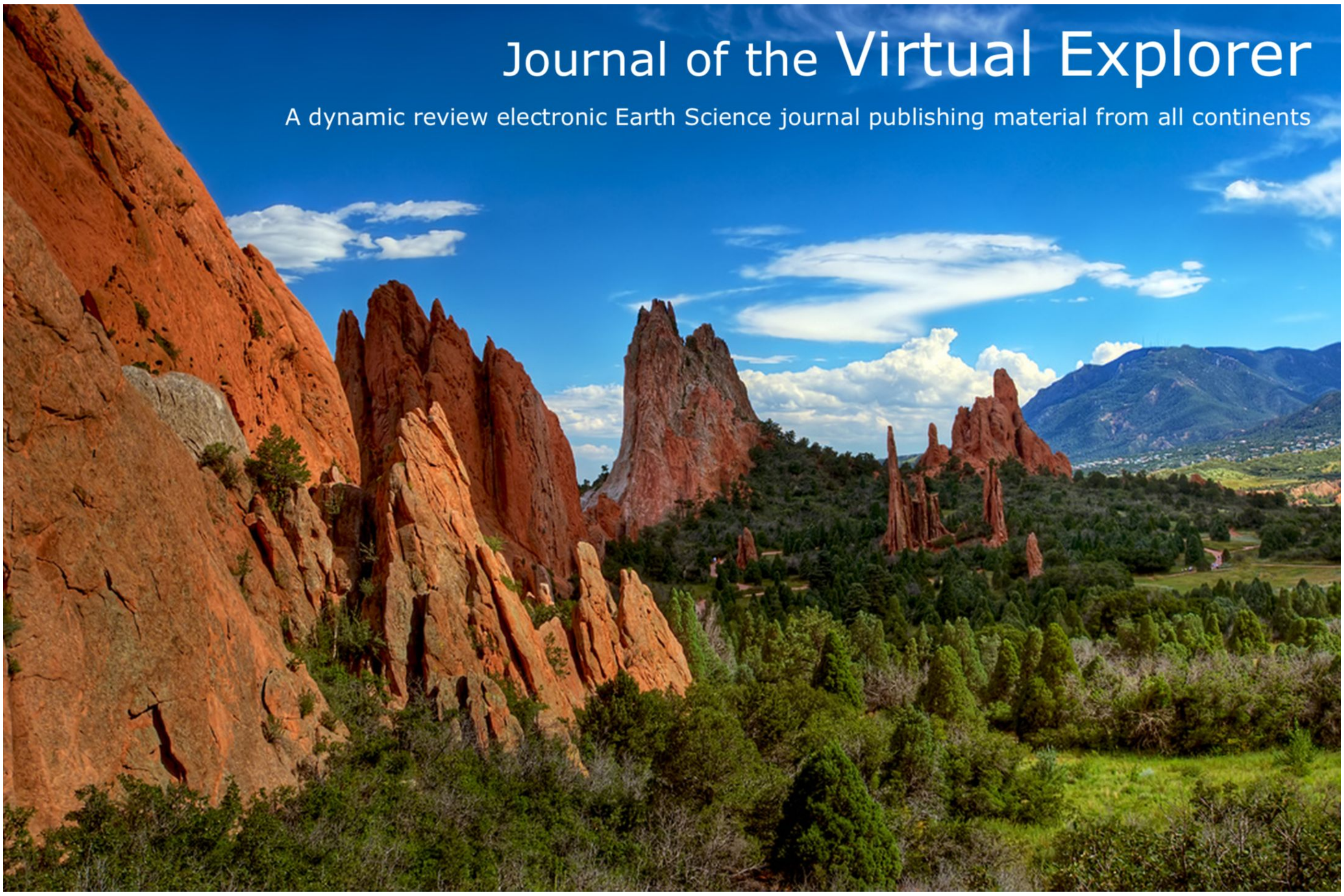

Field Trip 2 - General Architecture and tectonic evolution of Alpine Corsica. Insights from a transect between Bastia and the Balagne region.

Jacques Malavieille, Giancarlo Molli, Michel Ferrandini, Jean Ferrandini, Marie Madeleine Ottaviani Spella, Alberto Vitale Brovarone, Olivier Beyssac, Laurent Ciancaleoni, et al

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Field Trip 2 - General Architecture and tectonic evolution of Alpine Corsica. Insights from a transect between Bastia and the Balagne region.

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Abstract: CorseAlp 2011 was held in April 2011. A link to the website is here: CorseAlp 2011. The CorseAlp field trips were designed to introduce the participants to the general features of the geology of Alpine Corsica and to recent specific discoveries.

The three-day post-workshop field trip (Field trip 2) focused on the general architecture of Alpine Corsica in an EW transect between the Balagne region and Bastia.

First day: The internal zones of Alpine Corsica: from Bastia to Saint Florent. Intinerary: Saint Florent, Serra di Pigno, Cima Malaspina, Patrimonio, Saint Florent.

Second day: The Tenda massif and its eastern margin. Intinerary: Saint Florent, Casta, Fontana Porragghia, Rue de Morello, Fontana a Murello quarry; Fornali.

Third day: The external zones of Alpine Corsica in the Balagne region. Intineray: Saint Florent, Colle di U'Vezzu, Ostriconi, Belgodere, San Sebastiano Chapel, San Colombano, Ponte Leccia, Francardo, Bastia.

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Field trip 2: first day

The internal zones of Alpine Corsica: between Bastia and Saint Florent

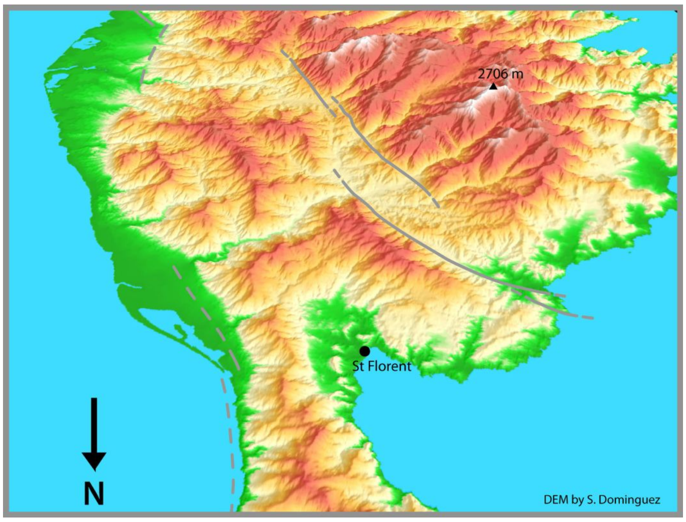
Intinerary: Saint Florent, Serra di Pigno, Cima Malaspina, Patrimonio, Saint Florent

The Internal zones of Alpine Corsica are well exposed between Bastia and S. Florent where most of the geological history of the orogen can be investigated. This multi-stage evolution ranges from the early stages of subduction to the late stages of post-orogenic extension and opening of oceanic domains, which are recorded in the structures found within the different tectonic units and in the sedimentary basins exposed in the area.

The nappe structure characterizing the area crops out in a regional scale open antiform-synform pair: the NS trending Serra di Pigno antiform and the Nebbio synform, both with an amplitude of nearly 5km. The Serra di Pigno antiform is developed in a doubly plunging axial depression of the c. 50km long Cap Corse-Castagniccia antiform.

The eastern and western limbs of the Serra di Pigno antiform are reworked by high angle normal faults (of dominant extensional to transtensional type), Miocene to present in age, which are responsible for the regional morphology (Fig. 2.1).

Figure 2.1. Digital Terrain Model of northern part of Corsica by S. Dominguez.



East of Serra di Pigno the alluvial plain of Marana (south of Bastia) occupies the northern part of the Aleria coastal plain. This domain represents the on-land part of the Corsica-channel basin, which separates Corsica from the Tuscan Archipelago (e.g. Mauffret *et al.*, 1999). West of Aleria, the oldest onland marine sediments of this basin are Mid-Burdigalian in age (St. Antoine fm., 18.7-18.3 Ma), as determined with nannoplacton and planktonic foraminifera (Loye-Pilot *et al.*, 2004).

West of Serra di Pigno the Neogene depositional system (Fig. 2.2) can be observed within the c.500 m thick carbonate rich sequence of the Saint Florent basin (Durand Delga, 1976; Dallan and Puccinelli, 1986; Rossi *et al.*, 1994; Dallan and Puccinelli, 1996; Ferrandini *et al.*, 1998; Fellin *et al.*, 2005; Cavazza *et al.*, 2007). The oldest sedimentary unit (F.Albino fm.) is formed by continental alluvial/fluvial deposits (conglomerates and pebbly sandstones) inferred to be early Burdigalian in age (Dallan and Puccinelli, 1996; Ferrandini *et al.*, 1998). Conglomerate clasts in the F.Albino formation were mostly derived from the different metamorphic and non-metamorphic alpine units. The basal deposits are covered by ~250 m thick marine carbonates starting with sandy marls, grey limestones and calcarenites (Mt. Torra fm.) followed by calcarenites characterized by spectacular cross-laminations (Fig. 2.3) and a fauna rich in bryozoans of the Mt. S.Angelo fm. (Orszag-Sperber and Pilot, 1976; Ferrandini *et al.*, 1998). The age of this unit is Mid-Burdigalian-early Langhian based on foraminifera and nannoplacton biostratigraphy (Dallan and Puccinelli, 1996; Ferrandini *et al.*, 1998, Cavazza *et al.*, 2007) and on magnetostratigraphy (Vigliotti and Kent, 1990; Ferrandini *et al.*, 2003). The Mt. S.Angelo fm. is followed by marls, muddy arenites, limestones (Farinole fm) of Langhian-Serravalian age and by the Tortonian S. Florent conglomerates (Ferrandini *et al.*, 1998; Ferrandini *et al.*, 2003; Fellin *et al.*, 2005; Cavazza *et al.*, 2007).

Figure 2.2. Stratigraphic log of the Saint Florent Miocene basin after Ferrandini *et al.* 1996.

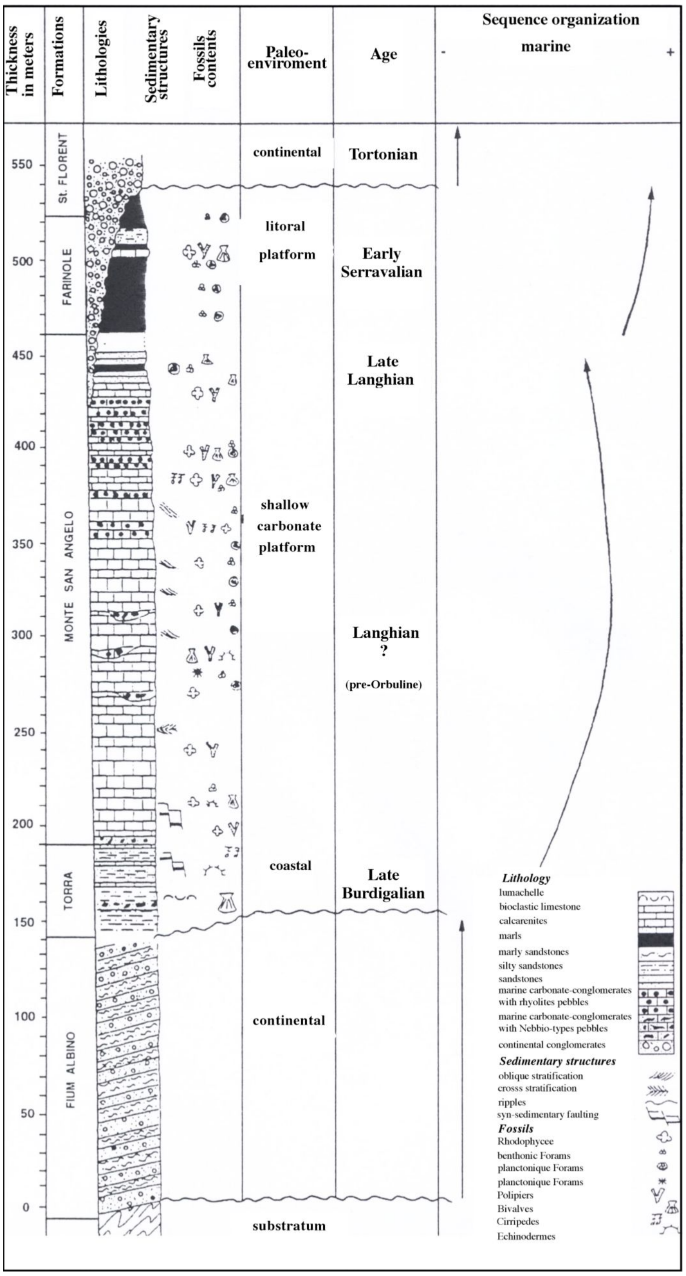


Figure 2.3. Cross-bedded medium grained carbonatic sandstones.

Cross-bedded medium grained carbonatic sandstones (interpreted as shoreface deposits) of the Miocene of Saint Florent.

The Neogene deposits of the S. Florent basin unconformably overlie the Alpine nappe stack, including the non-metamorphic Nappe Supérieure (Nebbio) (Fig. 2.4, 2.5) and the H*P*/L*T* Schistes Lustrés nappe (Dallan and Puccinelli, 1996; Rossi *et al.*, 2001). The basal unconformity is deformed by a kilometer-scale open synform which records, altogether with the internal geometry of sediments and their thickness variation (Ferrandini *et al.*, 1998; Fellin *et al.*, 2005), the syn- and post-sedimentary activity of high angle normal to transtensonal fault-systems. The faults bounding the basin were active at ~18-16 Ma, during the final stages of the Corsica-Sardinia block rotation (Vigliotti and Kent, 1990; Ferrandini *et al.*, 2003; Speranza *et al.*, 2002; Gattacceca *et al.*, 2007).

Figure 2.4. Schematic cross-section from Bastia to Saint Florent.

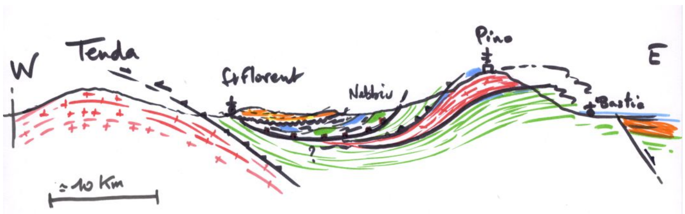
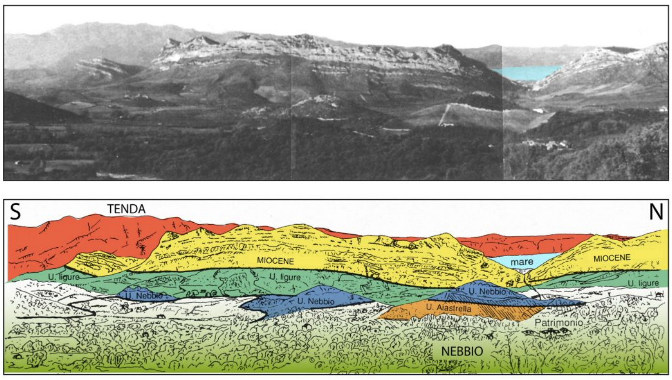


Figure 2.5. Panoramic view from Serra di Pigno



Panoramic view from Serra di Pigno toward west and schematic line drawings showing the different units of the Nebbio Nappe System (modified after Dallan and Puccinelli, 1996).

