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Deformation behavior of continental crust during subduction and exhumation: Strain distribution over the Tenda massif (Alpine Corsica, France)

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Abstract

In order to address the question of strain localization within continental units during subduction and exhumation, a large-scale portion of an exhumed continental crust was structurally revisited. The Tenda massif (Alpine Corsica) has recorded burial (D_1 ; top-to-the-SW kinematics) down to blueschist-facies conditions followed by exhumation (D_2 ; top-to-the-NE kinematics). It was so far regarded as a quite rigid unit with strain localization at the upper contact with the overlying oceanic material, the East Tenda Shear Zone (ETSZ), where previous studies were focused. A structural analysis carried out from the core to the boundaries of this continental unit shows instead that deformation is pervasive in the whole

section. A reappraisal of the finite structure shows that this unit corresponds to a N140°E elongated dome of D₂ fabrics severely overprinting earlier D₁ structures, locally preserved in the core of the dome. Field observations show that deformation was distributed at maximum burial, pervasively affecting both the Tenda massif and the overlying tectonically coupled oceanic material. Strain remained distributed during the first steps of exhumation that probably started during the last stages of subduction. After a regional switch from compression to extension, deformation localized toward the ETSZ and other internal shear zones such as the newly described Saleccia Shear Zone, as shown by strain intensity mapping in the eastern granitic protoliths. Doming occurred during this localization when crossing the brittle-ductile transition. This large-scale structural study shows that lithological boundaries controlled the place where strain localized, while the general style of deformation (distributed vs localized) was controlled by depth (i.e. pressure-temperature conditions), interacting with other parameters such as fluid circulation and protolith composition.

Key-words: Alpine Corsica; Tenda massif; Continental crust; Exhumation; Shear zone; Strain localization.

1. Introduction

The progressive deformation of continental basement units from the thrust front to the internal zones of subduction wedges largely depends upon the rheology of the subducting lithosphere. In particular, the style of deformation near the thrust front is diagnostic of the initial rheological stratification of the foreland domain of orogens (Mouthereau et al., 2013). Although the continental basement is often considered rigid in forelands, the western Alps or the Zagros show thick-skin deformation near the thrust front where both basement rocks and

cover series experienced significant amounts of shortening along localized shear zones active in the conditions of the greenschist facies (e.g. Leloup et al., 2005; Mouthereau et al., 2007; Bellahsen et al., 2012; Bellanger et al., 2014). In the internal zones of orogens, continental units are intensely deformed and sheared in high-pressure and low-temperature (HP-LT) conditions (blueschists and eclogite facies) during both subduction and exhumation (e.g. Huet et al., 2009). Thus, from the external to the internal zones, deformation is less and less localized mainly because temperature increases, but the evolution of continental units between these two end members is poorly known.

The Tenda massif in Alpine Corsica provides an example of subduction of a continental basement which experienced both intermediate P - T conditions (lower blueschists facies) and strong strain localization along the East Tenda Shear Zone (ETSZ), which was activated during subduction and exhumation (Cohen et al., 1981; Mattauer et al., 1981; Jourdan, 1988; Jolivet et al., 1990; 1991; Daniel et al., 1996; Molli and Tribuzio, 2004; Molli et al., 2006; Maggi et al., 2012; 2014; Rossetti et al., 2015).

The Alpine Corsica HP-LT orogenic wedge includes both Ligurian oceanic and European continental units (e.g. Mattauer et al., 1981; Agard and Vitale-Brovarone, 2013; Vitale-Brovarone et al., 2013). It is part of a larger orogenic system of which the most external zones are the Provence fold-and-thrust belt and the poorly deformed Western Corsica Variscan basement (Lacombe and Jolivet, 2005). The internal parts of the wedge, resting on top of Western Corsica, correspond to a westward-verging stack of nappes characterized by an eastward metamorphic gradient from lower blueschist-facies in the western continental Tenda massif to eclogite-facies conditions in the eastern oceanic Schistes Lustrés nappe (e.g. Mattauer et al., 1981; Lahondère, 1988; Fournier et al., 1991; Caron, 1994; Vitale-Brovarone et al., 2013). Several tectonometamorphic units are superimposed and separated by tectonic contacts and sharp metamorphic gaps (Fournier et al., 1991; Caron, 1994; Malavieille et al.,

1998; Vitale-Brovarone et al., 2013). The Tenda massif represents a part of the continental margin involved in the orogenic wedge. It is regarded as a quite rigid block that escaped significant internal deformation during subduction and exhumation (Mattauer et al., 1981; Daniel et al., 1996; Gueydan et al., 2003; Molli et al., 2006; Maggi et al., 2012; 2014; Rossetti et al., 2015) with shearing mostly localized along its eastern margin, the ETSZ, where most of earlier studies have focused so far. This apparent rigidity resulted in contrasting interpretations regarding strain distribution and exhumation mechanisms. Although this major structure has probably accommodated a significant part of the finite deformation, the internal deformation and its evolution toward the ETSZ have never been fully described.

In this study, we reviewed the style, magnitude, and significance of deformation of the entire Tenda massif based on a new and extensive structural characterization. Finite strain markers relative to burial and exhumation were identified, mapped and characterized across a large portion of this continental unit. A scale of relative strain intensity and partitioning in the eastern granitic complex of the Tenda massif was built with the aim to reconstruct a succession of superimposed structures in a continuum of deformation. Thus, a new high resolution map of the intensity of deformation is here presented, having implications on strain distribution, strain localization factors and exhumation processes during the whole subduction/exhumation story.

