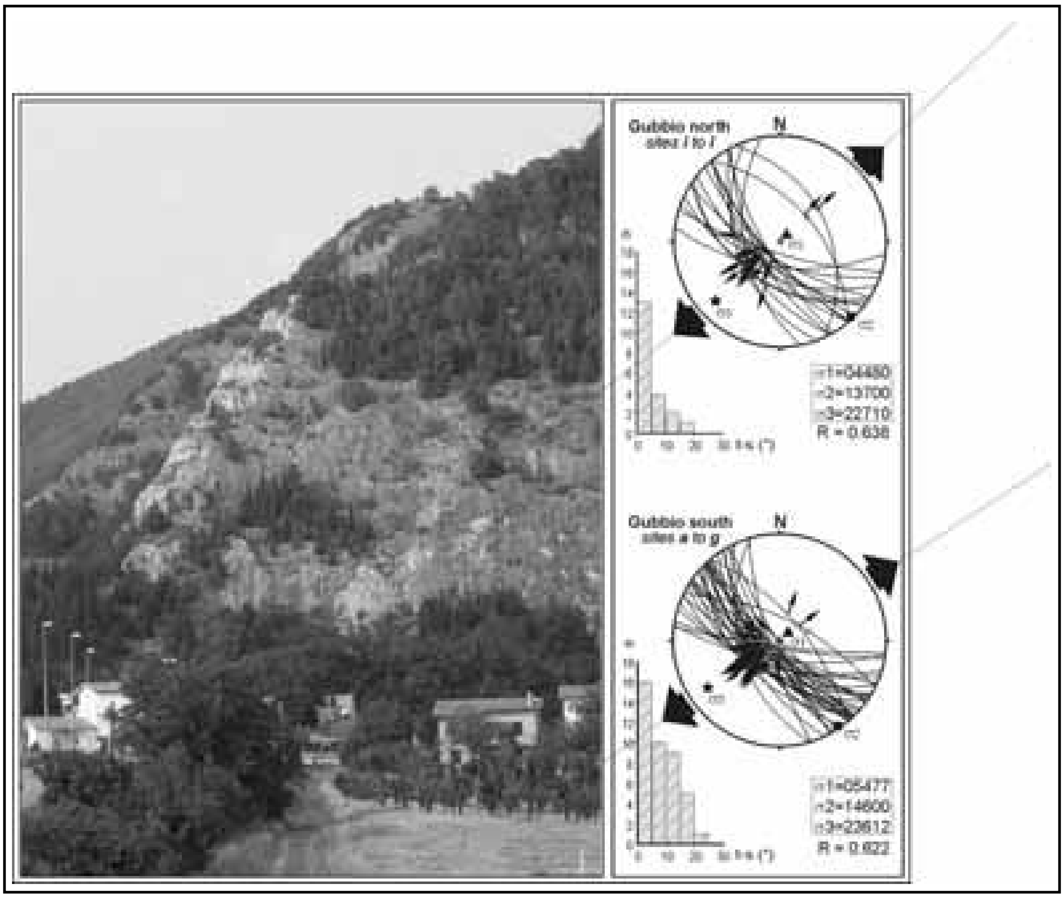

Figura 54 - Spectacular exposure of the Gubbio Fault plane near the Gubbio cemetery (site of Stop 35); Scaglia Rossa and Scaglia Variegata fms crops out in the hanging Wall, Late Pleistocene debris lies upon the plane and is dragged and deformed within the fault zone.The results of a tensorial analysis performed (Carey's 1976 inversion method) in several survey sites along the fault is shown in the two associated stereoplots.

References cited

Abers, G. A., 1991, Possible seismogenic shallow-dipping normal faults in the Woodlark-D'Entrecasteaux extensional province, Papua New Guinea, *Geology*, *19*, 1205-1208.

Abers, G. A., 1997, Shallow dips of normal faults during rapid extension: earthquakes in the Woodlark-D'Entrecasteaux rift system, Papua New Guinea, Jour. Geoph. Res., 102, 301-317.

Abbate, E., Bruni, P. and Sagri m., 1991, Sezione geologica dai monti del Chianti al Passo dei Mandrioli, Stud. Geol. Cam. Vol. Spec. 1991/1.

Alvarez, W., 1976, A former continuation of the Alps. Geol. Soc. Am. Bull., 87, 891-896.

Amaudric du Chaffaut, S., 1982, Les unités alpines à la marge orientale du massif cristallin corse, Travaux du Laboratoire de Géologie: Paris, presses de l'école normale supérieure, 133 p.

Ambrosetti, P., Bosi, C., Carraro, F., Cianfarini, N., Panizza, M., Papani, G., Vezzani, L. and Zanferrari, A., 1987, Neotectonic map of Italy: scale 1: 500.000. Prog. Fin. Geod. It., C.N.R..