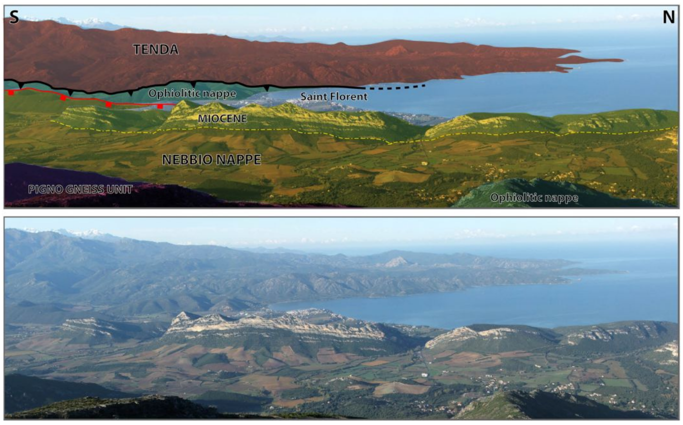
Nappe Architecture

The nappe architecture of the internal zones cropping out between Bastia and S. Florent, includes from the top to the bottom:

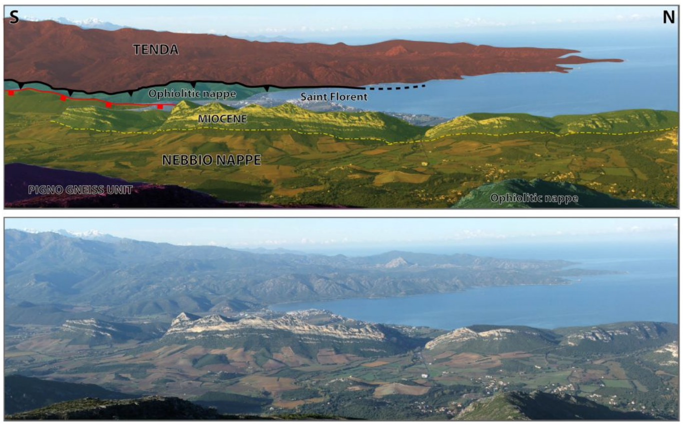
The Nappe Supérieure System

The "Nappe Supérieure" System ("Allochtone du Nebbio"), which consists of three major units that escaped significant deformation/metamorphism in the accretionary-collisional orogenic evolution. These units, in superstructure position, show a very low grade alpine overprint and structures developed at shallow crustal depth. Their complete description will be fully analyzed during the third day of the field trip 2. Following Dallan and Puccinelli (1996) and Rossi *et al.* (2001) from top to bottom the following units (Fig. 2.3) can be recognized (see panoramic view of Fig. 2.6 and Fig. 2.4):

Figure 2.6. Panoramic views



Panoramic view from Serra di Pigno toward Tenda Massif, with the main stack of units from the Serra di Pigno to the Tenda Massif.



Schematic panoramic view of the Nebbio region.

The Ligurian unit (also called Mortola-Canta Furmigola and Tramonti units) characterized by Jurassic basalts and a Late Jurassic-Late Cretaceous sedimentary cover including radiolarites, Calpionella limestone, and deep water pelagic Cretaceous sediments (Lydienne). The sequence is capped by Late Cretaceous ophiolite-rich breccias (Toccone breccia) and mixed ophiolite and continental crust-derived conglomerates (Alturaja fm); 1.2 The Nebbio Unit made up of a Late Cretaceous calcareous flysch (cfr. Flysch d'Ostriconi and Macinaggio) and Eocene siliciclastic deposits (coarse grained sandstones and conglomerates) including slices and/or olistoliths of Triassic-Jurassic limestone (Monte Tuda and Tramonti limestone);1.3 The Aiastrella unit (Unité inférieure du Nebbio) including pre-Hercynian phyllites and schists ("roches brunes"), Permo-Triassic conglomerates and quarzites and Triassic dolomites.

Bad outcropping conditions in the Nebbio region prevent a clear definition of the relationship between the different terms of the Nappe Supérieure system, which have been interpreted in the literature as sliced remnants of coherent units and/or detritic components within the Eocene deposits (see Dallan and Puccinelli, 1986; Rossi *et al.*, 2001 and references therein).

The Schistes Lustrés nappe

The “Schistes Lustrés” nappe, is composed by several tectonic units, some of which include metaophiolites (mantle-derived rocks, metagabbros and metabasalts) and related metasedimentary sequences (quarzites, marbles and calcschists). Slices of polycyclic continental basement (para- and orthogneiss with gabbro bodies and basic dykes) and Mesozoic cover series can be observed as well as extensive metasedimentary units whose original attribution to continental or oceanic sequence has been disputed (Durand Delga, 1984; Lahondere 1996; Rossi *et al.* 1998; Rossi *et al.*, 2001 and references therein). Recent studies (Serié, 2002; Meresse, 2006; Vitale Brovarone, 2011) showed that the association of continental slices and ophioliteswas established along a Mesozoic OCT (ocean-continent transition domain; see Field trip 1).

Following this interpretation and the results of new petrological studies (Ravna *et al.*, 2010; Vitale Brovarone *et al.*, 2010) including Raman Spectroscopy of Carbonaceous Material (RSCM) thermometry (Vitale Brovarone *et al.*, 2010; 2011) the Schistes Lustrés nappe system can be subdivided from top to bottom into four different tectonic units:

1 - an epidote-blueschist (Ep-BS) oceanic unit mainly exposed in the two limbs of the Nebbio synform, resting to the west on top of the Serra di Pigno lawsonite-blueschist (Law-BS) unit and to the east on top of the Tenda unit;

2 - a (Law-BS) OCT-type unit that, similarly to the previous unit, consists of metasediments associated with ophiolites and continental crust slices outcropping on top of Serra di Pigno;

3 - a lawsonite-eclogite (Law-Ecl) OCT-type unit formed by rare metasediments associated with ophiolites and continental crust slices, which crops out at the bottom of the Lancone valley and extensively to the west of Bastia (between Bastia and Serra di Pigno). *P-T* conditions in these units have been recently redefined by Vitale Brovarone *et al.* (2011) at 520 ± 20°C and 2.3 GPa;

4 - the lower Castagniccia unit characterized by a metasedimentary cover (calcschists and impure marbles) inferred to be Cretaceous in age (Caron, 1990). Lawsonite and chloritoid characterize these metasediments that locally preserve carpholite. In this unit the absence of mafic rocks precludes the definition of precise pressure estimates whereas RSCM thermometry yields an average value at 470°C (Vitale Brovarone, 2011);

All units show H*P*/L*T* peak metamorphism testifying their involvement in subduction processes at different depth, from 1 GPa and 300-400°C up to 2.3 GPa and 470-520°C (Vitale Brovarone *et al.*, 2011); later exhumation-related retrogression is highly variable (depending on rock-types and structural positions). A pervasive composite mylonitic foliation with relicts of earlier fabrics can be found in the Law-eclogite unit. Stretching lineation and shear sense indicators (Fig. 2.4 Map Serra di Pigno) associated with eclogite facies fabrics provide evidence of scarce top north/northwest shearing (Lahondère 1996), whereas the blueschist retrograde fabric formed the mappable main foliation associated with an east–west stretching lineation and dominant top-to-west kinematics (Faure & Malavieille 1981; Mattauer *et al.* 1981; Malavieille 1983; Harris 1985a, b). Greenschist retrogression is generally static, except for the metasediments, where it is locally associated with planar fabrics. In continental rocks and metabasalts the greenschist-facies retrogression is shown by the pseudomorphic growth of albite and chlorite porphyroblasts after blue-amphibole or Na pyroxene and garnet, respectively.

3) The Tenda unit. In the westernmost limb of the Nebbio synform the continent-derived gneisses and granitoids of the Tenda unit are exposed below the composite Schistes Lustrés nappe represented by the epidote-blueschist oceanic unit. The Tenda unit and the features of its eastern boundary will be the main topic of the second day of the Field trip (see below).

The conditions of coupling of the different units in the Cap Corse-Castagniccia antiform and the related kinematics are still to be worked out in details, being resolved only locally for the contact between the lower Castagniccia unit (2.4) and the Law-eclogite unit (2.3). The juxtaposition between these two units is well constrained to the exhumation history in the Mont San Petrone area (Vitale Brovarone et al, 2011), where the contact is characterized by epidote-blueschist fabrics parallel to the eclogitic foliations of the Law-eclogite unit.

The relationships between the Tenda and the overlying ophiolitic unit will be discussed during the second day of this Field trip.

Our general interpretation follows the early propositions of Mattauer *et al.* (1981) in considering the different contacts between internal units as related with sheared folds, refolding nappe contacts interleaving continental-derived and OCT-derived units during continental subduction and syn-contractional exhumation (Molli and Malavieille, 2010).

Day One Itinerary

Stop 1.1

Locality: Serra di Pigno summit (UTM 32T 532786 E 4727064)

Themes: Panoramic view and overall architecture of Alpine Corsica from Tenda Massif to Serra di Pigno and Cap Corse. The Nebbio synform and S.Florent sedimentary basin. Deformation and metamorphism of continental slices and Schistes Lustrés units.

After the description of the panoramic view from the Serra di Pigno summit (Fig. 2.6) we will analyze different aspects of the rocks by walking along the ridge crest towards the north.

The Serra di Pigno is the largest continental sliver inherited from the passive margin and OCT of Corsica. The Pigno and associated alpine tectonic units (Fig. 2.7; Fig. 2.8) are characterized by a lawsonite blueschist-facies metamorphism (Law-Bs). They consist of metaophiolites, and continental gneissic rocks with associated metasediments (e.g. Faure & Malavieille 1981). *P-T* estimates indicate *P-T* conditions of about 0.6/0.8 GPa and 300±50°C in the Serra di Pigno unit and 1.0 GPa and 350°C in the overlying Campitello unit (Lahondère, 1996). Omphacite and garnet are not observed in metamafics, which are characterized by Na-amphibole + lawsonite + phengite mineral assemblages, instead.

Figure 2.7. Geological-structural map of the Serra di Pigno area (modified after Malavieille, 1981).

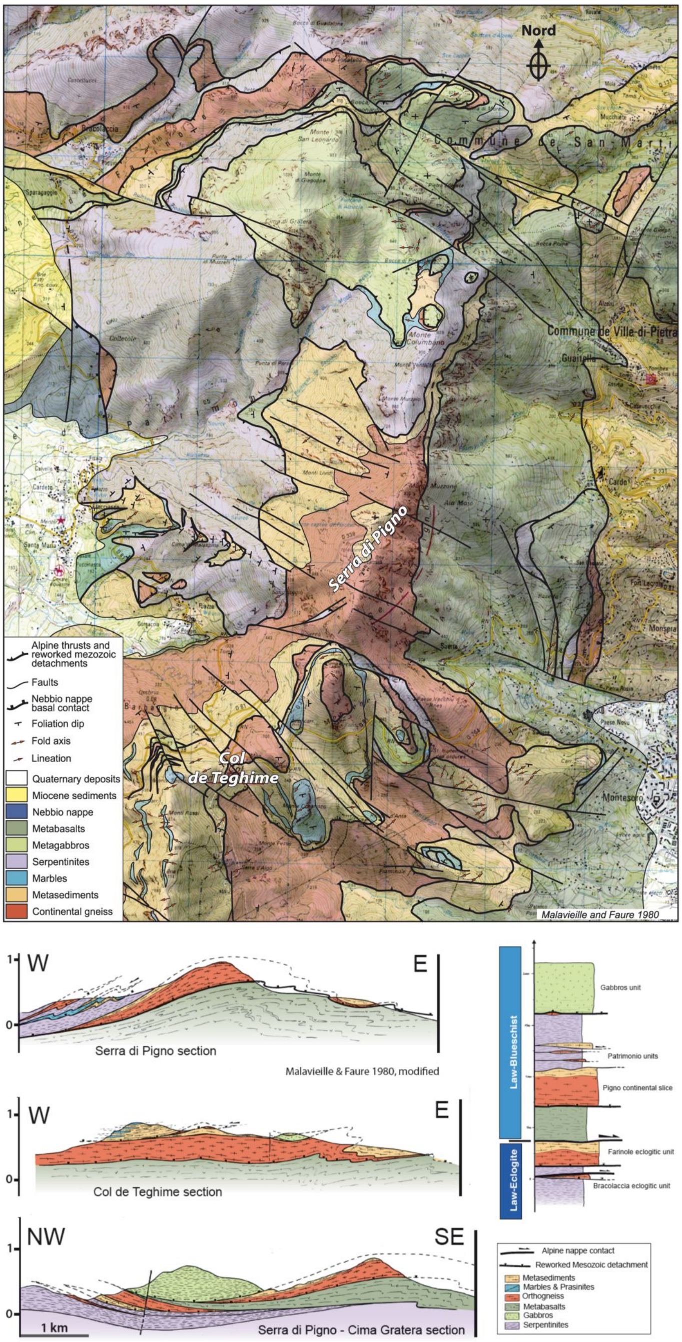
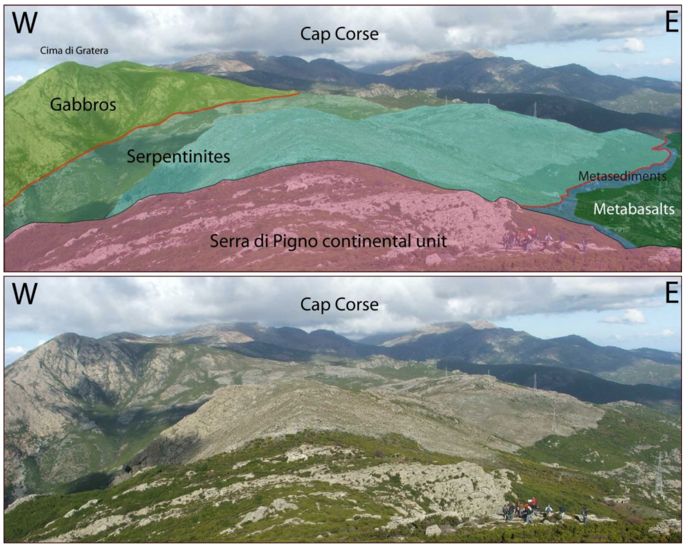


Figure 2.8. Panoramic view toward north of the Serra di Pigno with distribution of the main rock-types.



We will cross the different tectonic units along the crest going structurally from the bottom to the top. The Serra di Pigno continental unit rest structurally above a thick series of metabasalts. A major tectonic contact outlined by a thin layer of strongly deformed metasediments bounds the two units,. Walking to the North, we cross the contact between the Pigno gneissic rocks and ophiolitic nappes resting on top of it. Above the thin sequence of metasediments that represents the Mesozoic cover of the continental slice, a thick unit of strongly deformed serpentinites can be observed. On top of it, ophicalcite and metasediments including marbles and metacherts outline a major contact that could represent a Mesozoic intraoceanic detachment, which accommodated the exhumation of mantle rocks. Walking around, we can observe the relationships between gabbro dikes and serpentinites close to the contact and the effects of the strong alpine shearing deformation. The section will end in the upper unit, which consists ofa large flaser gabbro body. The Cima di Gratera gabbro is a cumulate-layered to varied-textured gabbro that is only partly re-equilibrated at blueschist facies conditions, except in shear zones (Fournier *et al.*, 1991) where the transformation has run to completion. Intense alpine strain affects these rocks, but parts of the gabbros that were incorporated into the subduction complex escaped deformation and recrystallization during both subduction and the subsequent exhumation.