2. Geological setting

2.1. Alpine Corsica

Corsica is divided in two geological domains: Variscan Corsica in the west and Alpine Corsica in the northeast (Durand-Delga, 1984) (Fig. 1). Alpine Corsica consists of a stack of

variably metamorphosed tectonic units belonging to a *HP-LT* subduction complex (e.g. Mattauer et al., 1981; Lahondère, 1988; Fournier, 1991; Caron, 1994; Vitale Brovarone et al., 2013). Formerly attached to the European continent, Corsica and Sardinia have migrated during the rifting of the Liguro-Provençal basin to reach their current position in Oligocene-Miocene times (e.g. Réhault et al., 1984; Speranza et al., 2002; Ferrandini et al., 2003; Maffione et al., 2008). Although multiple tectonometamorphic units can be distinguished (e.g. Vitale Brovarone et al., 2013), Alpine Corsica is conventionally divided into three major groups of units, from bottom to top: 1 - the external continental units, including the Tenda massif; 2 - the Schistes Lustrés nappe; and 3 - the Nappes Supérieures (Durand-Delga, 1984; Jolivet et al., 1990; Malavieille et al., 1998; Molli, 2008) (Fig. 1).

The external continental units correspond to autochthonous or parautochthonous units that were involved in Alpine subduction under greenschists to lower blueschists-facies conditions for the Sampolo and Popolasca units (Malasoma et al., 2006; Malasoma and Marroni, 2007; Garfagnoli et al., 2009; Vitale Brovarone et al., 2013; Di Rosa et al., 2016) up to middle-grade blueschists-facies conditions for the Tenda massif (Gibbons and Horak, 1984; Tribuzio and Giacomini, 2002; Molli and Tribuzio, 2004; Maggi et al., 2012; Vitale Brovarone et al., 2013).

The Schistes Lustrés nappe is mainly composed of more or less coherent pieces of metabasites, metasediments of oceanic affinity and few subordinate continental units (Centuri and Serra di Pigno units). These continental slices were incorporated in the Schistes Lustrés complex during the Alpine subduction (Faure and Malavieille, 1981), or prior to subduction, as extensional allochthons, in an Ocean-Continental Transition zone (OCT) (Lahondère and Guerrot, 1997; Martin et al., 2011; Vitale Brovarone et al., 2011a; Meresse et al., 2012). The Schistes Lustrés nappe was deeply buried during subduction as evidenced by the widespread occurrences of blueschist- and eclogite-facies assemblages present in both internal continental

units and metabasites (Caron et al., 1981; Caron and Péquignot, 1986; Lahondère, 1988; Fournier et al., 1991; Caron, 1994; Jolivet et al., 1998; Ravna et al., 2010; Vitale Brovarone et al., 2011a; 2011b).

The Nappes Supérieures (Balagne, Nebbio, Macinaggio, Pineto and Inzecca units) essentially consist of ophiolites of Jurassic age and Cretaceous flyschs of Ligurian affinity (Beccaluva et al., 1981; Ohnenstetter et al., 1981; De Wever et al., 1987) of low metamorphic grade (i.e. $T_{\max} = 300\text{ °C}$, Vitale Brovarone et al., 2013).

These nappes are overthrust on the autochthonous crystalline basement of Variscan Corsica and Eocene foreland deposits (Fig. 1) ensuring a Bartonian biostratigraphic constraint on the timing of Alpine deformation at the front of the belt (Bonnal et al., 1973; Bézert and Caby, 1988; Jourdan, 1988; Egal and Caron, 1989; Vitale Brovarone et al., 2013). *HP-LT* metamorphism in the Schistes Lustrés nappe yielded scattered ages spanning from controversial Late Cretaceous ages (Sm-Nd on whole-rock, garnet, jadeite and glaucophane, Lahondère and Guerrot, 1997; $^{40}\text{Ar}/^{39}\text{Ar}$ on glaucophane, Maluski, 1977) to Late Eocene ages ($^{40}\text{Ar}/^{39}\text{Ar}$ on phengite, Brunet et al., 2000; U-Pb on zircon, Martin et al., 2011; Lu-Hf on Garnet; Vitale Brovarone and Herwartz, 2013). *HP-LT* rocks from the Tenda massif yielded almost similar ages from $54 \pm 8\text{ Ma}$ (U-Pb on acmite-phengite assemblages, Maggi et al., 2012) and $\sim 36\text{ Ma}$ ($^{40}\text{Ar}/^{39}\text{Ar}$ on phengite, Brunet et al., 2000). Similar ages (37-33 Ma) were retrieved from the most external Alpine shear zones ($T = 300\text{ °C}$; $P > 0.6\text{ GPa}$) affecting the Variscan basement ($^{40}\text{Ar}/^{39}\text{Ar}$ on phengite; Di Vincenzo et al., 2016).