Anderson, 1951, The dynamics of faulting. Oliver and Boyd, Edinburgh, 206 pp.

Arisi Rota F., Brondi A., Dessau G., Franzini M., Monte Amiata s.p.a., Stab. Minerario del Siele, Stea B. & Vighi L., 1971, I Giacimenti minerari della Toscana meridionale. Rend. Soc. It. Miner. e Petr., 27, 357-544.

Aruta, G., Bruni, P., Cipriani, N., 1998, The siliciclastic turbidite sequences of the Tuscan domain in the Val di Chiana-Val tiberina area (eastern Tuscany

and north-western Umbria). Mem. Soc. Geol. It., 52, 579-593.
Baldi, P., Bellani, A., Ceccarelli, A., Fiordelisi, P., Squarci P. and Taffi L., 1994, Correlazioni tra le anomalie termiche ed altri elementi geofisici e strutturali della toscana meridionale, St. Geol. Cam., Vol. Spec., 1994/1.
Baldwin, S.L., Lister G.S., Hill, E.,J., Foster, D.A. and Mc Dougall, I., 1993, Thermochronological constraints on the tectonic evolution of active metamorphic core complexes, D'Entrecasteaux Islands, Papua New Guinea, Tectonics, 12 611-628,.
Barchi, M. R., A. De Feyter, M. B. Magnani, G. Minelli, G. Pialli, and B. M. Sotera, 1998, Extensional tectonics in the Northern Apennines (Italy): Evidence from the CROP03 deep seismic reflection line, *Mem. Soc. Geol. It.*, *52*, 527-538.
Bartole R., 1995, The north Tyrrhenian-northern Apennines post collisional system: constraints for a geodynamic model. Terranova, 7, 7-30.
Bartolini, C., Bernini, M., Carloni, G.C., Costantini, A., Federici P.R., Gasperi, G., Lazzarotto, A., Marchetti G., Marranti, R., Papani, G., Pranzini, G., Sandrelli, F., Vercesi, P.G., Costaldini D. and Francavilla F., 1982, Carta neotettonica dell'Appennino Settentrionale. Note illustrative. Boll., Soc., Geol., It., 101, 523-549.
Benvenuti, M., 1997, Physical stratigraphy of the fluvio-lacustrine Mugello Basin (Plio-Pleistocene, northern Apennines, Italy), *G. Geol.*, *59*, 91-111.
Bertini G., Cameli, G.M., Costantini, A., Decandia F.A., Di Filippo M., Dini, I., Elter 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. St. Geol. Cam., Vol. Spec, 1991/1.
Bézert, P., and Caby, R., 1988, Sur l'âge post-bartonien des évènements tectono-métamorphiques alpins en bordure orientale de la Corse cristalline (Nord de Corte): Bull. Soc. Géol. France, v. 6, p. 965-971.
Boccaletti, M., Elter, P. and Guazzone G., 1971, Plate tectonics model for the development of the western Alps and Northern Apennines, Nature, 234, 108-111.
Boccaletti M., Decandia F.A., Gasperi G., Gelmini, R., Lazzarotto A. and Zanzucchi G., 1987, Carta strutturale dell'Appennino settentrionale. Note illustrative. Prog. Fin. Geodin. It, CNR, Pubbl. 429, 1-203.
Boncio, P., F. Brozzetti, F. Ponziani, M. R. Barchi, G. Lavecchia, and G. Pialli, 1998, Seismicity and extensional tectonics in the northern Umbria-Marche Apennines, *Mem. Soc. Geol. It.*, *52*, 539-556.
Boncio, P. and Lavecchia, G., 2000, A structural model for active extension in central Italy. Jour. Geodyn., 29, 233-244.
Boncio, P., Brozzetti, F. and Lavecchia G., 2000, architecture of a regional low-angle normal fault zone in central Italy. Tectonics, 19, 6, 1038-1055.
Bortolotti , V., Fazzuoli, M., Pandeli, E., Principi, G., Babbini, A. & Corti, S. 2001. Geology of central and Eastern Elba island, Italy. *Ofioliti*, 2001, 26, 97-150.
Brozzetti, F., 1995, Stile strutturale della tettonica distensiva nell'Umbria occidentale: l'esempio dei Massicci Mesozoici Perugini. Stud. Geol. Cam., Vol. Spec., 1995/1.
Brozzetti, F. and Luchetti, L., 2002, Stratigraphic constraints to the tectonic evolution of the Miocene foredeep in northern Italy, Atti del 3° seminario sulla cartografia Geologica, Bologna, febb. 2002.
Brozzetti , F., Boncio, P. and Pialli G., 2002, Early-Middle Miocene evolution of the Tuscan Nappe-Western Umbria foredeep system: insights from stratigraphy and structural analysis, Boll. Soc. Geol. It., 52, 319-332.
Brunet, C., Monié, P., Jolivet, L., and 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: Tectonophysics, v. 321, p. 127-155.
Buck, W. R., 1988, Flexural rotation of normal faults, Tectonics, 7, 959-975.
Burchfiel, B., Hodges K.V. and Royden L.H., 1987, Geology of the Panamint Valley-Saline Valley pull-apart system, California:Palinspastic evidences for low-angle geometry of a Neogene range-bounding fault, J.Geoph. Res., 92, 10, 422-426.
Burrus, J., 1984, Contribution to a geodynamic synthesis of the Provençal basin (north-western Mediterranean): Marine Geology, v. 55, p. 247-269.
Byerlee, J., D., 1978. Friction of rocks, Pure Appl. Geoph., 116, 616-626.
Calamai A., Cataldi, R., Squarci R. and Taffi, L., 1970, Geology, geophysics and hidrogeology of the Monte Amiata geothermal fields: I-Maps and comments. Geothermics, special issue, 1, 1-9.
Calamita, F., Cello, G., Centamore, E., Deiana, G., Micarelli, A., Paltrinieri, W. and Ridolfi M., 1991, Stile deformativo e cronologia della deformazione lungo tre sezioni bilanciate dall'Appennino Umbro Marchigiano alla Costa Adriatica, Studi Geol. Cam.,