If time allows it, we will try to observe the pseudotachylyte veins exposed around the peak of Cima di Gratera. The pseudotachylytes are located in the best preserved parts of the gabbro. These interesting rocks have been studied here by Austrheim & Andersen (2004) in gabbros and mantle peridotites and interpreted as pseudotachylytes formed by frictional melting on faults at seismic strain rates (related to subduction zone earthquakes).

Further to the North, in the Farinole area, the Serra di Pigno blueschist unit overlies a lawsonite eclogite-facies, metaophiolite-rich domain, which is the lowermost tectonic unit of this part of Cap Corse. This unit also includes slivers of continental basement rocks. Ophiolitic and continental metamafic rocks are characterized by variably preserved omphacite + Na-amphibole + Ca-amphibole + lawsonite + garnet + phengite mineral assemblages. This High Pressure nappe unit is relatively continuous from north to south of Alpine Corsica. *P-T* estimates for this unit are locally well defined, whereas modern studies of the microstructural and petrological evolution are lacking in other places. Estimates range from 1.5 GPa and 500 ± 50°C in the Farinole area (Lahondère, 1996), to 0.8 GPa and 300°C in the Sant'Andrea di Cotone area (Caron *et al.*, 1981) and 2.2 – 2.6 GPa and 520 ± 20°C in the San Petrone area (see Field trip 1, Vitale Brovarone *et al.*, 2011).

- Back to the road, walking down from the pass, we will observe the structural features of the Pigno continental slice, including the pre-Alpine magmatic rocks and structures reworked by alpine deformation. Afterwards, the cover rocks preserved on top of the continental sliver will be observed along the road. They reflect the pre-compressional sedimentary environment that was associated with such extensional allochthons. All these rocks are strongly deformed during continental subduction as shown by the sheath folds obseved in marbles of the metasedimentary sequence, which are associated with spectacular stretching lineations parallel to the fold axes.

Stop 1.2

Locality: Poubelle de Bastia UTM 32T 533823 E 4727217.

Themes: subduction-related deformation and metamorphism in mylonitic orthogneiss and panoramic view toward the east

This short stop will allow to observe the typical orthogneisses of the Pigno unit. Ductile deformation and shear sense indicators developed under H*P*/L*T* metamorphic conditions during continental subduction can be observed.

About one hundred meters above, after a short walk to the hill summit, metasedimentary rocks show spectacular structures. Beautiful sheath folds with eye-like structures are present in the impure marbles in the metasediments that cover the orthogneisses. A huge pebble (olistolith) of Triassic dolomitic limestone is included in the marble and wrapped by the foliation. Such early detrital input of exotic rocks in the sediments deposited close to the stretched continental margin is controlled by extensional tectonic processes and suggests the presence of a rough submarine topography, characterized by fault scarps.

A panoramic view towards the East will allow to discuss the role of Corsica in the tectonic framework of the northern Tyrrhenian sea.

Stop 1.3 (this stop will be cancelled for the large number of partecipants)

Locality: Cima Malaspina (UTM 32T 530789 E 4727332)

Themes: Metasediments/serpentinites contact

Slivers of a metasedimentary sequence ranging from marble to carbonate quartzite in direct contact and refolded (Fig. 2.9) with a large serpentinite body is exposed above the village of Patrimonio, along the Malaspina ridge (Fig. 2.7), which leads to the Serra di Pigno mountain. Ultramafics and metasediments overlie the H*P*/L*T* Farinole/Serra di Pigno gneissic units, which themselves lie structurally above H*P*/L*T* meta-ophiolitic units. Although being strongly deformed during alpine tectonics, the marble/serpentinite contact is concordant with the sedimentary bedding, as judged from lithological heterogeneities, and is marked by a continuous, centimetre-thick, weathering-resistant reaction rim of calcsilicates. The tectono-stratigraphic meaning of this contact will be discussed as it may represent an original and inherited ocean-continent transition or alternatively an Alpine tectonic contact.

Figure 2.9. Outcrop photograph of the contact between serpentinites and sediments at Malaspina.

Insert: alteration nicely reveals the zonation in the metasediments with a hard calc-silicates rim at the interface with serpentinites, a grey depressed zone with pure wollastonite, a dark zone enriched in CM and the bulk marble with calcite + quartz above.

The interface between serpentinites and sediments is nicely underlined by a jade-like cm-thick layer of diopside with minor garnet (andradite/garnet). In the sediments, the following mineralogical zoning is observed starting from the serpentinites:

A whitish 1 to 5 cm thick zone composed almost exclusively of wollastonite with minor grossular.

A 5 to 20 cm thick dark zone consisting of wollastonite + quartz + graphitic carbon (± grossular ± diopside), with no carbonate.

The bulk calcite+quartz bearing sediment, which is wollastonite free.

The transition between the two latter zones (wollastonite and carbon bearing vs. wollastonite free and low carbon content) is sharp.

Altogether, the association of wollastonite with local enrichment in carbonaceous material (CM) is interpreted as due to the local reduction of the Ca-carbonate to form elemental carbon and wollastonite according to the reaction

$SiO_2 + CaCO_3 + H_2 = CaSiO_3 + C + H_2O$.

The occurrence of this reaction is supported by a detailed geochemical and structural study of the carbonaceous material (CM) along the reaction front. The total organic carbon (TOC) significantly and sharply increases through the reaction front from about 0.5% wt in the original sediment to over 5% wt. in the reaction zone. Raman spectroscopy shows that CM is much more graphitic in the reaction zone than in the original rock. In addition, marked isotopic differences are observed on both sides of the reaction front with δ13C (CM) and δ13C (calcite) around -19‰ and 1‰ respectively in the original rock far from the reaction zone, whereas δ13C (CM) is around -2 ‰ in the reaction zone. These data are compatible with the formation of graphitic CM from the reduction of calcite, which may be a consequence of the diffusion of reducing fluids (H_2 and CH_4) from the underlying serpentinites.

This reaction front is locally associated with some spectacular mineralogical features that will be described in the field, such as the presence of aragonite-garnet intergrowths in the sediments close to the contact with serpentinites.

Field trip 2: second day

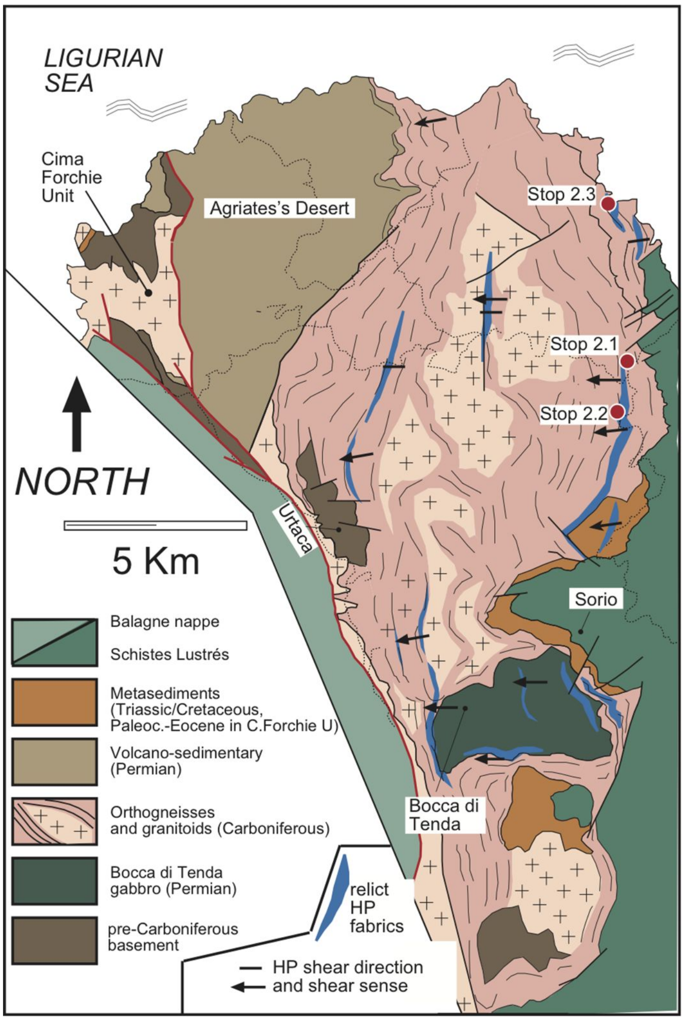
Deformation and metamorphism of the Tenda Massif and sourrounding Areas

Intinerary: S.Florent, Cava Fontana Morello, F.Poragghia, Fornali, Calaverte, S.Florent

The Tenda Massif

The Tenda Massif forms an elongated culmination of continental-derived metamorphic rocks interposed between the western “authochtonous” and “parauthochtonous” units (plus the Upper nappes of the Balagne) and the eastern “Schistes Lustres” composite nappe system. The Tenda Massif, exposed for nearly 200km^2 from S. Florent-Ile Rousse southward to Ponte Leccia (Fig. 2.10) is formed by two major tectonic units: the Cima delle Forchie unit, exposed in the north-west side of the massif, and the Tenda unit (Nardi *et al.*, 1978; Durand Delga 1978; Rossi *et al.*, 2001).

Figure 2.10. Geological map of the Tenda massif



Geological map of the Tenda massif with location of relicts of HO fabrcs. Map modified after Molli and Tribuzio (2004) and based on: Delcey and Meunier (1966); Nardi *et al.* (1978); Jourdan (1988); Dallan *et al.* (1983); Rossi *et al.* (1994); Dallan *et al.* (1995); Rossi *et al.* (2001).

The latter is characterized by Palaeozoic granitoids (mainly amphibole/biotite granodiorites and leuco-monzogranites), which are 280-300 My old (Rossi *et al.*, 1994), intruded in the southern part by a gabbroic complex (the Bocca di Tenda gabbro). This gabbroic complex shows well preserved magmatic features and is dated at 274±4 My (Ohnestetter & Rossi 1985). It consists mainly of olivin-gabbronorites, gabbronorites and horneblende-bearing diorites/tonalites (Ohnestetter & Rossi 1985; Rossi *et al.*, 2001; Tribuzio *et al.* 2001). Granitoids and gabbros are crosscut by doleritic dykes and peralkaline rhyolites both showing chilled margins against the host rocks (Tribuzio *et al.* 2001).

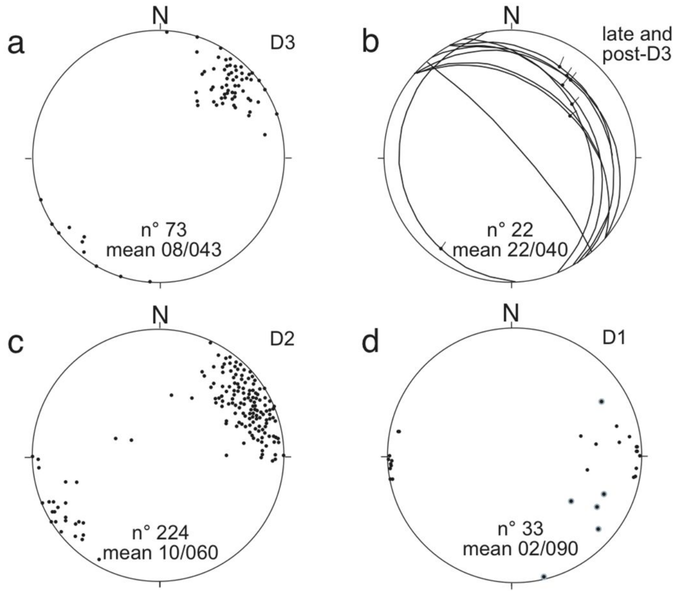
Remnants of a pre-Carboniferous basement are locally described in the southern and in the northern-western side of the massif (Delcey & Meunier 1966; Nardi *et al.* 1975; Jourdan 1988; Rossi *et al.* 1999; Rossi *et al.*, 1994; Rossi *et al.*, 2001). The basement rocks are associated with a Permian volcano-sedimentary cover well exposed in the northern part of the massif (Agriates's Desert). A reduced Mesozoic metasedimentary cover (Durand Delga 1984; Mattauer *et al.* 1981; Rossi *et al.*, 2001) crops out in the eastern and southern parts of the massif. These sediments are mainly represented by quarzites and micaceous arkoses (Trias), dolomitic marble and marble (Jurassic) and polygenic metaconglomerates principally formed by basement-derived debris with subordinate carbonatic clasts. These conglomerates are considered as Cretaceous or Cretaceous to Eocene in age (Durand Delga, 1984; Rossi *et al.*, 2001 and references therein).

The Tenda Massif (Fig. 2 and Fig. 2.10) forms a large-scale antiform with an amplitude of ~20km in the north, between S.Florent and Ostriconi, and less than 5km at the southern periclinal termination north of Ponte Leccia. The geometry of this elongated dome is partially related to the activity of a major system of wrench faults (mainly post-Eocene and pre-Burdigalian in age) described at the western boundary of the Tenda Massif (Central Corsica Fault zone of Maluski *et al.* 1973; Waters, 1990; Rossi *et al.*, 2001; Molli, 2008) and to a late phase of antiformal stacking occurring during the Late Eocene.

Structures related to a polyphase deformation history can be observed in cover rocks of the Tenda unit, where at least two generations of isoclinal folds are overprinted by a late phase of open to kink folds. On the contrary, basement rocks commonly exhibit a simpler structural pattern. More specifically, a regional-scale main foliation wrapping around undeformed domains (meter to kilometer in scale e.g. Casta granodiorite of Rossi *et al.*, 1994), can be recognized in granitoids. The main foliation is generally characterized by greenschist facies assemblages, although relict domains preserving pre-greenschist structures can be found (see below). On the basis of shear directions, overprinting relationships and petrologic data, the structures formed in greenschist facies metamorphic conditions have been subdivided into two groups, interpreted as developed at different structural levels (Molli and Tribuzio, 2004; Molli *et al.*, 2006).

The younger structures (D3) are characterised by localized zones of deformation from centimeter to meter in size, forming shear band systems associated with northeast/southwest trending stretching and mineral lineations (Fig. 2.11) and top-to-northeast shear sense indicators. These structures can be easily recognized toward the eastern border of the Tenda Massif, where they were firstly described by Waters (1990), Jolivet *et al.* (1990) and later on by Daniel *et al.* (1996), Egger and Pinaud, (1998) and Guyedan *et al.* (2003). D3 structures are also present inside the massif, where conjugate shear zones were described by Molli and Tribuzio (2004) and interpreted as related to partitioned coaxial strain and vertical shortening far from the eastern border of the Tenda Massif.