Timing of exhumation and extensional tectonics appears also quite well constrained. Extensional regime is ensured by the development of extensional sedimentary basins that overprinted the structure of the subduction complex, notably the St-Florent basin that developed on top of the Nebbio and the Schistes Lustrés nappe. Activity of the bounding normal faults is ascribed to the syn-tectonic deposition of Burdigalian to Serravallian marine

sediments (Ferrandini et al., 1998; Cavazza et al., 2007). Besides, the youngest retrogression ages associated to exhumation retrieved from greenschist-facies rocks are well constrained in both the Schistes Lustrés nappe and the Tenda massif between 25 and 20 Ma (Brunet et al., 2000; Rossetti et al., 2015). Final exhumation stages of the internal zones including the Tenda massif as constrained by fission tracks low-temperature thermochronology on both zircon and apatite yielded Late Oligocene to Miocene ages (Jakni et al., 2000; Fellin et al., 2006).

At first glance, the finite structure of Alpine Corsica therefore results from a quite simple two-stepped tectonometamorphic evolution of subduction (burial) and exhumation. The first event corresponds to ductile deformation recorded in HP-LT conditions in both the Schistes Lustrés and Tenda units. This event is associated with a top-to-the-SW shearing localizing toward the contact between the two units that accommodated the thrusting of the Schistes Lustrés nappe over the Tenda massif (Cohen et al., 1981; Faure and Malavieille, 1981; Mattauer et al., 1981; Jourdan, 1988; Molli and Tribuzzio, 2004; Molli et al., 2006). It is followed by a post-HP event including series of top-to-the-E shearing and E-verging folding extensional increments in the Schistes Lustrés nappe (Faure and Malavieille, 1981; Jolivet et al., 1990; 1991; Fournier et al., 1991) and the Tenda massif (Jourdan, 1988; Jolivet et al., 1990; 1991; Daniel et al., 1996; Molli and Tribuzio, 2004; Molli et al., 2006). These late structures were interpreted as reactivating in the opposite direction some early thrusts either as backthrusts (Faure and Malavieille, 1980), or as extensional shear zones (Jolivet et al., 1990; 1991; Fournier et al., 1991; Daniel et al., 1996; Molli and Tribuzio, 2004; Molli et al., 2006; Rossetti et al., 2015). Interpreted as the result of syn-orogenic extension or regional post-orogenic extension, the geodynamic context that prevailed during early stages of exhumation is a matter of debate, leaving interpretation of some first order contacts, such as the ETSZ, controversial (Molli et al., 2006).

2.2. The Tenda massif and the ETSZ

The Tenda massif (Fig. 1) is bounded to the east by the ETSZ that marks the contact with the Schistes Lustrés nappe and to the west by the Ostriconi fault, which instead puts it in contact with the Nappes Supérieures of Balagne (Jourdan, 1988). It consists primarily to a Variscan basement portion overlain by Meso-Cenozoic cover series all involved into Alpine tectono-metamorphic events (Fig. 2). To the west, it is represented by Proterozoic to Paleozoic medium-grade rocks (gneisses and micaschists) unconformably overlain by Late Variscan volcano-sedimentary rocks (Rossi et al., 1994). At variance, the eastern part of the massif is mainly composed of a Late Carboniferous-Early Permian granitic complex (Rossi et al., 1993), intrusive in the volcano-sedimentary rocks. It includes from east to west the Mt Asto monzogranite, directly under the Schistes Lustrés nappe, the Casta granodiorite and the Mt Genova leucomonzogranite, in contact with the volcano-sedimentary and micaschists sequences (Rossi et al., 1994). The Triassic to Eocene sedimentary cover that remains united with the basement is only locally preserved, notably near Santo-Pietro-di-Tenda in the east (Jourdan, 1988; Rossi et al., 2001; Vitale Brovarone et al., 2013).

Evidence of the Alpine subduction of the Tenda massif is expressed by both an intense deformation near the ETSZ and the local development of *HP-LT* index metamorphic associations. They are often limited to celadonite-rich phengite in all types of intrusive rocks, and sodic blue-amphibole (riebeckite-ferroglaucophane), jadeite-bearing aegirine or even clinopyroxene + rutile, in granitoids of peculiar composition (Gibbons and Horak, 1994; Tribuzio and Giacomini, 2002; Molli and Tribuzio, 2004; Molli et al., 2006; Maggi et al., 2012; Rossetti et al., 2015). Peak metamorphic conditions estimates fall in the blueschists-facies conditions in the range of 1-1.2 GPa and 400-450 °C (Tribuzio and Giacomini, 2002; Molli and Tribuzio, 2004; Molli et al., 2006; Maggi et al., 2012) confirmed by independent

temperature estimates retrieved on metasediments from cover series (Vitale Brovarone et al., 2013). Exhumation is marked by an overall pressure decrease inferred from the decrease in Si-content in phengite (Molli et al., 2006; Maggi et al., 2012; Rossetti et al., 2015). An exhumation path was recently recognized from ~1 GPa and 450 °C to < 0.5 GPa and 300-400°C (Tribuzio and Giacomini, 2002; Molli and Tribuzio, 2004; Molli et al., 2006).