Vol. Spec., 1991/1
Cann, J.R., Blackman, D.K., Smith, D.K., McAllister, E., Janssen, B., Mello, S., Avgerinos, E. Pascoe, A.R. and Escartin J., 1997. Corrugated slip surfaces formed at North Atlantic ridge-transform intersections. Nature 385, 329-332.
Carmignani, L., and Kligfield, R., 1990, Crustal extension in the northern Apennines: the transition from compression to extension in the Alpi Apuane core complex. Tectonics, 9, 1275-1303.
Carmignani, L., Decandia F.A., Disperati L., Fantozzi, P.L., Kligflield R., Lazzarotto A., Liotta D and Meccheri D., 2001, Inner northern Apennines, in Vai G.B. and Martini, I.P., Anatomy of an Orogen, Kluwer Academic Publishers.
Caron, J. M., 1977, Lithostratigraphie et tectonique des schistes lustrés dans les Alpes cottiennes septentrionales et en Corse orientale, 326 p.
Caron, J. M., Kienast, J. R., and Triboulet, C., 1981, High pressure-low temperature metamorphism and polyphase Alpine deformation at Sant' Andrea di Cotone (Eastern Corsica, France): Tectonophysics, v. 78, p. 419-451.
Chamot-Rooke, N., Gaulier, J. M., and Jestin, F., 1999, Constraints on Moho depth and crustal thickness in the Liguro-Provençal basin from a 3D gravity inversion: geodynamic implications, *in* Durand, B., Jolivet, L., Horvàth, F., and Séranne, M., eds., Geol. Soc. Special Publication: London, Geological Society, p. 37-62.
Civetta, L., Orsi, G., Scandone, P. and Pece, R., 1978, Eastward migration of the Tuscan anatectic magmatism due to anticlockwise rotation of the Apennines., Nature, 26, 604-606.
Collettini, C. and Sibson, 2001, Normal faults normal friction?, Geology, 29, 927-930.
Collettini, C. and Barchi, M. R., 2002. A Comparison of Structural Data and Seismic Images For Low-Angle Normal Faults in the Northern Apennines (Central Italy): Constraints on Activity. *Special Publication Journal of the Geological Society*, in press.
Dallan Nardi, L. and Nardi R., 1972, Schema stratigrafico e strutturale dell'Appennino settentrionale. Mem. Acc. Lunigianense G. Cappellini , 42, 214 pp.
Damiani, A.V., Minelli, G. and Pialli G., 1991, L'Unità Falterona-Trasimeno nell'area compresa fra la Val di Chiana e la Valle Tiberina : sezione Terontola – Abbazia di Cassiano. Stud. Geol. Cam. Vol. Spec. 1991/1.
Daniel, J. M. & Jolivet, L. 1995. Detachments faults and pluton emplacement: Elba island (Tyrrhenian Sea). *Bulletin Societ*è geologique France, 166, 341-354.
Daniel, J. M., and Jolivet, L., 1995, Interaction of detachments and granitic plutons during extension in the Tyrrhenian Sea (Elba island): Bull. Soc. géol. France, v. 166, p. 341-354.
Daniel, J. M., Jolivet, L., Goffé, B., and Poinssot, C., 1996, Crustal-scale strain partitioning: footwall deformation below the Alpine Corsica Oligo-Miocene detachement: J. Struct. Geol., v. 18, p. 41-59.
Davis, G. H., 1980, Structural characteristics of metamorphic core complexes, in Cordilleran Metamorphic Core Complexes, edited by M.D. Crittenden Jr., P.J., Coney and G.H., Davis, Mem Soc Geol Am, 153, 35-77
Davis, G.A., and Lister G.S., 1988, Detachment faulting in continental extension; perpectives from southwestern U.S. cordillera, Spec. Pap.Geol. Soc.Am.,218, 133-160
Davis, G.A., Anderson, J.L., Frost, E.G., Shackelford T.S., 1980, Mylonitization and detachment faulting in the Whipple-Buckskin Rawhide Mountains terrane, southeastern California and western arizona, in cordilleran Metamorphic core complexes, edited by M.D. Critenden jr, P.J., Comney and G.H. Davis, Mem Geol. Soc. Am., 153, 79-130.
Davison, I., 1989, Extensional domino faults tectonics: kinematic and geometrical constraints. Annales Tectonicae, 3, 12-24.
Decandia, F. A., A. Lazzarotto, D. Liotta, L. Cernobori, and R. Nicolich, 1998, The CROP 03 traverse: Insights on post-collisional evolution of the northern Apennines, *Mem. Soc. Geol. It., 52*, 427-439.
Durand Delga, M., 1984, Principaux trait de la Corse alpine et corrélation avec les Alpes ligures: Mem. Soc. Geol. It., v. 28, p. 285-329.
Egal, E., 1992, Structures and tectonic evolution of the external zone of Alpine Corsica: J. Struct. Geol., v. 14, p. 1215-1228.
Elter P., Giglia G. and Trevisan L., 1975, Tensional and compressional areas in the recent (Tortonian to present) evolution of the Northern Apennines Boll. Geofis. Teor. Appl., 27, 3-18.
Elter P. and Marroni M., 1991, Le unità liguri dell'Appennino settentrionale: sintesi dei dati e nuove interpretazioni. Mem Descr.Carta geol. It., 44, 121-138.
Ferrandini, M., Ferrandini, J., Löye-Pilot, M.-D., Butterlin, J., Cravatte, J., and Janin, M. C., 1996, Le Miocène du bassin de Saint Florent (Corse): modalités