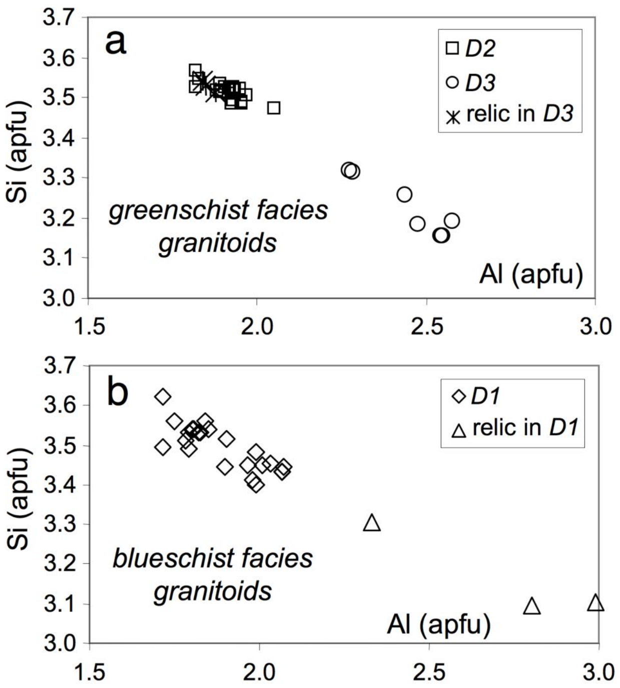
Figure 2.11. Equal area projection plots



Equal area projection plots (lower hemisphere) of: (a) stretching/mineral lineation of D3; (b) late- and post-D3, great circles for shear bands and faults with sense of movement and slicklines; (c) D2 and (d) D1 stretching/mineral lineations. In (d) black dots and mean trend refer to lineation in D1domains unaffected by D2/D3 overprint, whereas black dots with grey rims refer to D1 lineation in relicts domains within D2 or D3 dominant fabrics. After Molli *et al.* (2006).

D3 fabrics in granitoids are associated with celadonite-poor phengite (Si = 3.2-3.3 apfu, Fig. 2.12) + epidote + albite + quartz (± calcite ± chlorite) assemblages. The dolerite dykes developed chlorite + albite + epidote (± quartz ± calcite ± phengite). Metabasites are thus characterized by the absence of amphibole and can be related to the "lowermost greenschist facies" (Spear, 1993). These mineral assemblages indicate that the D3 deformation occurred at temperatures of 300-400°C and pressure conditions lower than 0.5 GPa.

Figure 2.12. Phengite compositions



Phengite compositions from metagranitoids deformed under greenschist (a); and blueschist (b) facies conditions; Si vs. Al tot (atoms per formula unit, apfu). D1, D2 and D3 = phengites related to D1, D2 and D3 fabrics, respectively.

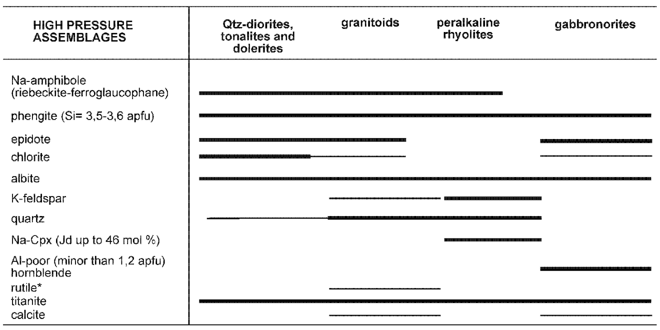
Meter- to pluridecameter D3 open to close folds associated with sub-horizontal to moderately dipping crenulation cleavage can be locally observed. They commonly show fold axis sub-parallel to the stretching lineation of the surrounding D3 shear zones. Semi-brittle shear bands and cataclasite-bearing fault zones (late- and post-D3 structures) geometrically and kinematically coherent with D3 structures are locally present (Fig. 2.11). They are interpreted as related to a progressive deformation at lower temperature conditions in the uppermost crust. D3 structures therefore evolved toward shallower structural levels with late- and post-D3 structures.

The main fabric along the eastern border of the Tenda Massif (as well as inside of it) can be referred to the D2 stage of greenschist facies deformation. This fabric shows stretching and mineral lineations (Fig. 2.11) oriented east-northeast/west-southwest (mean attitude toward 60°). Shear sense indicators such as asymmetric porphyroclasts and shear band systems of C- and C'-types show both top-to-west (southwest) and predominant top-to-east (northeast) kinematics (Fig. 2.11). D2 structures in granitoids are associated with celadonite-rich phengite (Si ~3.5 apfu, Fig. 2.12) + epidote + albite + quartz (± calcite ± chlorite ± actinolite) assemblages.

The heterogeneous deformation pattern that characterizes the Tenda unit allowed us to recognize and analyse domains of mylonitic orthogneisses and mylonitized doleritic and peralkaline rhyolitic dykes with a well preserved H*P*/L*T* fabric (D1 structures, Fig. 2.11) unaffected by greenschist facies static and/or dynamic retrogression.

The occurrence in the Tenda unit of lithotypes showing large compositional variations, allowed to estimate the peak metamorphic conditions (Fig. 2.13; 2.14, more details in Tribuzio and Giacomini 2002 see also Maggi *et al.*, 2011 and stop 2.1). Evolved rocks of the gabbroic sequence (quartz-diorite/tonalites), basalt doleritic dykes and granitoids are characterized by epidote-blueschist assemblages, as they show the coexistence of riebeckite/ferroglaucophane, epidote, celadonite-rich phengite (Si= 3.5-3.6 apfu) and albite. The peralkaline rhyolites develop a metamorphic paragenesis, defined by jadeite-bearing (up to 46 mol%) aegirine, riebeckite, celadonite-rich phegites (Si= 3.5-3.6 apfu), quartz, albite and K-feldspar. The Mg-rich lithotypes (olivine gabbronorites to gabbronorites) are characterized by the absence of blue amphibole. In particular, we have found that the gabbronorites develop a mineral association that can be related to the epidote-amphibolite facies, as they display the coexistence of Al-poor horneblende (Al_2O_3 < 6.6 wt%), albite, epidote and phengite. These multisystem assemblages allow the peak *P*-*T* metamorphic conditions to be constrained at 1.0 ± 0.1 GPa and 450 ± 50°C (Tribuzio & Giacomini, 2002; Molli & Tribuzio, 2004). These values attest to a geothermal gradient (d*T*/d*P*) of 10/13°C km^{-1}, thus suggesting a subduction-related tectonic setting (Tribuzio & Giacomini 2002 Molli & Tribuzio, 2004; Molli *et al.*, 2006).

Figure 2.13. H*P*/L*T* mineral assemblages



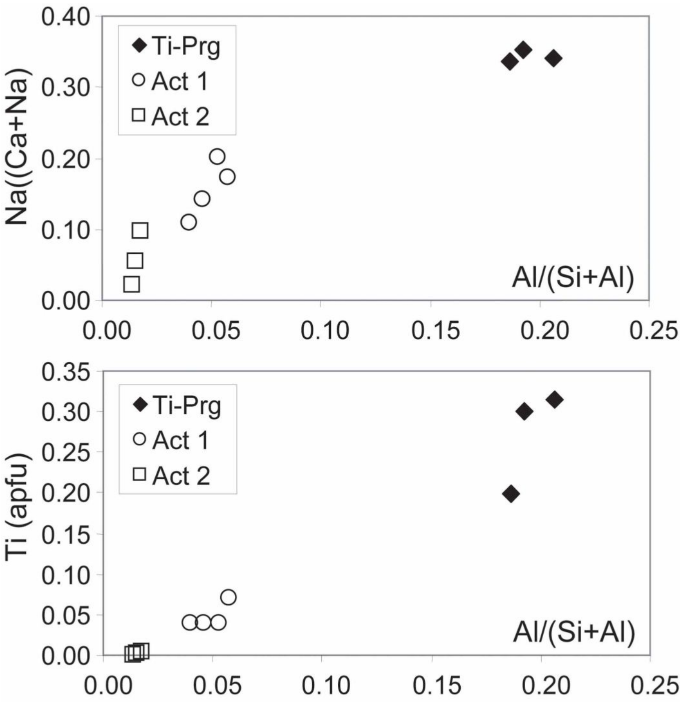
Significant H*P*/L*T* mineral assemblages in different rock types in the Tenda unit (thin lines represent mineral that are only locally present) (modified after Molli and Tribuzio, 2004).

Moreover, the D1 relict structural domains are characterized by E-W oriented stretching and/or mineral lineations defined by Na-amphibole and are associated with a top-to-the-west shear sense. These kinematic indicators, observed at the kilometric distance, for their constant attitude at regionale scale were considered relevant for the deep deformation and history of the Tenda unit, which has to be related with a east-dipping subduction (e.g. Mattauer *et al.*, 1981; Cohen *et al.*, 1981; Gibbson & Horack, 1984; Warburtoun, 1986; Waters, 1990; Malavieille *et al.*, 1998; Marroni & Pandolfi, 2003; Molli & Malavieille, 2010 and reference therein).

At its eastern border the Tenda unit is in contact with a typical ophiolitic unit, consisting of serpentinites, metabasites and a metasedimentary cover. The latter shows significant variations from the north, where the metabasites are stratigraphically followed by quartzitic-micaschists and calcschists, to the south, near S. Pietro di Tenda, at the south-eastern edge of the Tenda massif where pure quarzites, marbles and calcschists crop out.

The peak metamorphic assemblages in the metabasites derived from lava flows are characterized by Na-amphibole + albite + chlorite + epidote + phengite (± calcite). Therefore, the metabasalts display typical epidote-blueschist facies (Evans, 1990) assemblages, similar to those observed in the metadolerites within the Tenda orthogneisses and in the metadiorites from the Bocca di Tenda gabbroic sequence (Tribuzio and Giacomini, 2002). In the ophiolite unit, blueschist facies metabasites derived from Fe-Ti-rich gabbroic protoliths have also been found, for instance at Monte S. Angelo (kilometric coordinate 566.9/249). These rocks commonly preserve mineral relicts of the igneous paragenesis (i.e. ilmenite and Ca-clinopyroxene), (Fig. 2.14) and display a peak metamorphic assemblage with Na-amphibole + aegirine + albite + chlorite. Ophiolitic Mg-Al-rich gabbros are also found and are characterised by the absence of Na-amphibole, similar to the gabbronorites from the Bocca di Tenda gabbroic sequence (Tribuzio and Giacomini, 2002; Molli and Tribuzio, 2004). Remarkably, the chemical composition of Na-amphiboles from blueschist facies metabasites of the Tenda Massif and from the ophiolitic unit are similar (Fig. 2.14), suggesting that both unitsunderwent similar peak metamorphic conditions. This conclusion has been confirmed recently with Raman Spectroscopy on Carbonaceous Material (Vitale Brovarone *et al.*, submitted).

Figure 2.14. Amphibole compositions from Tenda microgabbro



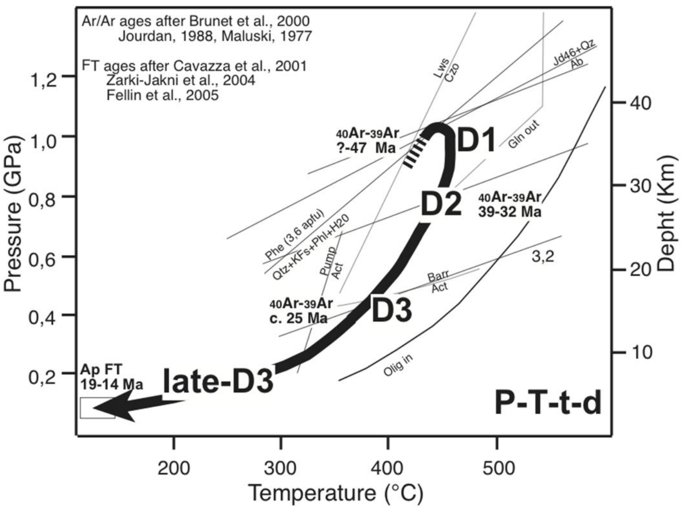
Amphibole compositions from Tenda microgabbro with greenschist facies metamorphism, involved in the D2 deformation event. Na/(Ca + Na) vs. Al/(Si + Al) and Ti vs. Al/(Si + Al) in apfu. Ti-Prg = Ti-pargasite of igneous origin; Act 1 = actinolite in coronas around Ti-Prg; Act 2 = actinolite in top-to-west shear band systems (D2 event).

Blueschist facies assemblages in metabasites are commonly partially overprinted by greenschist facies parageneses, with development of chlorite + albite + epidote

(± phengite ± calcite ± actinolite). A widespread re-equilibration under greenschist facies conditions is common in the ophiolitic metasediments. The ophiolitic unit is characterised by distributed deformation associated with pervasive folding from map to thin section scale. Two generations of isoclinal folds (D1 and D2) are associated with the two main stages of metamorphism (under blueschist and greenschist facies conditions, respectively), whereas later stages of folding are characterised by sub-horizontal to sub-vertical axial-planar disjunctive crenulation cleavage and kink bands (D3 and late-D3).

Figure 2.15 reports the *P-T-t-d* evolution of the units in contact along the eastern border of the Tenda Massif. Both oceanic and continental (Tenda) units are characterised by peak metamorphic conditions of ~450°C and 1.0 GPa, which were attained during D1 deformation. D2 deformation and structures can be related to the exhumation history, which started at H*P*/L*T* metamorphic conditions and evolved towards lower pressure, when the early greenschist facies assemblages formed. The following D3 structures formed at *T* = 300-400°C and P<0.5 GPa and were overprinted at shallow structural levels by the late- and post-D3 semibrittle and brittle structures. D3 and late- to post-D3 structures are localized and individual structures accommodated little displacement (from millimeters to meters). In addition, their impact on the overall crustal geometry is negligible, as suggested by the undisturbed D2 large scale folds at the contact between oceanic and continental units. Recently acquired structural and petrologic data (Maggi *et al.*, 2011 and Maggi, 2011) largely confirmed the structural history described above. Furthermore, a U/Pb TIMS study on syn-kinematic rutile in D1 phyllonites yelded an age of 48 ± 18 Ma (MSWD 7.3), while an intercept at 54 ± 8 Ma (MSWD = 48) was obtained from coexisting acmite-phengite and ox/sulph coatings.

Figure 2.15. Pressure-Temperature-deformation



Pressure-Temperature-deformation path of continental Tenda and ophiolitic units justapoxed across the eastern border of Tenda Massif after Molli *et al.* (2006). Gln out taken from Maresch (1977), the reaction pumpellyite + chlorite = actinolite + epidote after Liou *et al.* (1983), lawsonite-clinozoisite transition after Barnicoat and Fry (1986), the lower stability limit of barroisite from Ernst (1979), oligoclase-in reaction from Maruyama *et al.* (1983), the reaction curve for Na-clinopyroxene (Jd46) + quartz = albite was calculated with the 3.1 version of THERMOCALC program (Holland and Powell, 1998, and references therein). The reaction curve for Mg-phengite (Si = 3.6 apfu) = quartz + K-feldspar+ phlogopite + H_2O) is after Massonne and Szpurska (1997). Ar/Ar data after Brunet *et al.* (2000), Jourdan (1988) and Maluski (1977). Fission track apatite ages (Ap FT) after Cavazza *et al.*, 2001; Zarki- Janki *et al.* (2004) and Fellin *et al.* (2005). Note that new data of Maggi *et al.* 2011 suggest for the H*P*/L*T* event an age between 55 and 48 Ma.