Just as the whole Alpine Corsica, the finite structure of the Tenda massif was achieved through a two-stepped history. All studies agree on a first event, D_1 , responsible for the development of a first planar-linear fabric (referred as S_1/L_1 in this study) accompanied with a top-to-the-SW sense of shear during subduction and burial in blueschists-facies conditions (Cohen et al., 1981; Mattauer et al., 1981; Jourdan, 1988; Daniel et al., 1996; Molli and Tribuzio, 2004; Molli et al., 2006; Molli and Malavieille, 2011; Maggi et al., 2012; 2014; Rossetti et al., 2015). This event corresponds to the underthrusting of the Tenda massif below the Schistes Lustrés nappe. Except for D_1 , models and history of subduction/exhumation and strain localization of the Tenda massif in the frame of Alpine Corsica can be very different depending on papers. Three models are proposed:

- For Jolivet et al. (1990; 1991), Daniel et al. (1996) and Gueydan et al. (2003), D_1 is placed during the Eocene. Then, a post-orogenic exhumation of Alpine Corsica started near 30 Ma, with top-to-the-NE deformation localized notably toward the ETSZ that is reworked in greenschist-facies conditions. This D_2 phase is associated to folds affecting the former S_1 , and a crenulation cleavage (referred as S_2 in this study) preserved in less non-coaxial deformation domains between top-to-the-NE shear zones. A second lineation (referred as L_2 in this study) developed on S_2 planes. The feldspar-to-phengite reaction is responsible for strain localization near the contact with the Schistes Lustrés nappe (Gueydan et al., 2003). A D_3 phase is used in this model for brittle extensional structures in a continuum of deformation during exhumation (Daniel et al., 1996).

- A more recent model rather considers most of the exhumation syn-orogenic (Molli and Tribuzio, 2004; Molli et al., 2006). The D_1 phase is therefore placed earlier, before the Early Eocene and until ~ 40 Ma. S_1 folding and crenulation cleavage are interpreted occurring in low-strained domains in between zones of localized top-to-the-SW shearing during D_1 (Molli et al., 2006; Rossetti et al., 2015). After ~ 40 Ma, during thrusting of the most external continental domains, a first step of syn-convergence exhumation (D_2) occurred for both the Schistes Lustrés nappe and the Tenda massif until ~ 32 Ma. It is accommodated by a top-to-the-SW shear zone contact below the Tenda massif and a top-to-the-NE shear zone at the top of the Schistes Lustrés nappe (Molli and Tribuzio, 2004; Molli et al., 2006). Associated structures are D_2 folds, crenulation and both top-to-the-NE and -SW shear bands formation in the Tenda massif, and penetrative folding in the Schistes Lustrés nappe. The contact between the two units is itself involved in large-scale D_2 folds, in a general delocalization of strain within the orogenic wedge. The ETSZ is finally reactivated in a post-orogenic context until brittle conditions at the end of the exhumation (D_3). Those final phases of exhumation evolved from the ductile field (top-to-the-NE shearing and folds with axes parallel to the stretching lineation) up to the brittle field, with the development of semi-brittle shear bands and cataclasite-bearing fault zones.

- A third model (Maggi et al., 2012; 2014; Rossetti et al., 2015) is in agreement with the timing of events and the syn-orogenic character of exhumation described in the previous model of Molli and Tribuzio (2004) and Molli et al. (2006). However, the ETSZ remained active during exhumation with progressive strain localization from distributed in gneiss lenses to localized in top-to-the-SW shear zones. The ETSZ is characterized by a transient rheology with softening during feldspar-to-micas reaction and hardening by noeblastesis of K-feldspar during fluid metasomatism. The ultimate stage of localization is top-to-the-SW phyllonites,

which are the only structures reactivated semi-brittlely during post-orogenic extension. In their model, the Tenda massif remains undeformed beneath the ETSZ.

3. Structure and finite strain of the Tenda massif

A field survey was carried out on the whole Tenda massif including a detailed study of structural features and geological mapping in order to complement the map of the French Geological Survey (BRGM; Rossi et al., 1994; 2001). Strain markers were retrieved over more than 700 measurement stations throughout the massif. In the deformed parts, all metamorphic rocks are often pervasively foliated and generally exhibit a well-defined stretching lineation. Results are projected on the maps of figures 2 (foliations) and 4 (stretching lineations with sense of shear) and are illustrated by representative outcrops and cross sections (Figs. 3 and 5), complementing earlier studies focused on the ETSZ (Daniel et al., 1996; Molli and Tribuzio, 2004; Molli et al., 2006; Maggi et al., 2014).

3.1. Foliations and folds

At least two planar fabrics, related to D_1 and D_2 , are observed, some forming large-scale features apparent in the landscape (Fig. 3). A total of 207 S_1 and 482 S_2 planes were measured in all lithologies of the Tenda massif and in the Schistes Lustrés nappe close to the contact (Fig. 2a). Both S_1 and S_2 are found in almost the whole volume of the massif showing that deformation is not restricted to the ETSZ. S_2 commonly dips away from the center of the Tenda massif with trajectories showing a concentric pattern depicting a N140°E elongated structural dome (Fig. 2). The apparent prevalence of NE-dipping planes is due to oversampling of S_2 planes within the ETSZ. S_1 planes show instead a N170°E preferred