de la transgression du Burdigalien supérieur et mise en évidence du Serravalien: Geobios, v. 31, no. 1, p. 125-137.
Fournier, M., Jolivet, L., Goffé, B., and Dubois, R., 1991, The Alpine Corsica metamorphic core complex: Tectonics, v. 10, p. 1173-1186.
Gattacceca, J., 2000, Cinématique du bassin liguro-provençal entre 30 et 12 Ma. Implications géodynamiques (Thèse de Doctorat thesis), 293 p.
Gautier, P., and Brun, J. P., 1994, Crustal-scale geometry and kinematics of late-orogenic extension in the central Aegean (Cyclades and Evvia island): Tectonophysics, v. 238, p. 399-424.
Gibbons, W., and Horak, J., 1984, Alpine metamorphism of Hercynian hornblende granodiorite beneath the blueschists facies Schistes Lustrés nappe of NE Corsica: J. Metamorphic Geol., v. 2, p. 95-113.
Giglia G. and Radicati di Brozolo, R., 1970, K/Ar age of metamorphism in the Apuane Alps (northern Tuscany). Boll. Soc. Geol. It.,89, 485-497.
Goetze, C., and Evans, B., 1979, Stress and temperature in the bending lithosphere as constrained by experimental rock mechanics: Geophys. J. R. Astron. Soc., v. 59, p. 463-478.
GSA National Annual Meeting, Abstracts with program, v. 16, no. 6, 1984.
Gueguen, E., 1995, Le bassin Liguro-Provençal: un véritable océan. Exemple de segmentation des marges et de hiatus cinématiques. Implications sur le processus d'amincissement crustal. (Phd thesis): Université de Bretagne Occidentale.
Gueydan, F., 2001, La transition fragile-ductile de la croûte continentale en extension, du terrain à la modélisation: Université Pierre et Marie Curie, 356 pp.
Gueydan, F., Leroy, Y., and Jolivet, L., 2001, Grain-size sensitive flow and shear stress enhancement at the brittle to ductile transition of the continental crust: Int. Journ. Earth Sciences, v. 90, no. 1, p. 181-196.
Gueydan, F., Leroy, Y., Jolivet, L., and Agard, P., 2003b, Analysis of continental midcrustal strain localization induced by microfracturing and reaction-softening: J. Geophys. Res., v. 108, no. B2, p. 2064, doi:10.1029/2001JB000611.
Hamilton W., 1988, Extensional faulting in the death Valley region, Geol. Soc. Am. Abs. with programs, 20, 165-166.
Hoogerduijn Strating, E.H., E. Rampone, G.B. Piccardo, M.R. Drury, and R.L.M. Vissers., 1993, Subsolidus emplacement of mantle peridotites during incipient oceanic rifting of the Mesozoic Tethys (Voltri Massif, NW Italy). J. of Petrology, 34, 901 – 927.
Howard K.A. and John B.E., 1987, crustal extension along a rooted system of imbricate low-angle faults: Colorado River extensional corridor, California and Arizona, in Continental Extensional Tectonics, ed. By Coward M.P., dewey, J. F., and Hancock, Geol. Soc., Spec. Publ., London, 28, 299-311.
Jackson, J.A., and White, N.J., 1989, Normal faulting in the upper continental crust: observation from regions of active extension. J. Struct. Geol. 11, 15-36.
Johnson, R. A., and K. L. Loy, 1992, Seismic reflection evidence for seismogenic low-angle faulting in southeastern Arizona, *Geology*, *20*, 597-600.
Jolivet, L., Daniel, J. M., and Fournier, M., 1991, Geometry and kinematics of ductile extension in alpine Corsica: Earth and Planetary Science Letters, v. 104, p. 278-291.
Jolivet, L., Dubois, R., Fournier, M., Goffé, B., Michard, A., and Jourdan, C., 1990, Ductile extension in Alpine Corsica: Geology, v. 18, p. 1007-1010.
Jolivet, L., Faccenna, C., d'Agostino, N., Fournier, M., and Worrall, D., 1999, The Kinematics of Marginal Basins, examples from the Tyrrhenian, Aegean and Japan Seas, *in* Mac Niocaill, C., and Ryan, P. D., eds., Continental Tectonics: Geol. Soc. Spec. Pub.: London, Geological Society, p. 21-53.
Jolivet, L., Faccenna, C., Goffé, B., Mattei, M., Rossetti, F., Brunet, C., Storti, F., Funiciello, R., Cadet, J. P., and Parra, T., 1998, Mid-crustal shear zones in post-orogenic extension: the northern Tyrrhenian Sea case: J. Geophys. Res., v. 103, no. B6, p. 12123-12160.
Jolivet, L., and Patriat, M., 1999, Ductile extension and the formation of the Aegean Sea, *in* Durand, B., Jolivet, L., Horvàth, F., and Séranne, M., eds., Geol. Soc. Special Publication: London, Geological Society, p. 427-456.
Jourdan, C., 1988, Balagne orientale et massif du Tende (Corse septentrionale). Etude structurale, interprétation des accidents et des deformations, reconstitutions géodynamiques: Paris Sud, Orsay (France), 246 p.
Lavecchia, G., 1988, The Tyrrhenian-Apennines system: Structural setting and seismotectogenesis, *Tectonophysics*, *147*, 263-296.
Lavecchia, G. and stoppa F., 1989, Tettonica e magmatismo nell'Appennino Settentrionale lungo la Geotraversa Isola del Giglio- Monti Sibillini. Boll. Soc. Geol. It., 108, 237-254.