Day Two Itinerary

Stop 2.1:

Locality: Fontana Morello Quarry (UTM 32T 522329 E 4722712)

Themes: Heterogeneous strain in the ETSZ, gneissic fabrics and the Acmite-Rutile-bearing phyllonites

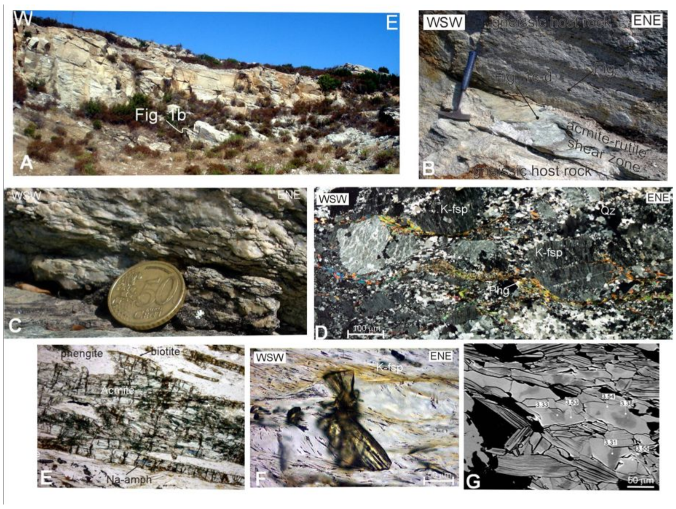
A panoramic view of the eastern margin of Tenda Massif will be shown to illustrate the heterogeneous strain in this areaThe analysis of the Fontana Morello Quarry will allow to describe the range of fabrics and mineral assemblages of the ETSZ More specifically, this stop aims at illustrating (i) the heterogeneous distribution of shear strain within the ETSZ, with zones of localized ductile-to-brittle shearing (SZ) that wrap around lens-shaped massive bodies with prominent gneissic fabric

(ML); (ii) the effects of the intense fluid-rock interaction along the ETSZ with associated changes in textures and mineralogy; and (iii) the high-pressure (Na-amphibole/acmite-rutile bearing) phyllonitic shear zones.

The NW-SE striking main foliation in the ML, which generally dips to the NE at a shallow angle, is concordant with that in the overlying Schistes Lustrés. The ML mineralogy consists of a high-variance assemblage made of quartz + phengitic muscovite + epidote and (relict) feldspar (Fe-oxides, zircon and allanite as accessory phases). The SZ mineralogy is invariably dominated by Si-rich (Si^{4+} = 3.5-3.7 apfu) phengite (40-60 vol%), modally abundant quartz (30-50 vol%), albite (10-20 vol%) epidote (5-10 vol%) and microcline (5-10 vol%). Locally, Na-amphibole (10-20 vol%) also occurs in the SZ assemblage to form thin (up to 1 meter thick) dark mylonitic levels. Collectively, due to the predominance of phengite in the shear zone assemblages (locally > 60%), the SZ consists of phyllonites. with stretching lineations striking WSW-ENE to E-W. quartz-phengite-albite characterize ML and whereas Na-amphibole-quartz-albite-phengite may be found in SZ. Kinematic indicators are abundant within the two main rock types and invariably point to top-to-the-W/NW sense of shear.

At about 100 meters below the main contact, a 10 cm thick phyllonitic layer (Fig. 2.16b) contains the high-pressure association rutile-acmite-phengite-albite-epidote-Na-amphibole. Transition from the gneissic host rocks to the shear zone is marked by a green alteration halo made of phengite aggregates (Fig. 2.16b). Kinematic indicators both at the meso- and micro-scale, systematically indicate top-to-the-west ductile shear sense (Figs. 2.16c-d). Acmite and rutile form porphyroblasts along the main foliation; acmite grains host inclusions of Na-amphibole and are rimmed by retrograde biotite (Figs. 2.16e-f). The modal composition consists of phengite (>60% vol.), acmite (25% vol.), rutile (5% vol.), quartz plus accessory phases (albite + microcline + epidote + relict zircon; 10% vol.). Microprobe data and BSE images indicate a rather homogeneous composition both for phengite (3.6-3.7 Si^{4+} apfu) and acmite (average composition $Jd_{17}Di_9Hed_{10}Px_{0.05}$) grains. On the other hand, phengite shows a strong core-to-rim chemical zoning (Si^{4+} content ranging from 3.2 to 3.6 apfu, respectively) (Fig. 2.16g).

Figure 2.16. Stop 2.1



(a) Panoramic view of the Stop 2.1, with indicated the acmite-rutile phyllonites. The dominant lithological type consists of massive gneisses with main foliation dipping shallowly towards the E; (b) The acmite-rutile phyllonite in the field; (c) Kinematic indicators in the host rocks (exposure normal to foliation and parallel to the stretching direction). Shear sense id dominantly provide by S-C fabrics and s-type porphyroclasts made of relict K-feldspar; (d) Kinematic indicators at the thin section scale (section cut normal to foliation and parallel to the stretching direction); (e) Porphyroblastic acmite crystals, hosting Na-amphibole as inclusion and rimmed by biotite; (f) Porphyroblastic rutile in the phyllonite matrix; (g) - BSE images showing the phengite chemical zoning (numbers refer to the measured Si^{4+} a.p.f.u).

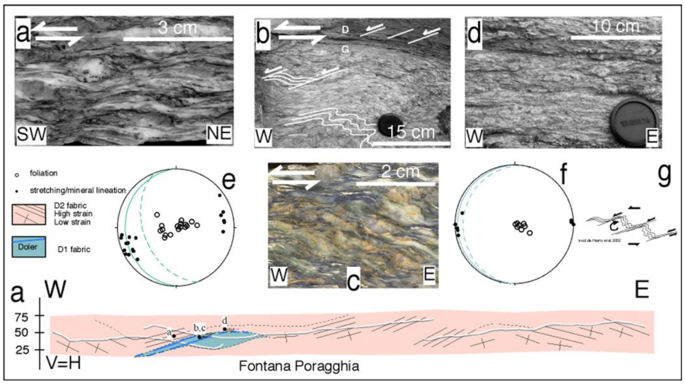
Stop 2.2

Locality: Fontana Poragghia (UTM 32T 521706E 4720892)

Themes: blueschist orthogneiss and overprinting fabrics.

Widespread relicts of pre-greenschist facies assemblages and structures (D1) can be observed all along the eastern border of the Tenda Massif as shown in Fig. 2.5. An easily accessible exposure will be visited along the D62 (kilometric coordinate 567.5/262.07), near the Fontana Porraghia site (Fig. 2.17). Here, orthogneiss and metadolerite dykes hosting H*P*/L*T* fabrics will be observed and the styles of GS structural overprinting analysed.

Figure 2.17. Fontana di Porragghia



(a) Cross-section of Fontana di Porragghia and surroundings showing D1 shear zone relicts and D2 foliation with gradient of D2 strain; equal area projection plots (lower hemisphere) of pole of foliations, great circles of shear bands and stretching lineations D2 (e) and for D1 (f) deformations; (a) D2 fabric with well developed top-to-southwest shear bands; (b) contact between mylonitized dolerites (D) with top-to-west shear bands and granitic mylonites (G) with top-to-west shear bands and centimeter-scale backfolds; (e) D2 crenulation with greenschist retrogression of former D1 fabric in the microlithons; (g) schematic diagram showing development of east-vergent folds between shear band systems associated with top-to-west shearing, on the basis of Harris *et al.* (2002).

In Fontana di Poragghia, the orthogneisses locally show a main foliation defined by Na-amphibole+ celadonite-rich phengite (Si ~3.5 apfu, Fig. 2.12) + epidote + albite + quartz (± chlorite ± K-feldspar). White mica cores locally preserve high-celadonite (Si = 3.1-3.3 apfu) and Ti-poor compositions, which can be attributed to a low-pressure stage of the prograde path.

The blueschist-facies fabric is characterized by mineral (Na-amphibole) and stretching (quartz aggregates) lineations striking east/west (mean toward 268°), with top-to-west kinematics, defined by σ-type asymmetric porphyroclast and widespread shear bands in metadolerites (Fig. 2.17). Centimeter to decimeter-scale E-vergent folds deforming the blueschist foliation can be observed in orthogneisses located inbetween mylonitic metadolerites. The folded Na-amphibole fabric is not retrogressed to greenschist-facies assemblages, thus allowing us to interpret the folds as related to the back-rotation of the blueschist foliation in a progressive history of top-west shearing. This event possibly occurred during the development of shear bands in metadolerites, in the latest stages of blueschist shearing. Fig. 2.17g illustrates fold/shear zone relationships as observed from decimeter to microscale and their interpretation according to the model described by Harris *et al.* (2002).

The blueschist foliation grades to orthogneisses characterised by greenschist facies assemblages, and the transition shows a spatial variability normal to the strike. Toward the west, the blueschist facies foliation coplanarly grades to a greenschist foliation characterised by a well-developed lineation of stretching and mineral type oriented toward 248° southwest (Fig. 2.17). In this domain, asymmetric porphyroclast systems and well-developed shear bands indicate top-to-the-southwest kinematics (Fig. 2.17). East of Fontana Porraghia, the transition to the greenschist foliation occurs through a decameter thick domain of strongly crenulated orthogneisses. This domain shows the partial to total breakdown of blueschist facies Na-amphibole (Fig. 2.17).

Stop 2.3:

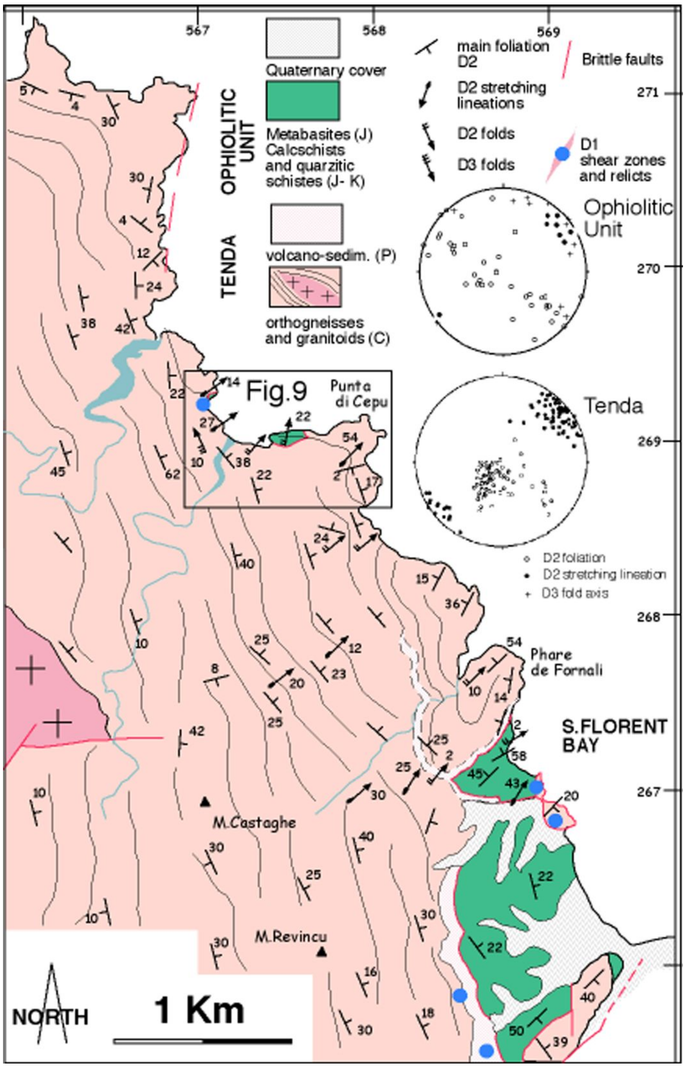
Locality: Walking along the coast from Fornali to Calaverte (UTM 32T 521788 E 4727588)

Themes: The ETSZ and the geometries of the contact between the Tenda and the ophiolitic unit.

This stop will involve a walk towards the north, along the coast from Fornali lighthouse to Calaverte bay. During this walk, it will be possible to observe structural geometries within the ETSZ and the relationship between the Tenda crystalline massif and the ophiolitic unit.

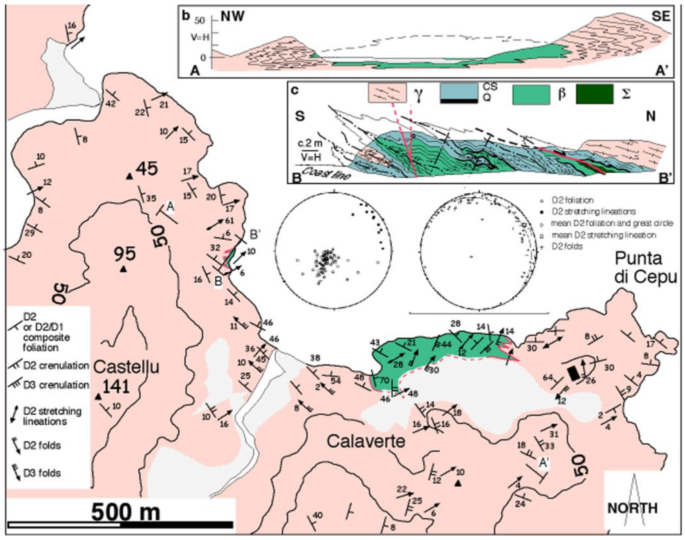
As shown in Fig. 2.18, between St.Florent and the Fornali lighthouse, the ophiolitic unit, consisting of metabasites and calcschists, are overlain by orthogneisses and a mylonitised volcano-sedimentary cover, both Tenda-derived. According to the interpretations of Waters (1990) and Dallan and Puccinelli (1995), the orthogneisses are not Tenda-related and form tectonically interleaved continental–derived slices similar to those found east of St.Florent (Unità degli ortogniess superiori di Dallan & Puccinelli, 1995; Fornali nappe of Waters, 1990). Rossi *et al.*, (1994) and (2001) proposed a similar interprerpretation, although the tectonic slices were condidered as Tenda-derived.

Figure 2.18. Geologic sketch map of the area west of S. Florent



Geologic sketch map of the area west of S. Florent, with inset of equal area projection plots (lower hemisphere) of pole of foliation, stretching/mineral lineation, and fold axes of D2 structures in the Tenda and ophiolitic units. Rectangle indicates the location of Fig. 2.19.