orientation (Fig. 2). S_2 generally displays dips between 0 and 45° with very low dispersion (Fig. 2c). S_1 dip shows instead a higher dispersion (from 0 to 90°) depending on the degree of D_2 reworking. Structural relationships between S_2 and S_1 were studied in detail along the main road from St-Florent to Bocca di Vezzu (see location on Fig. 2a). Deformation appears heterogeneously distributed with lenses-shaped domains associated to rocks showing no or only a weak record of D_2 deformation (see grey domains in Fig. 2b) and surrounded by anastomosed D_2 corridors (see S_2 trajectories in Fig. 2b). Low-strain domains consist in lense-shaped volumes of both unstrained Variscan magmatic rocks and weakly deformed facies, with incipient flat to gently dipping S_1 . The largest one occurs in the core of the Tenda massif and roughly corresponds to the cartographic outline of the Casta granodiorite (Fig. 2). When weakly reworked by D_2 , S_1 exhibits small- to large-scale undulations (Fig. 3a). With increasing D_2 reworking, S_1 is involved into recumbent isoclinal D_2 folds whose axial planes correspond to shallowly dipping S_2 crenulation cleavage (Figs. 3b and 3c). Both S_1 and S_2 carry a mineral association mainly made of quartz, albite and phengite in most granites, but in more mafic facies, blue-amphiboles highlight the folded S_1 , and is retrogressed into chlorite in S_2 cleavage (Fig. 3d). In those typical outcrops characterized by D_2 folds, envelopes of S_1 are steeply dipping and strike NNW-SSE while S_2 , in the same range of strike, is shallowly dipping (Fig. 3e). The associated D_2 folds axes also trend NNW-SSE with gently dipping plunges.

With increasing D_2 strain intensity, S_1 can be almost completely transposed into the S_2 fabric, especially along the Saleccia Shear Zone (SSZ) and the ETSZ where it can be observed at landscape-scale (Figs. 3f and 3g, respectively). The uppermost structural levels of the ETSZ show a single S_2 fabric that is itself involved in metric to decametric recumbent isoclinal folds in a progressive D_2 deformation (i.e. the D_3 folds of Molli et al., 2006). The geometry of these late D_2 folds evolves from open recumbent folds to overturned, then to

sheath folds, with trends axes evolving from NW-SE to ENE-WSW (Figs. 3h and 3i), parallelizing to the local stretching (Fig. 4). Dimensions of D_2 folds across the massif vary from a few centimeters to several tens of meters.

3.2. Stretching lineations

Stretching lineation is defined by various markers, depending primarily on the nature of the protolith and the strain intensity. For both D_1 and D_2 , stretching lineation is marked in granites by stretched quartz rods, truncation of feldspar, strain shadows, and more commonly by elongated phyllosilicate aggregates such as phengite. Similarly, stretching lineation is conspicuous in metasediments. It is marked by fine-grained slates of phyllosilicates in pelitic layers or by the elongation of stretched pebbles in metaconglomerate layers. More rarely, it is marked by the alignment and truncation of prismatic minerals such as blue-amphiboles. In most outcrops, the orientation of the various types of lineation is fairly constant. Just as S_2 , L_2 is ubiquitous across the massif but is particularly marked in the vicinity of the ETSZ and second-order shear zones. Conversely, L_1 is only locally expressed and restricted to low-strain domains.

A total of 166 L_1 and 408 L_2 orientations have been measured in all lithologies across the Tenda massif (Fig. 4a). L_1 planes were only measured where they have not been verticalized during D_2 . Trend and plunge of the different stretching lineations are almost comparable and show low dispersion (Fig. 4b) with a mean orientation of $N050^\circ E$ - $N060^\circ E$ and a dominant NE-plunge related to the oversampling of S_2/L_2 fabrics within the ETSZ. However, a slight yet reproducible difference in trends can however be detected at the scale of the outcrop. Consistent with the overall dome geometry, both L_1 and L_2 lineation mostly

plunge to the northeast in the northeastern flank and to the southwest in the southwestern flank.

4. Kinematics of the polyphased deformation

Kinematics of deformation recorded in the Tenda massif is complex and characterized by the progressive overprint of opposing shear senses. The early subduction-related top-to-the-SW deformation is largely overprinted by the late top-to-the-NE extensional shear that prevailed during exhumation (e.g. Mattauer et al., 1981; Jolivet et al., 1990; 1991; Fournier et al., 1991; Daniel et al., 1996; Gueydan et al., 2003; Molli and Tribuzio, 2004; Molli et al., 2006; Maggi et al., 2012, 2014; Rossetti et al., 2015). In this section, D_2 features are analyzed first and D_1 next in order to properly unfold the sequence of top-to-the-NE and SW-directed deformations and adequately characterize their relative contribution to the finite geometry of the massif.

4.1. Description and distribution of top-to-the-NE deformation (D_2)

Top-to-the-NE deformation is heterogeneously distributed within the Tenda massif (Figs. 4a and 5). It concentrates, at the first glance, toward the higher parts of the dome in the vicinity of the ETSZ and therefore mantles a relatively less deformed core preserving D_1 features (Fig. 5a). This deformation is spectacularly exposed within the northern outcrop of the ETSZ where a ca. 200 m-thick NE-dipping mylonites are classically described (Jolivet et al., 1990; 1991; Fournier et al., 1991; Daniel et al., 1996; Gueydan et al., 2003; Molli and Tribuzio, 2004; Molli et al., 2006; Maggi et al., 2012, 2014; Rossetti et al., 2015). Due to the dome-shaped architecture of the Tenda massif, top-to-the-NE deformation displays gentle to

moderate northeastward dips and normal-sense relative displacements on the northeastern flank of the dome while it often presents flat or even “reverse” geometry on the southwestern flank in the vicinity of the Ostriconi fault (Fig. 5a). This first-order geometry hides smaller-scale yet abundant top-to-the-NE structures that are particularly relevant for the discussion. Those observations are projected along the synthetic cross section presented in Fig. 5.