Lavecchia, G., Minelli G., and Pialli G., 1984, L'Appennino Umbro-marchigiano: tettonica distensiva ed ipotesi di sismogenesi. Boll. Soc. Geol. It., 103, 467-476.

Lavecchia, G., Creati, N. and Boncio, P., 2002, The intra-montane Ultra-alkaline province (IUP) of Italy: a brief review with considerations on the thickness of the underlying lithosphere. Boll. Soc. Geol., It., Special Vol. 1, 87-98.

Lazzarotto, A., 1967, Geologia della zona compresa fra l'alta valle del fiume Cornia ed il Torrente Pavone (Prov. Pisa e Grosseto). Mem. Soc. Geol. It., 6, 151-197.

Le Pichon, X. and sibuet J.C., 1982, Passive margins: a model of formation. J. Geophys. Res. 86, 3708-3720.

Liotta D., 2002, Stratigraphic and structural outline of the Montagnola senese area (southern Tuscany), Boll. Soc. Geol. It., vol. spec. 2002, 1, 705-713.

Lister, G.S., Davis, G.A., 1989, The origin of metamorphic core complexes and detachment faults formed during tertiary continental extension in the northern Colorado River Region, USA. J.Struct. Geol., 11, 65-93.

Keller, J.V.A., Minelli, G. & Pialli, G. 1994. Anatomy of late orogenic extension: the Northern Apennines case. *Tectonophysics*, 238, 275-294.

Keller, J. V. A. & Coward, M. P., 1996, The structure and evolution of the Northern Tyrrhenian Sea. *Geological Magazine*, 133, 1-16.

Keller, J. V., and Pialli, G., 1990, Tectonics of the island of Elba: a reappraisal: Boll. Soc. Geol. It., v. 109, p. 413-425.