Figure 2.19. Geological map of the Calaverte bay



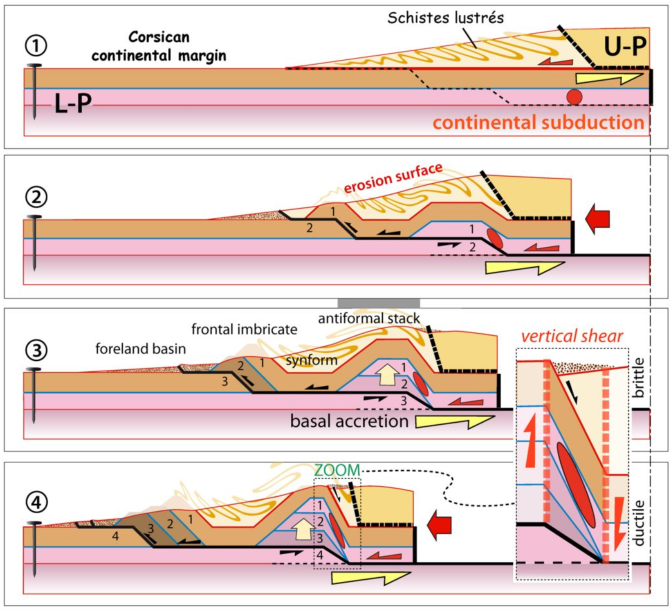
Geological map of the Calaverte bay and surroundings, with inset of equal area projection plots (lower hemisphere) of pole of foliation, stretching/mineral lineation, and fold axes of D2 structures; (b) geological cross section of the Calaverte area, traced normal to the trend of D2 stretching lineation, with the interpretation of the ophiolite unit exposure within the Tenda orthogneiss as lateral termination of D2 sheath-fold; (c) schematic cross-section of north-west exposure of ophiolitic unit along the coast with observable structures. γ are orthogneiss; CS and Q are calcschists and quarztites; β, metabasic rocks and Σ actinolite-tremolite schists.

The geological maps of Fig. 2.18, 2.19 describe in some details the geometry of the contact between the Tenda and the ophiolite unit west of S.Florent. Key exposures are located to the south of the Fornali lighthouse, between Fornali and Punta Cepu, where the ortogneiss and the mylonitized Tenda-derived cover are affected by pervasive isoclinal folds at different scales, withstrongly curved intersection lineations on foliation planes. The axial planar foliation of the isoclinal folds shows greenschist facies assemblages with phengite composition (Si = 3.4-3.5 apfu) similar to D2 structures in other parts of the Tenda.

Further north in the Calaverte bay (Fig. 2.19), the ophiolitic unit reappears with a complete litostratigraphic sequence, with actinolite-tremolite schists, metabasites, quartzitic-micaschists and calcschists forming tight folds within the orthogneiss (Fig. 2.19). Meso- and microstructural features show that the axial planar foliation of the folds is associated with greenschist assemblages deforming a previous blueschist fabric. These two foliations (and the associated isoclinal folds D1 and D2) are both deformed by a west-dipping disjunctive crenulation cleavage that is axial planar to metric scale east-vergent open to close folds (D3). Our interpretation for the exposure of the Calaverte Bay is that of the lateral terminations of a strongly non-cylindric recumbent fold (Fig. 2.19). The ophiolitic unit (calcschists and metabasites) is pinched within the orthogneisses, which therefore were more competent during deformation. These structures are considered as being due to the buckling of the original thrust contact (with the ophiolitic unit arranged in an inverted limb of kilometer-scale west vergent isoclinal D1 fold above the Tenda orthogneiss), followed by rotation and stretching of the earlier fold during southwest/northeast directed shearing.

Similar geometries of deformation can be also observed south of S.Pietro di Tenda, around the M.Buggentone area and further south, in the Ponte Leccia surroundings. Therefore, at the scale of the whole eastern border of the Tenda, the contact between the continental Tenda and the ophiolitic units appears tightly folded in large scale recumbent isoclinal non-cylindric D2 folds. This has important implication for the interpretation of the ETSZ which can hardly be interpreted as a core complex of Cordilleran type (Jolivet *et al.*, 1990; Daniel *et al.*, 1996; Gueydan *et al.*, 2003). On the contrary, the Eastern Tenda Fault Zone shows the main tectonic grain related with the deep seated tectonic processes associated with subduction and syn-collision intrawedge-deformation possibly developed during the underplating of more external units. Thus, although extension is commonly invoked to explain exhumation of metamorphic rocks and synchronous enigmatic zones of normal shearing observed in most mountain belts, in many cases, this cannot be the dominant mechanism. Since many years, uplift induced normal sense shear zones and concommitant brittle normal faults (developed at lower depth) are described in both ancient (e.g. the Variscan belt, Pérez-Estaùn *et al.*, 1991) or active mountain belts (e.g. the Taiwan belt, Crespi *et al.*, 1996). Interpretation of such deformation features in the frame of mountain building is still controversial today and proposed models range between end-members involving compressional (convergent) or extensional (divergent) settings. Analog experiments with décollements and erosion (e.g. Konstantinovskaia & Malavieille 2005; Malavieille, 2010) suggest an alternative way to develop crustal scale normal sense shear deformation during continental subduction. Such structures can be the result of the vertical shear induced by strain partitioning within the orogenic wedge (Fig. 2.20). Continuous uplift of underplated crustal units relative to comparatively stable surrounding rocks favors vertical shear and as a consequence a strong stretching and thinning of the formerly stacked tectonic units. At depth these domains are characterized by the development of foliation zones of combined pure and simple shear deformation with a normal sense shearing component. They can evolve to brittle normal faults superimposed on former ductile foliation when reaching upper-crustal domains during synconvergence erosion assisted uplift.

Figure 2.20. Cartoon after Malavieille (2010)



Cartoon after Malavieille (2010) showing deformation partitioning and kinematics of units in a décollement type wedge. Notice the deformation of upper plate (orogenic lid) resting on top of the former refolded décollement. Early folds are suggested to show evolution of U-P geometry. A possible deformation mechanism responsible for vertical shear inducing stretching and thinning of the underplated units is schematized. Red ellipsoids show resulting strain. U-P = pre-structured upper-plate, L-P = basement lower-plate.

Remarkably, across the ETSZ, there is no significant difference in geochronological ages across the eastern border of the Tenda Massif, between footwall and hanging wall of the supposed detachment fault, for both high ($^{40}Ar/^{39}Ar$ on white micas) and low (apatite fission track) temperature systems (Cavazza *et al.*, 2001; Zarki-Janki *et* *al.*, 2004; Fellin *et al.*, 2005). This agrees with the structural data observable, which indicate a weak (low displacement) extensional-related reactivation of previous oldest and deepest fabric. In addition, the presence of clasts of Tenda-derived granitic mylonites within the lowermost part of the S.Florent sequence (see also Dallan and Puccinelli, 1995; Ferrandini *et al.*, 1998) argues against the interpretation of the Miocene of S.Florent as a supra-detachment extensional basin related to the activity of the East Tenda shear zone at depth, as proposed by Jolivet *et al.* (1990).

Field trip 2: third day

The External Zones of Alpine corsica in the Balagne Region.

Intineary: Saint Florent, Colle di U'Vezzu, Ostriconi, Belgodere, San Sebastiano Chapel, San Colombano pass,km 59, Francardo, Bastia

The Balagne region is located in the northern part of Corsica between Calvi to the west and the Tenda Massif to the east (Fig. 2.21, 2.22a,b). The external and uppermost structural units of the Alpine Corsica orogen crop out in the easternmost part of the Balagne, between Ostriconi and Lozari in the north and Moltifao and Pietralba in the south, . These units are preserved within a north-south trending kilometer-scale synformal structure locally controlled by transtensional strike-slip faults. Although they were probably originally present also further to the west, they are largely eroded today in the so-called "Hercynian Corsica."

Figure 2.21. Schematic field drawings of D2 deformation structures along the ETSZ.

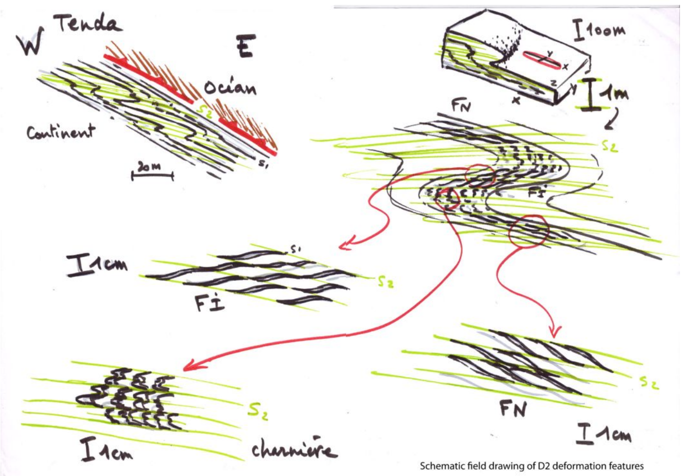
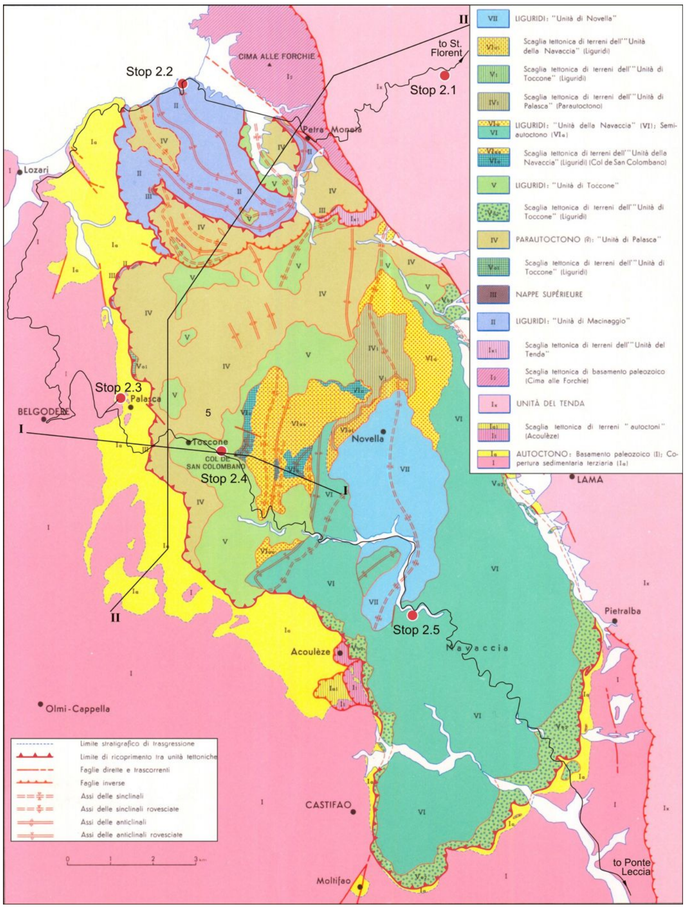
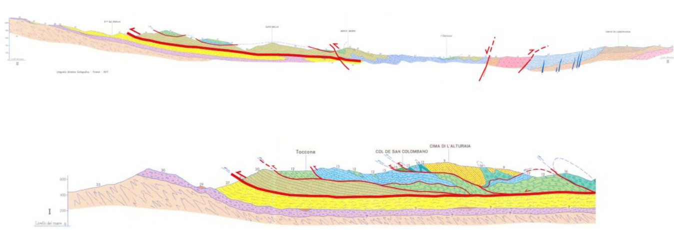


Figure 2.22. The Balagne region



Tectonic scheme of the Balagne region with stops and trace of cross-section of Fig. 2.21 after Nardi *et al.* (1978) modified.



Geological cross-section of the Balagne region after Nardi *et al.* (1978) modified.

From bottom to top the following units can be recognized (Fig. 2.22a, 2.22b):

The "autochthonous" Hercynian basement.

It is represented by the high-grade metamorphic rocks of the Belgodere complex, which comprises leptynite–amphibolite associations with eclogite boudins, orthogneiss and metasediments derived from an Early Palaeozoic protolith affected by early-Devonian high grade metamorphism (Palagi *et al.*, 1984; Menot and Orsini, 1990; Rossi *et al.*, 2001; Rossi *et al.*, 2009). The Belgodere complex was intruded at c. 340 Ma (Rossi *et al.*, 1988; Menot and Orsini, 1990) by Mg-K plutons, mainly consisting of monzonites and associated ultrapotassic mafic rocks. During the emplacement of Mg-K intrusion, the Belgodere gneiss underwent anatexis under amphibolite conditions (Rossi et al 1991). A second composite magmatic association is testified by granitic and mafic intrusions dated at 300-280 Ma (Rossi *et al.*, 2001; Tribuzio *et al.*, 2008), by a volcano-sedimentary sequence including calc-alkaline ignimbrites and volcanoclastites and dykes swarms (Rossi *et al.*, 2001). A third Permian magmatic cycle is associated with the alkaline volcano-plutonic magmatism (Menot and Orsini, 1990; Rossi *et al.*, 2001; Tribuzio *et al.*, 2008; Rossi *et al.*, 2009) e.g. the Monte Cinto caldera and the metaluminous intrusion of Popolasca. The magmatic history of the basement ends with a later set of mafic dykes (Rossi *et al.*, 2001).

The Hercynian “autochthonous” basement is covered by Mesozoic-Tertiary sediments, all characterized by the incomplete and rather thin sequences, with several unconformities (Durand Delga, 1976; Nardi *et al.*, 1978; Rossi *et al.*, 2001; Durand Delga *et al.*, 2001). In the area of interest, Early? -Mid Triassic reddish to green conglomerates, pelites and sandstones (Petra Moneta Conglomerates), lie alternatively on the Hercynian basement and on the Permian volcano-sedimentary sequences. The cover sequence is completed by massive conglomerates characterized by basement-derived clasts (high and low grade metamorphic rocks, granites, volcanites) having a maximun thickness of c. 250m. Within it, conglomerate-matrix nummulites of Paleocene-Early Eocene age have been locally found (Amaudric du Chaffaut, 1973; Ferrandini *et al.*, 2010). The Conglomerates grade to discontinuous levels of grey limestones and calcarenites (“Nummulitic limestones”) roughly stratified and characterized by Mid-Late Lutetian nummulites (Nardi *et al.*, 1978; Ferrandini *et al.*, 2010), in turn followed by grey sandstone, siltite, microconglomerates and pelites (Lozari sandstone) well exposed along the R.N. 199. Sandstone levels locally include, at their base, Upper Lutetian nummulites. The sequence ends with black shales, siltstone and fine grained sandstones forming the so called “Shaly Flysch” (“Flysch noir”, “Flysch Argilloso“), with a maximum thickness of c.250 m. The “Shaly Flysch” can directly overlie the basal conglomerates, the “Nummulitic limestones” or the Lozari sandstones. The maximum thickness of the “autochtonous” sedimentary cover is around 500 m. Fragments and/or slices of Mesozoic limestone, classically considered as small to large (decameter in size) olistoliths (Nardi *et al.*, 1978; Durand Delga, 1978; Rossi *et al.*, 2001), are included in the upper part of the Middle Eocene sequence near Pietralba).

The autochthonous basement and cover sequence were affected by polyphase deformation (Egal & Caron, 1989; Egal, 1992), which occurred at shallow crustal levels, at $T < 280/300°C$. The sedimentary sequence is characterized by heterogeneous deformation localized along bedding-parallel fault zones and distributed within folded domains. West vergent north-south trending parallel folds and top-to-the west fault zones, usually overprinted by late folds and faults, are common deformation features (Egal, 1992).