First, top-to-the-NE deformation also concentrates toward another shear zone of prime importance, the Saleccia Shear Zone (SSZ) that is described for the first time in this study (see Fig. 3f). There, D_2 deformation occurs within granitoids roughly 1 km structurally below the ETSZ and evolves upward from gently NE-dipping protomylonites preserving D_1 subvertical fabrics, to mylonites and ultimately rare stripes of ultramylonite bodies over a ca. 150 m-thick strain gradient where deformation localizes within the Saleccia rhyolite (Fig. 5b). In this gradient, the mylonitic S_2 strikes NW-SE and moderately dips to the northeast (Fig. 3f). In the most deformed domains of the SSZ, spacing between D_2 shear bands is locally as low as 0.5 to 1 cm (Fig. 5b) and it increases downward, leaving lens-shaped domains where S_1 is preserved.

Top-to-the-NE deformation also occurs as 10-50 cm-thick localized shear zones that rework previous fabrics or even undeformed protolith. Within the most deformed rock volumes, spacing between shear bands is locally as dense as 1-2 cm. Away from the core of shear zones, asymmetry and more generally imprint of D_2 deformation decreases over a short distance (Fig. 5c). Locally, shear zones at high angle with respects to the main foliation suggest a brittle component during top-to-the-NE deformation (Fig. 5d). These narrow shear zones are consistently NE-dipping from the west to the east of the Tenda massif (Figs. 5a, 5c and 5d), crosscutting the large-scale dome structure.

In various lithologies cropping out in the west (i.e. volcano-sedimentary rocks), shear bands are the most ubiquitous shear-sense indicators. There, competent objects (i.e. remnants

of compositional layering or dykes) are asymmetrically boudinaged within more ductile metapelites. On the main D81 road, lenses of meta-rhyolite intrusions display a sigmoidal internal S_2 foliation consistent with an overall top-to-the-NE sense of shear (Fig. 5e). In the bulk of those intrusions, the S_1/L_1 fabrics are involved in recumbent isoclinal D_2 folds with the development of a S_2 crenulation cleavage with a low ($\sim 20^\circ$) westward dip (Fig. 5f).

The reactivation of pre-existing shallow-dipping discontinuities also occurs in the vicinity of the intrusive contact between the western granitic complex and the volcano-sedimentary country-rocks. Such is the case at the Punta di Mignola (see Fig. 2 for location) where top-to-the-NE mylonites occur as a 20 m-thick shear zone that overprints the primary intrusive contact that is preserved as partially transposed granitic dykes asymmetrically boudinaged within the metasedimentary matrix (Fig. 5g). Despite the preservation of primary intrusive boundaries, their igneous protolithic texture is hardly recognizable. Here again, shear bands present an apparent flat/reverse geometry.

Due to the scarcity of index minerals in most of the rocks, the physical conditions that prevailed during top-to-the-NE deformation are not easy to retrieve continuously. In the most mafic lithologies, top-to-the-NE shearing can be associated with the boudinage or even the truncation of blue-amphiboles (Figs. 5h and 5i). In these rocks, a second generation of small-scale blue-amphibole can occur during this D_2 stretching but gaps between the fragments of large blue-amphiboles and along the top-to-the-NE shear bands are mostly recorded during crystallization of chlorite (Fig. 5i). In more felsic granitic lithologies that represent most of the Tenda massif, the top-to-the-NE mylonites of the ETSZ are made of quartz, albite and phengite that are sometimes associated with chlorite and pyrite (Fig. 5j).

4.2. Description and distribution of top-to-the-SW deformation (D_1)

Top-to-the-SW deformation is located in the same structural level as rocks only affected by S_1 that wrap unstrained Variscan magmatic rocks (compare Figs. 2b and 4a). Study of top-to-the-SW deformation has therefore been restricted to the exploration of the lower parts of the dome along sections showing the upward development of S_1 and the associated simple shear markers at the expense of the Variscan magmatic protoliths.

In the core of the less deformed domains (Fig. 2b), the inherited Variscan magmatic paragenesis including K-feldspar, plagioclase, quartz, biotite and hornblende is metastably preserved within incipient coronitic reactions. Among the most obvious reactions, hornblende is statically transformed to blue-amphibole. These domains, which display no perceptible deformation, are surrounded by rocks affected by S_1/L_1 tectonites. While the base of this strain gradient is dominated by less non-coaxial deformation, it rapidly evolves toward top-to-the-SW simple shear deformation marked by the development of a single set of shear bands (Figs. 4a and 5a). Besides, this strain gradient is further complicated by more local strain localization, even at the scale of a single outcrop, as exemplified by the base of the ETSZ, along the main D81 road (Figs 6a, 6b, 6c). Here, top-to-the-SW deformation can occur as localized shear bands that overprint a volume of rocks devoid of visible deformation (Fig. 6a). In this case, S_1 only develops near shear bands and passes laterally to the undeformed protolith. Shear bands can also work as limits between isotropic granite on one side and oblique S_1 on the other side. Those observations suggest that prior or during formation of S_1 , D_1 deformation started along discrete precursors in the form of shear bands. In this case, the flat orientation of shear bands appears controlled by the orientation of the precursor. Within the most deformed volumes of this outcrop, D_1 deformation is more distributed, with regularly spaced top-to-the-SW shear bands with S_1 sigmoidal (Fig. 6b). In this context, typical spacing between shear planes varies between 1 and 3 cm, depending on the size of magmatic K-feldspar clasts, and they dip to the west. S_1 appears generally flat, highlighted by stretched quartz ribbons