Kligfield, R., 1979, The Northern Apennines as a collisional orogen. Am. Jour. of Science, 279, 676-691.

Kligfield, R., Hunziker, J., Dallmeyer R.D., Schamel S., 1986, dating of deformational phases using K-Ar and 40 Ar/39Ar techniques: results from the northern Apennines. Jour. Struct. Geol., 8, 781-798.

Kuznir, N. J., and Park, R. G., 1987, The extensional strength of the continental lithosphere: its dependance on geothermal gradient, and crustal composition and thickness, *in* Coward, M. P., Dewey, J. F., and Hancock, P. L., eds., Continental Extensional Tectonics: Geological Society Special Publication, p. 35-52.

Lahondère, D., 1988, Le métamorphisme éclogitique dans les orthogneiss et les métabasites ophiolitiques de la région de Farinole (Corse): Bull. Soc. géol. France, v. 4, p. 579-586.

Lahondère, D., and Guerrot, C., 1997, Datation Sm-Nd du métamorphisme éclogitique en Corse alpine : un argument pour l'existence au Crétacé supérieur d'une zone de subduction active localisée sous le bloc corso-sarde: Géologie de la France, v. 3, p. 3-11.

Lavecchia, G., 1988, The Tyrrhenian-Apennines system: Structural setting and seismotectogenesis: Tectonophysics, v. 147, p. 263-296.

Lavecchia, G., and Stoppa, F., 1989, Il 'rifting' tirrenico: delaminazione della litosfera continentale e magmatogenesi: Boll. Soc. Geol. Ital., v. 108, p. 219-235.

Le Pichon, X., and Chamot-Rooke, N., 1991, Extension of continental crust, Controversies in Modern Geology, Academic Press Ltd, p. 313-338.

Lister, G. S., and Davis, G. A., 1989, The origin of metamorphic core complexes and detachment faults formed during Tertiary continental extension in the northern Colorado River region, U.S.A.: Journal of Structural Geology, v. 11, no. 1/2, p. 65-94.

Malinverno, A., and Ryan, W., 1986, Extension in the Tyrrhenian sea and shortening in the Apennines as result of arc migration driven by sinking of the lithosphere: Tectonics, v. 5, p. 227-245.

Mascle, J., and Réhault, J. P., 1990, A revised seismic stratigraphy of the Tyrrhenian Sea: implications for the Basin evolution, *in* Kastens, K. A., Mascle, J., and al., e., eds., Proc. ODP, Scientific Results: Proc. ODP, Scientific Results, p. 617-636.

Mattauer, M., Faure, M., and Malavieille, J., 1981, Transverse lineation and large scale structures related to Alpine obduction in Corsica: J. struct. Geol., v. 3, p. 401-409.

Mauffret, A., and Contrucci, I., 1999, Crustal structure of the North Tyrrhenian Sea: first results of the multichannel seismic LISA cruise, *in* Durand, B., Jolivet, L., Horvàth, F., and Séranne, M., eds., The Mediterranean basins: Tertiary extension within the Alpine Orogen: Special Publications: London, Geological Society, p. 169-194.

Miller, E. L., Gans, P. B., and Garling, J., 1983, The Snake river decollement: an exumed mid-Tertiary brittle-ductile transition: Tectonics, v. 2, p. 239-263.

Mitra, G., 1978, Ductile deformation zones and mylonites: The mechanical processes involved in the deformation of crystalline basement rocks: Amer. Jour. Sci., v. 278, p. 1057-1084.

Orszag-Sperber, F., and Pilot, M.-D., 1976, Grand trait du Néogène de Corse: Bulletin de la Société Géologique de France, v. 28, p. 1183-1187.