Parautochthonous or External continental units

Between the “autochthonous” basement and the overlying units belonging to the “Nappe Supérieure”, slices of a “parautchtonous” basement and cover sequences can be locally observed. These can be considered as the northward equivalent of tectonic units exposed along the western border of the “Hercynian Corsica” between Vecchio in the south and Popolasca-Ponte Leccia in the north, which are labelled “External continental units” (Fig.6; Bezert, 1990, Bezert and Caby, 1989; Molli, 2008). These units, which have also been studied near Corte (“Corte slices” of Amaudric du Chaffaut *et al.*, 1973), are characterized by basement rocks (mainly Permian granites and Paleozoic host rocks) and a metasedimentary cover consisting of Mesozoic carbonates (Corte and Restonica marbles) including Late Jurassic and Late Cretaceous metabreccias. The sedimentary sequence ends with siliciclastic turbidites (metasandstones) dated to early Mid-Eocene on the basis of the presence of Nummulites biarritzensis; Discocyclina sp., (Bezert & Caby 1988). In the Balagne area very limited exposures of these units are reported. The Volparone breccias (Late Cretaceous? breccia which includes clasts of Malm limestone and rift-related metabasalts) in the Bocca di Fuata area are an example (Malasoma and Marroni, 2007).

The "Parautochthonous" or “External units” are characterized (Bezert & Caby 1988) by a low grade high pressure/low temperature metamorphic overprint (H*P* greenschist facies of Bousquet *et al.*, 2008) recently constrained at ~0.7 GPa for a temperature of 325°C (Malasoma *et al.* 2006; Malasoma and Marroni, 2007). The high pressure/low temperature parageneses are mainly observable in suitable rock types, e.g. within the chlorite-phengite matrix of the metabreccias and in metapelites as well as within the mafic dykes. In the granitoids, millimetre-scale blue-amphibole bearing shear zones have been observed (Molli *et al.* 2005 and unpublished). These shear zones show evidence of east–west transport associated with top-to-the-west kinematics, similarly to structures described elsewhere all along the boundary between Alpine and Hercynian Corsica by Bézért (1990).

The Nappe Supérieure System

The "Nappe Supérieure" System includes two major tectonic units: the Nappe du Bas Ostriconi and the Balagne Nappe.

The Nappe du Bas Ostriconi

It is characterized by a Late Cretaceous to early Tertiary sedimentary cover detached from the original basement, inferred to be of continental -type by Rossi *et al.* (2001). It is is formed by a calcareous-marly sequence associated with siliciclastic deposits including microconglomerates and sandstones supplied by a basement consisting of continental crust. The mainly carbonate flysch (named in local literature as Calcareous Flysch, Narbinco Flysch or Ostriconi Flysch) is dated to Late Cretaceous and correlated with the Calcareous Flysch of the Nebbio Nappe (Nardi, 1968; Durand Delga, 1976; Durand Delga, 1984) and the Macinaggio flysch in eastern Cap Corse. Some authors (Durand Delga, 1984, Dallan and Nardi, 1984; Caron, 1990; Rossi *et al.* 2001), proposed a correlation between these Late Cretaceous deposits and those observable in the Santa Lucia Nappe west of Corte i.e. the Tralonca Flysch and the "Anchesa Limestone." Moreover, since the end of the sixties, analogies between the Nappe du Bas Ostriconi and the Helminthoid Flysch of the Ligurian Alps and northern Apennines has been suggested (Nardi, 1968; Durand Delga, 1976; Durand Delga, 1984).

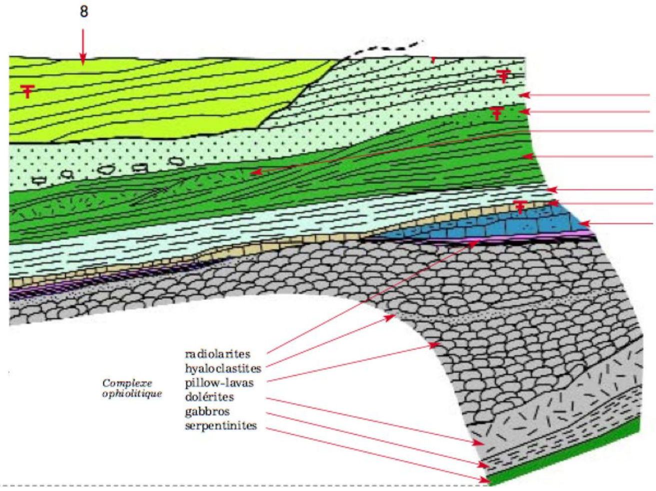
The Late Cretaceous sequence is completed by the continental crust-derived conglomerates and coarse grained sandstones (Paleocene?-early Eocene?) of Punta d'Arco (Nardi *et al.*, 1978; Rossi *et al.*, 2001). Some authors (see Rossi *et al.*, 2001) include in the Nappe du Bas Ostriconi Palombini shales (S.Martino fm.) and Lydiennes Flysch (see below), locally found at the base of the Nappe as sheared lenses.

The Nappe du Bas Ostriconi shows internal structures represented by NW-SE striking and west-vergent folds associated with an east-dipping cleavage and with top-to-the west thrusts. These structures were formed at shallow crustal levels, at *T* <280°C (Vitale Brovarone 2011; Vitale Brovarone *et al.*, submitted). It directly overlies the Eocene cover of the "autochthonous" basement in the north whereas it overthrust the Balagne Nappe system toward the South (Nardi *et al.*, 1978; Rossi *et al.*, 2001). This late thrust is considered as a splay of the Ostriconi sinistral strike-slip fault system.

The Balagne Nappe

It includes Jurassic ophiolites, a supra-ophiolitic Late Jurassic to Late Cretaceous sedimentary sequence and Paleocene-early Eocene siliciclastic deposits (Fig. 2.23). The ophiolitic sequence comprises Jurassic basalts, mainly pillow lavas lying on gabbros and serpentinized mantle peridotites. U-Pb (SHRIMP) data on zircon from a trondhjemite vein intruded in a gabbro indicates late magmatism at 169±3 Ma (Rossi *et al.*, 2002). Geochemical features (GLOM, 1977; Durand Delga, 1997; Saccani, 2003) of basalts reveal E-MORB affinity, typical of crust developed during the initial stages of oceanic spreading (Marroni & Pandolfi, 2006). The ophiolitic sedimentary cover is formed by radiolarites (Late Dogger-Late Malm, Bill *et al.*, 2001; Dalenian *et al.*, 2008 and ref. therein), Calpionella Limestone (Tithonian-early Berriasian) and early Cretaceous S.Martino fm (Marroni *et al.*, 2000). The S.Martino fm. consists of c.100 m thick calcareous turbidites and shales interbedded with minor quartz-rich siltites, which have been correlated since the late seventies with the Palombini shales of Northern Apennine (Nardi, 1968; Durand Delga, 1978; Marroni *et al.*, 2000). The S.Martino fm. is followed by Lydienne Flysch (Routhier, 1956), consisting of silicified thin-bedded calcareous and mixed turbidites (Pandolfi, 2007), grading to the medium to coarse grained arenites and rudites of the Novella Sandstone, which is part of the Late Cretaceous (Albian-Late Cenomanian) siliciclastic turbidite system (Nardi *et al.*, 1978; Durand Delga *et al.*, 1978; Durand Delga, 1984; Marino *et al.*, 1995; Marroni *et al.*, 2007). The Novella sandstones are not found everywhere and they are locally replaced by two different coarse grained deposits, named Toccone and Alturaja formations. The Toccone breccias consist of coarse grained polygenic breccias with clasts supplied from continental-crust units (low grade metamorphites, granites, volcano-sedimentary, triassic dolomites, liassic limestones) as well as from the reworking of an ophiolitic sequence (basalts, radiolarites, limestones, lydiennes; Rossi et al, 2001). The Alturaja fm. (Alturaja Arkose) consists of massive to roughly bedded conglomerates interbedded with minor turbiditic sandstones (Marroni *et al.*, 2004). The clasts of this formation derived mainly from granitoids, low-grade metamorphic rocks and Si-rich volcanics (rhyolites to dacites). These deposits, dated to Late Barremian to Middle Aptian with palynological assemblages, are interpreted by Marroni *et al.* (2004) as the proximal part of a marine turbidite system very close to the source area. Marroni *et al.* (2006) considered the Alturaja fm. and the Toccone breccias heteropic with the Lydienne Flysch and Novella Sandstone. An alternative interpretation is favoured by Nardi *et al.* (1978), Durand Delga (1984) and Rossi *et al.* (2001), who proposed that the Alturaja fm. overlies unconformably the entire ophiolitic cover and locally rests above basalts (e.g. at Bocca di Fuata).

Figure 2.23. Balagne Nappe



General stratigraphic setting of the Balagne Nappe (after Durand Delga, 1984 modified in Spella *et al.*, 2009).

The sedimentary sequence of the Balagne Nappe is completed by the 200-300 m thick Eocene deposits of the Annunciata Fm. (Palasca Sandstone in Nardi *et al.*, 1978). This formation, which is interpreted as a siliciclastic turbidite fed by a continental-crust source area, consists of meter thick medium to coarse grained sandstones alternated with siltites and shales,. A Middle Eocene depositional age (Lutetian 46-40 Ma) has been proposed based on the presence of Nummulites brogniarti (forme A), Discocyclina, Asterodiscus, Amphistegina (Nardi, 1968; Bonnal, 1972; Lacazeidieu, 1974; Durand Delga, 1976; Nardi *et al.* 1978). These findings have been confirmed recently with nannoplancton data in Marino *et al.*, 1995 (Rossi *et al.*, 2001).

The close sedimentological analogies and the similarities in age (Middle Eocene) of the Annunciata fm. and the Eocene cover of the “Autochthonous” basement represent a key point in the long standing debate about the Balagne region, in particular for the nature and significance of the ophiolitic Balagne nappe. This point is discussed below.

The Balagne Nappe has been deformed at shallow structural levels with max *P* not exceeding 0.3/0.4 GPa at a *T* ~150/200°C (Marroni and Pandolfi, 2003). Different generations of superimposed structures can be recognized in the supraophiolitic cover including an early slaty cleavage associated with meter to map scale tight folds and later crenulation cleavages associated with open to tight folds. Fold-facing and kinematics along localized zones of deformation point to a dominant top-to-the west sense of transport. In the Eocene Annunciata Flysch simpler deformation patterns, which developed at shallower depths, have been described (Egal, 1992). The Balagne Nappe system can be subdivided into several minor tectonic elements (Nardi *et al.*, 1978; Durand Delga, 1978; Egal and Caron, 1988; Egal, 1992; Rossi *et al.*, 2001; Marroni & Pandolfi, 2003), all of them characterized by constant and dominant top-to-the west kinematics.

The Balagne region has been the classical ground for geological debate since the beginning of the 20th century. A complete historical report of such history can be found in Nardi, 1968; Durand Delga, 1978; furthermore in Rossi *et al.* (2001) it is possible to trace back the issue to the early “autochthonist” vs. “allochthonist” debate, including the discussion on the sense of movement of the Nappe Supérieure from west to east (Steinman, 1907) or from east to west (Maury 1907). These topics were discussed also in a field trip organized in 1928 (first edition of CorseAlp!!) by French geologists (P. Termier, E.Maury, E.Raguin) with Alpine “tectoniciens” (Steinmann G., Staub R., Kober L., Tillman N.) as well as in the more recent field trip of the Réunion extraordinaire de la Société Géologique de France in 1976 (Amaudric du Chaffaut and Campreon, 1976). The debate on the significance, origin and tectonic evolution of the Balagne Nappe and of the Nappe Supérieure is still ongoing.

Some authors e.g. Durand Delga (1976); Durand Delga (1984); Principi and Treves (1984); Durand Delga (1998); Marroni and Pandolfi (2003, 2007) support an original location of the Balagne Nappe and the Nappe Supérieure close to the Corsican continental margin (Balano-Ligure domain of Durand Delga, 1984). This interpretation is mainly based on the stratigraphic and geochemical characteristics of the ophiolites and the related sedimentary cover. Following Durand Delga *et al.* (2002) these features can be summarized as follows:

1) Presence of debris derived from continental crust at different stratigraphic levels within the ophiolitic sequence and cover series:

(a) Late Malm zircon-bearing sandstone associated with basalts, with zircon identical to those of the neighbouring granitic Variscan batholith of Corsica;

(b) continental debris within latest Malm limestone (S.Colombano);

(c) continental source area for Late Cretaceous and Eocene siliciclastic deposits correlated with Hercynian Corsica rock-types;

2) E-MORB composition of the oceanic basalts formed during early stage of oceanization in the ocean-continent (Corsica) transition area.

Other authors (Nardi, 1968; Mattauer and Proust, 1975; Nardi *et al.*, 1978; Dallan and Nardi, 1984; Malavieille *et al.*, 1998; Molli, 2008) considered the Balagne Nappe and the Nappe Superiore as derived from a more “internal" position, in an oceanic domain located far east from the Corsican margin (“ultra-Schistes Lustrés”) toward the “Apenninic” oceanic domain.

Our view (Molli and Malavieille, 2010) envisages the Balagne-Nebbio-Macinaggio units (Balagne and Nappe Supérieure system) as the relict of an intraoceanic accretionary wedge formed by offscrapping and shallow underplating of an oceanic part of Ligurian Tethys far from the Corsican continental margin. The Ocean-Continent transition originally located directly to the east of the Corsican continent is in our view sampled in the eclogitic and blueschist units of Serra di Pigno, Farinole and Mont San Petrone area (see Field trip 1 and Field trip 2.1, Vitale Brovarone *et al.*, 2010; 2011; Vitale Brovarone 2011). In this context, the Balagne ophiolite could be derived from the opposite OCT, close to the microcontinent located east of the Mesozoic Corsican margin. The continental basement-derived Late Cretaceous sediments in the supra-ophiolitic cover could have been deposited in a forearc basin at the rear of a growing accretionary wedge. Therefore, the Nebbio-Calabria microcontinent (continental edge block) could have been the source area of the Late Cretaceous siliciclastic deposits, as also suggested by Malavieille *et al.* (1998); Michard *et al.* (2002); Molli and Malavieille (2010 and references therein); Vitale Brovarone (2011); Vitale Brovarone *et al.* (in prep).

Day Three Itinerary

Stop 3.1:

Locality: Colle di U’Vezzu (UTM 32T 521706 E 4720892)

Themes: General introduction to the geology of the Balagne region and to western limb of Tenda antiform, which are visible in the landscape

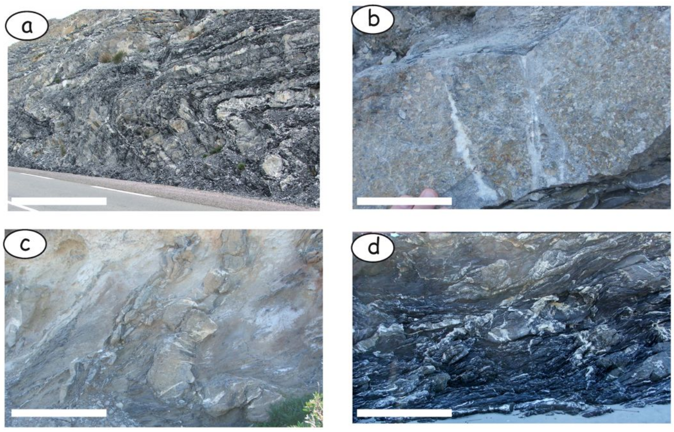
Stop 3.2

Locality: Ostriconi (UTM 32T 504843 E 4722876)

Themes: Nappe Supérieure (Flysch di Ostriconi), sedimentary and structural features.