alternating with feldspar ribbons. The most deformed parts of this outcrop are concentrated in a very gently dipping high strain shear band showing a fast lateral widening from less than 1cm up to ~10 cm (Fig. 6c). Near the core of the ETSZ, D_1 deformation is occasionally preserved from D_2 as small-scale lens-shaped domains that often show internal mylonitic deformation (Fig. 6d). There, both the S_1 and the top-to-the-SW shear planes show evidence of passive rotation as they now dip eastward. Field observations on those structural levels indicate that when both top-to-the-SW and -NE shear criteria coexist, the first ones are always crosscut by the second ones (Fig. 6e).

In granitic lithologies, physical conditions that prevailed during top-to-the-SW deformation are generally not easy to retrieve. Blue-amphiboles sporadically develop either along the S_1 or along the top-to-the-SW shear bands. Conversely, more mafic lithologies, top-to-the-SW shearing appears clearly associated with the growth of synkinematic blue-amphiboles (Fig. 6d).

5. Mapping the state of finite deformation within the Tenda granites

5.1. Approach and methodology

Variably strained granite constitutes the dominant lithology in the east of the Tenda massif. Granitic protoliths appear well suited to determine local and regional variations in the magnitude of deformation, either qualitatively or quantitatively. Quantitative estimates as retrieved from image analysis (e.g. Launeau and Robin, 2005) or anisotropy of magnetic susceptibility analysis (e.g. Borradaile and Jackson, 2010) are valuable in such contexts but are often local in extent. To put constraints on the large-scale pattern of deformation ($> 100 \text{ km}^2$), exhaustive mapping of the state of finite deformation has been applied qualitatively as

described in Huet et al. (2009), Charles et al. (2011), and Laurent et al. (2015). Laurent et al. (2015) have shown a good correlation between this qualitative approach and quantitative methods.

As discussed above, the effects of D_1 and D_2 can be generally unambiguously separated. The intensity of the D_2 overprint, however, is locally so severe as to allow only the D_2 finite deformation pattern to be continuously mapped across the massif. Five strain grades were distinguished to describe the intensity of deformation across the protolith (Fig. 7). Geometric embedding of the different strain grades is presented on figure 8.

- Grade 0 corresponds to rocks that did not experience significant D_2 deformation, such as D_2 folds, S_2 , L_2 or top-to-the-NE simple shear indicators. Within this grade, rocks may be either coarse grained or mylonitized during D_1 . Thus, two sub-grades were distinguished. Grade 0s.s. corresponds to the undeformed protolith characterized by an equant fabric where blue-amphiboles can statically replace magmatic hornblende (Figs. 7a and 7b). Besides, grade 0W corresponds to any other rocks deformed during D_1 and showing typical features such as top-to-the-SW shear bands, S_1 or L_1 but no trace of D_2 deformation (Fig. 7c).

- Grade 1 characterizes rocks that experienced incipient D_2 deformation superimposed on D_1 features. In the field, typical examples are outcrops showing D_2 folds associated with two different fabrics: a steep and folded S_1 and a sub-horizontal S_2 crenulation cleavage (Figs. 3b, 3c, 3d, 5f and 7d). There, S_1 can be either gently folded or strongly crenulated. However, it shows in any case a pure shear component corresponding to subvertical shortening. D_1 top-to-the-SW shear bands may be locally preserved associated with the sigmoidal character of S_1 . In those cases, top-to-the-SW shear bands and S_2 planes have comparable orientations, suggesting that shear bands represented weak zone where S_2 lately localized.

- Grade 2 is characterized by unequivocal top-to-the-NE simple shear indicators associated to asymmetric boudins and particularly to shear bands while D_2 folds remain the

main feature. In some cases, S_2 crenulation planes are used as shear bands (Fig. 7e). In this case, S_1 is comprised and between S_2 planes, indicating top-to-the-NE shearing. In other cases, newly formed top-to-the-NE shear bands affect a sigmoidal S_2 (Fig. 7f). Angle between shear planes and foliation is about 60° .

Increasing degree of simple shear from grades 2 to 4 is then recorded by an overall decrease of both spacing between top-to-the-NE shear bands and the angle between the S_2 and the shear bands, responsible for the progressive transposition of D_1 structures (Figs. 7f to 7h).

- Transition between grades 2 and 3 is marked by the occurrence of top-to-the-NE shear bands that are regularly spaced between 5 and 10 cm, affecting a sigmoidal S_2 (Fig. 7f). The angle between shear bands and foliation can be bracketed between 30 and 60° . D_2 folds are more rarely preserved, and shear bands dominate.

- Transition between grades 3 and 4 is marked by a decreasing angle between top-to-the-NE shear bands and S_2 , beneath 30° (fig. 7g). This flattening of mylonites is responsible for nearly complete obliteration of D_2 folds, rarely still visible, and the regular spacing of shear bands near 1 cm.