Pascucci, V., Merlini, S. & Martini, I. P. 1999. Seismic

stratigraphy in the Miocene-Pleistocene sedimentary basins of the Northern Tyrrhenian Sea and western Tuscany (Italy). *Basin Research*, 11, 337-356.
Pertusati, P.C., Raggi, G., Ricci C., A., Duranti , S. and Palmieri, R., Evoluzione post-collisionale dell'Elba centro-orientale. Mem. Soc. Geol. It.,49, 223-312.
Pialli G., BarchiM., Minelli G., eds: Results of the CROP03 deep seismic reflection profile. Mem. Soc. Geol. It., 52, pp 657.
Proffett, J.M., Jr.,1977, Cenozoic Geology of the Yerington District, Nevada, and implicatyion for nature and origin of Basin and Range faulting, Geol. Soc., Am., Bull., 88, 247-266.
Ranalli, G., 1995, Rheology of the Earth, second edition: London, Chapman & Hall, 413 p.
Réhault, J. P., Moussat, E., and Fabbri, A., 1987, Structural evolution of the Tyrrhenian back-arc basin: Mar. Geol., v. 74, p. 123-150.
Rocchi, R., Westerman, S. D., Dini, A., Innocenti, F., & Tonarini, S. 2002. Two-stage of laccoliths at Elba Island, Italy. *Geology*, 30, 983-986.
Rigo, A., Lyon-Caen, H., Armijo, R., Deschamps, A., Hatzfeld, D., Makropoulos, K., Papadimitriou, P., and Kassaras, I., 1996, A microseismicity study in the western part of the Gulf of Corinth (Greece): implications for large-scale normal faulting mechanisms: Geophys. J. Int., v. 126, p. 663-688.
Rossetti, F., Faccenna, C., Jolivet, L., and Funiciello, R., 1999a, Structural evolution of the Giglio island, Northern Tyrrhenian Sea (Italy): Mem. Soc. Geol. It., v. 52, p. 493-512.
Rossetti, F., Faccenna, C., Jolivet, L., Tecce, F., Funiciello, R., and Brunet, C., 1999b, Syn- versus post-orogenic extension in the Tyrrhenian Sea, the case study of Giglio Island (Northern Tyrrhenian Sea, Italy): Tectonophysics, v. 304, p. 71-93.
Sartori, R., 1990, The main results of ODP Leg 107 in the frame of neogene to Recent geology of peri-Tyrrhenian areas, *in* Kastens, K. A., Mascle, J., and al., e., eds.: Proc. Scientific Results, ODP, p. 715-730.
Sartori, R., Mascle, G., and Amaudric du Chaffaut, S., 1987, A review of circum-Tyrrhenian regional geology, *in* Kastens, K., Mascle, J., Auroux, C., and *et al*., eds.: Proc. Init. Repts, ODP, p. 37-63.
Saupé, F., Marignac, C., Moine, B., Sonet, J. & Zimmerman, J. L. 1982. Datation par les mèthodes K/Ar et Rb/Sr de quelques roches de la partie orientale de l'ile d'Elbe (Province de Livourne, Italie)., *Bulletin de Minéralogie*, 105, 236-245.
Serri, G., Innocenti F. and Manetti P., 1993, Geochemical and Petrological evidence of the subduction of delaminated Adriatic continental lithosphere in the genesis of the Neogene-Quaternary magmatism of central Italy. Tectonophysics, 223, 117-147.
Sibson, R. H., 1985, a note on fault reactivation. J.Struct.geol. 7, 751-754.
Spencer, J.E., 1984, Role of tectonic denudation in warping an uplift of low-angle normal faults, geology, 12, 95-98,1984.
Spencer, J.E., 1985, Miocene low-angle normal faulting and dike emplacement, Homer mountain and surrounding areas, southeastern California and southernmost nevada, Geol. Soc. Am. Bull., 9, 1140, 1155.
Speranza, F., Villa, I. M., Sagnotti, L., Florindo, F., Cosentino, D., Cipollari, P., and Mattei, M., 2002, Age of Corsica-Sardinia rotation and Liguro-Provencal basin spreading : new paleomagnetic and Ar/Ar evidence: Tectonophysics, v. 347, p. 231-251.
Trevisan, L., 1950. L'Elba orientale e la sua tettonica di scivolamento per gravità. *Memorie dell'Istituto Geologico dell'Università di Padova*, 16, 1-30.
Trevisan, L., Marinelli, G., Barberi, F., Giglia, G., Innocenti, F., Raggi, G., Squarci, P., Taffi, L. & Ricci, C. A. 1967. Carta Geologica dell'Isola d'Elba. Scala 1:25.000. *Consiglio Nazionale delle Ricerche*, Gruppo di Ricerca per la Geologia dell'Appennino centro-settentrionale e della Toscana. Pisa, 1967.
Warburton, J., 1986, The ophiolite-bearing Schistes Lustrés nappe in Alpine Corsica: a model for the emplacement of ophiolites that have suffered HP/LT metamorphism: Geol. Soc. Am. Memoir, v. 164, p. 313-331.
Waters, C. N., 1989, The metamorphic evolution of the Schistes lustrés ophiolite, Cap Corse, Corsica, *in* Daly, J. S., Cliff, R. A., and Yardley, B. W. D., eds., Evolution of Metamorphic Belts: Spec. Publs geol. Soc. Lond., p. 557-562.
Westphal, M., Orsini, J., and Vellutini, J., 1976, Le microcontinent corsosarde, sa position initiale: données paléomagnétiques et raccords géologiques: Tectonophysics, v. 30, p. 141-157.
Wernicke, B.,1981; Low-angle normal faults in the Basin and Range province: Nappe tectonics in an extending orogen, Nature, 291, 645-648.
Wernicke, B., and Burchfiel B.C., 1982, Modes of extensional tectonics. J. Struct. Geol., 4, 105-115,.
Wernicke, B., 1985, Uniform-sense normal simple shear of the continental lithosphere, *Can. J. Earth Sci.*, *22*, 108-125.