This stop aims to illustrate some of the sedimentary and structural features of the Flysch d’Ostriconi a Late Cretaceous (Senonian) deposit belonging to the Nappe Supériore System (Nardi *et al.*, 1978; Durand Delga & Magné, 1978). Although the best exposures are located along the new road cut, the old RN 199 allows observations of the calcareous-marly sequence (Fig. 2.24a) associated with siliciclastic deposits, including microconglomerates and sandstones sourced from continental basement (Fig. 2.24b). The sequence is affected by bedding-parallel extension of calcareous beds, testified by incipient to well developed boudinage structures and by asymmetric west-vergentfolds, ranging in size from one meter to several meters, associated with a east-dipping cleavage (Fig. 2.24c). Towards the basal contact of the unit it is possible to notice a marked increase of calcite/quartz veins sets (Fig. 2.24d), which are especially well exposed in the outcropsalong the coast.

Figure 2.24. Flysch d'Ostriconi



Some stratigraphic and structural features of the Flysch d'Ostriconi. See text for details.

Stop 3.3

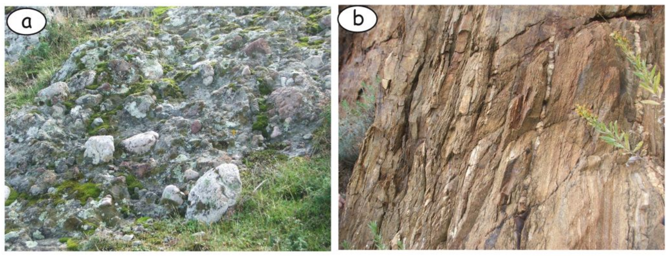
Locality: Cappella di San Sebastiano (UTM 32T 5032279 E 4715484)

Themes: Hercynian authochtonous basement and alpine sedimentary cover sequence.

Along the DN 163 toward Palasca, walking towards the old chapel of San Sebastiano, we will cross the unconformity separating the Hercynian basement from its sedimenatry cover belonging to the Alpine cycle. The basement rocks, characterized by a steep foliation, mainly consist of migmatitic gneiss (Fig. 2.25b) and minor bodies of orthogneiss dated at 338 Ma (Menot *et al.*, 1996) belonging to the Belgodere gneiss complex. These rocks are intruded by by Carboniferous pegmatoidal dykes.

The basement rocks are uncomformably covered by polygenic conglomerates (Fig. 2.25a) called 'Poudinges de Palasca', with clasts of granitoids, gneisses (dominant) and rare Permian cover rocks ranging in size from a few centimetres to half a meter. The conglomerates, dated to the Paleocene-early Eocene (Rossi *et al.*, 2001), show a variable thickness ranging from a few meters up to 300 meters and are locally associated with coarse sandstone.

Figure 2.25. Authochtonous Hercynian basement and its alpine sedimentary cover sequence.



(a) Paleocene-early Eocene Conglomerates; (b) Belgodere gneiss.

From the San Sebastiano chapel looking to the SE (Fig. 2.26) its is possible to observe the geometry of the nappe stack of the Balagne region. From SE (right) to NW (left) and from bottom to the top, the following units can be recognized:

1) the Hercynian basement and its Paleocene-early Eocene sedimentary cover (including the Flysch noir in this landscape view);

2) the Annunciata unit (early Mid Eocene sandstone) with a base of tectonic slices including exotic Jurassic limestones, slices of external continental units and elements of ophiolitic Balagne Nappe (ophiolitic breccias and Lydienne) (south of the view);

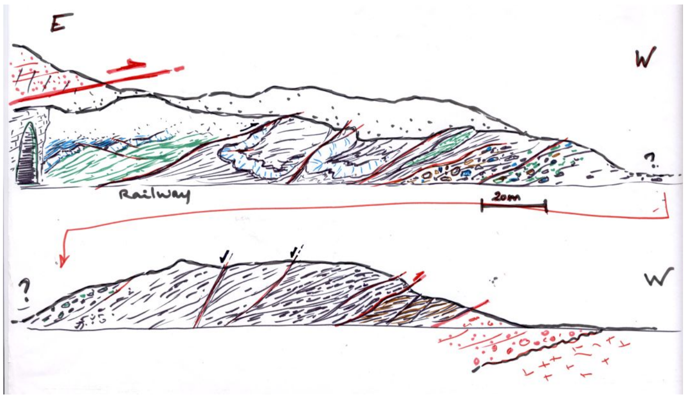
3) the Toccone unit, mainly consisting of Cretaceous Lydienne Flysch and ophiolitic breccias (Brecce di Toccone);

Figure 2.26. San Sebastiano Chapel

(a, b) Panoramic view from San Sebastiano Chapel of the nappe stack geometry of the Balagne region.

A detail of the main basal thrust of the Annunciata unit on the authochtonous basement and cover can be seen in Fig. 2.27

Figure 2.27. A schematic field drawing of the main basal thrust of the Annunciata unit on the authochtonous.



Stop 3.6

Locality: Col de San Colombano (UTM 32T 505813 E 4714121)

Themes: Panoramic view of the north Balagne Region, the ophiolitic Navaccia unit, main lithotypes and their structural features.

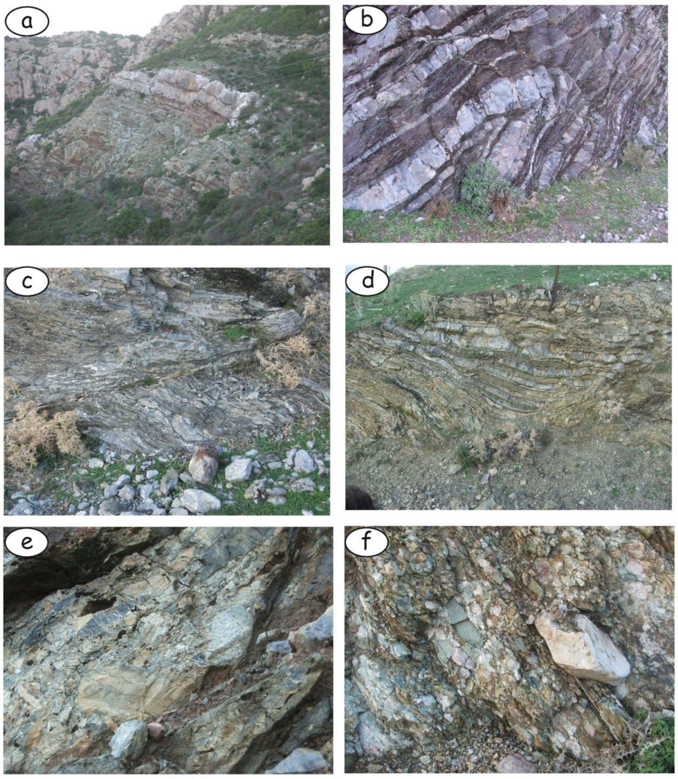
From the Col de San Colombano looking to the north a panoramic view of Balagne Region can be observed (Fig. 2.28) including, from left to right, the Hercynian basement, the Annunciata Eocene Sandstone (with beautiful west vergent fold systems) and the Balagne units. Further to the north, the central ridge with vegetation is located within the Nappe Superiore (Ostriconi), whereas towards the east the Cima delle Forchie unit (mainly granitoids), the volcano-sedimentary and Tenda orthogneiss, can be easily recognized. Polygenic Cretaceous Conglomerates (Alturaja fm) can be observed to the right, close to the lookout point.

Figure 2.28. Col de San Colombano

A Panoramic view of the Balagne Region From the Col de San Colombano looking toward north. See text for description.

The walk from the San Colombano Pass to the hill on the east allows to see different lithotypes and sub-units of the Balagne Nappe System (San Colombano, Toccone and Alturaja sub-units of Marroni & Pandolfi, 2003) and their internal structures (Fig. 2.29).

Figure 2.29. San Colombano track



Some rock-types and structures observable along the San Colombano track, see text for descriptions.

Most lithologies of the Balagne ophiolitic sequence can be analyzed here, including: Jurassic basalts (mainly pillow lava), Late Jurassic red Cherts (Late Dogger-Late Malm in age, Bill *et al.*, 2001; Dalenian *et al.*, 2008 and ref. therein) (Fig. 2.29a,b), Calpionella limestone (2.29a), including the 'Gran Rocher limestone,' which is well known in Corsican geological literature

The Calpionella and Grand Rocher limestone (Tithonian-early Berriasian) here include levels of coarse to medium grained breccia in carbonate matrix. Clasts consist mainly of continental basement (low and medium/high grade rocks, igneous intrusives and volcanic-subvolcanics rocks) as well as Triassic-early Jurassic carbonates. The Calpionella limestone is followed by the early Cretaceous S.Martino fm (Fig. 2.29c) (Marroni *et al.*, 2000 cfr. Palombini shales of Northern Apennine) and Lydienne Flysch (Fig. 2.29d) (Late Hauterivian-Early Barremian Marroni *et al.*, 2000), thin bedded turbidites locally with fine to coarse grained sandstone layers (mapped as 'Novella Sandstones' when especially abundant) and/or ophiolitic breccias (Fig. 2.29e; Toccone Breccia). The sequence is capped by a thick sequence of coarse grained conglomerates (Alturaja fm) early-Mid Aptian in age (Marroni *et al.*, 2004), which will be observed at the top of the hill (Fig. 2.29f).

All the units are strongly deformed, with localized shear zones characterized by foliated cataclasites that accommodated top-to-west displacement (Fig. 2.29c,d). Folds ranging in size from a few meters to several tens of meters are also common (Fig. 2.29b).

Stop 3.7

Locality:km 59 railway track (UTM 32T 510165 E 4710559)

Themes: Jurassic ophiolites and their sedimentary cover

A section through the top of the Tethyan ocean floor can be observed along the railway. Pillow lavas and lava tubes are covered by pillow breccias, themselves capped by deep marine sediments. The sedimentary sequence consists from bottom to top of red radiolarites (late Dogger, late Malm), interbedded calcareous turbidites and shales and late Cretaceous Lydienne Flysch.

Walking further along the road, a large body of pillow lavas allows observations of well preserved features of submarine volcanism.

Stop 3.8

Locality: Carrière de Taverna (UTM 32T 515768E 4693748)

Themes: Francardo Miocene basin. Sedimentary features and tectonic setting.

During the Oligocene, drifting and subsequent rotation of the Corsica-Sardinia block resulted in the development of the gulf of Lion oceanic domain. In Miocene time, rifting resumed and the subsequent cooling of the oceanic lithosphere led to thermal subsidence and marine invasion in parts of Alpine Corsica, where Miocene sedimentary basins controlled by transtensional faults were formed.

The last stop of the day is in the Francardo basin, which was related to these processes

Outcrops in the Taverna quarry allow observations of the different structures and sedimentary facies that characterize the paleoenvironmental and tectonic setting of this basin.

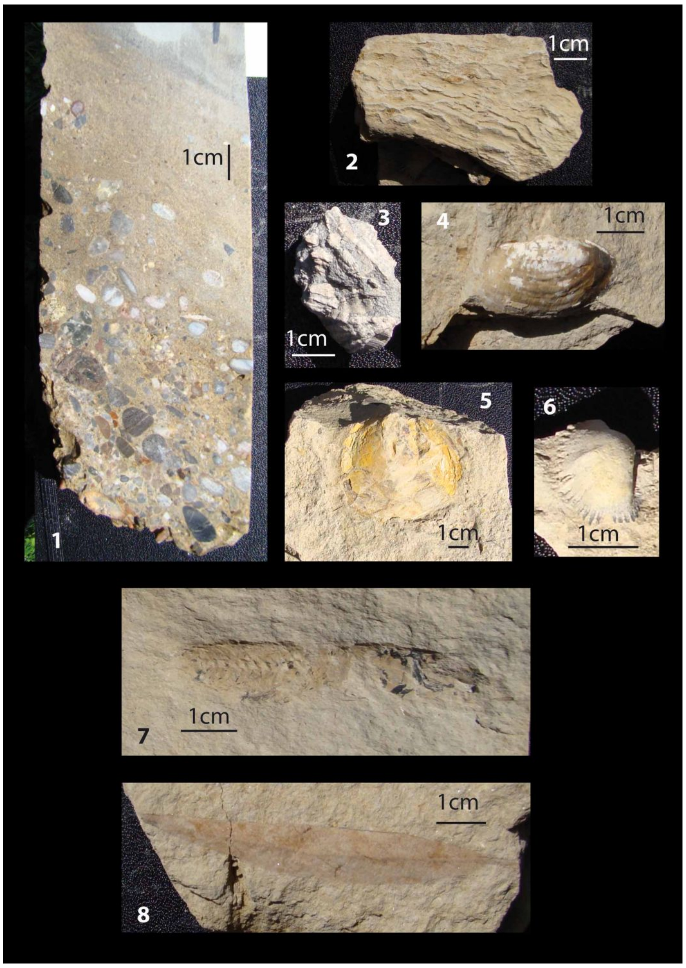
Sediments were deposited unconformably on, the autochthonous hercynian basement made by Permian alcaline granitoïds, Eocene para autochthonous units, pre-piemontese Caporalino-Pédani allochthonous units to the west and the Ligurian ophiolitic units to the North and East. Coarse detrital sediments deposited on paleoreliefs marks the base of the basin. The main stratification presents a slight dip (10-15°) to the west. North-South oriented high-angle normal faults are related to a transtensional tectonic setting during basin formation. They developed before tilting of the series related to a regional scale synformal structure that is bounded by similar normal faults to the west and to the east.

The Taverna formation, which crops out in the quarry, consists from bottom to top of :

- sandstones and clays including pit layers,
- marls and fossiliferous sandstones interlayered with limestones layers and conglomerates,
- conglomerates deposited in a shallow marine environment (presence of ripple-marks and oyster debris) covered by a thick sandstone sequence.

This formation contains Mid-Burdigalian to Upper Burdigalian fossils (Fig. 2.30). These fossils are typical of a shallow marine littoral. Clear water ostracods and fish living in brackish water characteristic of lagoons are also found. Fossils of plants and coal found in grey marls suggest the vicinity of the coast. All fossils indicate a hot climate.

Figure 2.30. Burdigalian fossils



1: Polygenic conglomerate associated with medium to fine sandstone and siltstones; 2: échantillon TAV 21, south face of quarry. Croûte carbonatée, stromatolite; 3: Balanus sp., 4: Tellina sp.; 5: Corbicula sp.; 6: Arca sp.; 7: Aphanius sp.; 6: Empreinte de feuille, bloc éboulé, front de taille nord de la carrière.

Analysis of conglomerates suggests that during basin develoment, the main source of sediments was from proximal reliefs located to the west. Indeed, most of the pebbles consist of rocks types typical of the crystalline basement and of its Mesozoic and Eocene cover. The absence of metamorphic rocks from the ophiolitic schistes lustrés nappe may indicate that during the Miocene these rocks remained below sea level.