Wernicke, B., 1995 Low-angle normal faults and seismicity: A review, *J. Geophys. Res.*, *100*, 20,159-20,174.

Wernicke, B., and G. J. Axen, 1988, On the role of isostasy in the evolution of normal fault system, *Geology*, *16*, 848-851.

Westaway, R., 1999, The mechanical feasibility of low-angle normal faulting, Tectonophysics, 308, 407-443.

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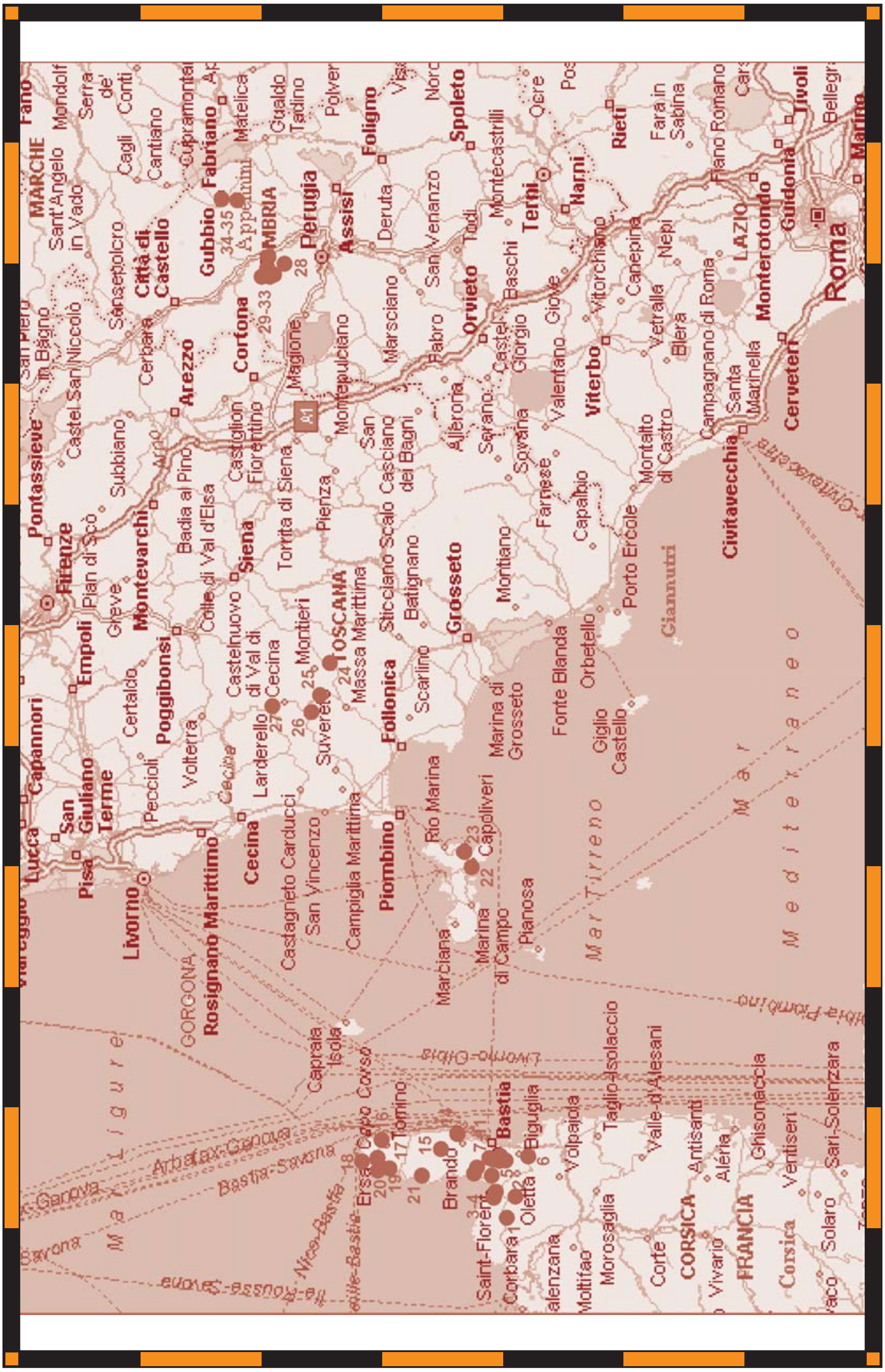
field trip itinerary

32nd INTERNATIONAL GEOLOGICAL CONGRESS

FIELD TRIP MAP



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