- Grade 4 corresponds to the most deformed volumes, being characterized by occurrence of ultramylonites and by the almost total transposition of S_2 (Fig. 7h).

The increase of finite strain intensity is therefore associated with a progressive increase of the simple shear component of deformation, from vertical shortening and folding of S_1 to top-to-the-NE simple shear deformation. In this frame, the angle between top-to-the-NE shear planes and S_2 progressively closes with the progressive reworking of the remaining D_1 structures (Fig. 8).

5.2. Grade distribution

Distribution of the deformation over the entire eastern Tenda massif was retrieved by applying this high-resolution strain intensity scale. In the cases of mixed grade for a same outcrop, overprinted structures (grade n-1) are generally boudinaged between anastomosed zones of overprinting structures (grade n). Local variations in deformation grade thus appear as snapshots of a continuum of deformation. This observation can be made at different scales (Fig. 9). To clarify the regional pattern, local mixed-grade stations are qualified with a single grade. Grade n-1 is chosen if it clearly represents a large part of the outcrop. For equal-proportion between grades n and n-1, grade n is chosen because it better represents the finite structure, as it appears as the latest phase. The resulting map, comprising 617 outcrops homogeneously distributed within the granitic complex, is shown on figure 10a. Four cross sections normal to the regional foliation and parallel to stretching are presented on figure 10b.

At the massif scale, the map shows that D_2 affects a large part of the mapped area, except for the Casta granodiorite, which remains isotropic or only affected by D_1 (cross sections BB' and DD' on Fig. 10). 75 % of the investigated outcrops are affected by D_2 . Among those outcrops, 28 % are represented by grade 1 and 30 % by grade 2. Grades 3 and 4 domains are of secondary importance, 24 % to 18 % respectively, a trend nicely indicating a localization of deformation with increasing finite strain. This distribution is shown both on the map and in the cross sections where the deformed zones become narrower with increasing strain intensity. At first-order, strain intensity clearly increases from west to east, toward the ETSZ (cross sections BB', CC', DD' on Fig. 10) and the SSZ (cross section AA' on Fig. 10). At a smaller scale, deformation zones also occur as isolated shear zones of moderate intensity (grade 2) in poorly deformed zones (grade 1) near Mt Genova (cross section BB' on Fig. 10). Similarly, isolated high-intensity shear zones (grade 4) occur in moderately deformed volumes within the ETSZ (cross section CC' on Fig. 10). The overall 3-D structure of the massif thus appears as a heterogeneous shear zone showing: i) a first-order strain gradient

toward the ETSZ and the SSZ, ii) isolated strain gradients at a smaller scale. As observed at different scales (Fig. 9), zones dominated by structures related to grade n-1 appear as lenses or boudins between zones affected by grade n (Fig. 10).

6. Discussion

Structural studies on the Tenda massif have been so far conducted on its eastern part, the ETSZ, an excellent example of polyphased contact accommodating both burial and exhumation of continental material (Daniel et al., 1996; Brunet et al., 2000; Gueydan et al., 2003; Molli and Tribuzio, 2004; Molli et al., 2006; Maggi et al., 2012; 2014; Rossetti et al., 2015). However, behavior of a large-scale continental crust portion and particularly strain accommodation of subduction and exhumation cannot be explored only by the detailed study of the major contacts where deformation is particularly concentrated. Our results show that deformation of the Tenda massif was not restricted to the ETSZ. Field results show the distribution of retrograde deformation related to exhumation and the pattern of partial to total transposition of the inherited top-the-SW fabric. One of the most important observation is that the later NE-directed (exhumation-related) structures are pervasively recorded across the entire massif. In the lights of these results, this section discusses the variation of strain distribution during burial and exhumation as well as the factors responsible for strain localization. Regional implications are presented first.

6.1. A review of the Tenda massif finite structure: regional implications on the model of subduction/exhumation

The various models of subduction and exhumation of the Tenda massif depend on: 1) the uncertain attribution of some structures to one of the different tectonic phases; 2) the focus of previous work on the ETSZ, that led to different descriptions of finite structure and patterns of strain distribution (Daniel et al., 1996; Brunet et al., 2000; Gueydan et al., 2003; Molli and Tribuzio, 2004; Molli et al., 2006; Maggi et al., 2012; 2014; Rossetti et al., 2015); 3) the scattering and various interpretation of chronological constraints. We discuss these three points in the next subsections.

6.1.1. To which deformation stage folds and crenulated structure belong?

Folds affecting S_1 and associated to crenulation cleavage is differentially explained by the models presented in the section 2.2. They can develop in coaxial domain of deformation between top-to-the-NE shear zones during exhumation in greenschist-facies conditions (Jolivet et al., 1990; 1991; Daniel et al., 1996) or can occur in low-strained domains in between zones of localized top-to-the-SW shearing in blueschist-facies conditions (Molli et al., 2006; Rossetti et al., 2015). Our study shows that: 1) top-to-the-SW shearing can display different amounts of strain localization from distributed to strongly localized in a same outcrop without developing any crenulation (Fig. 6). 2) On the contrary, crenulation is rather localized where top-to-the-NE shearing occurs at outcrop scale (Figs. 5e, 5f, 7e, 7f and 9). This is also true for larger scales. For example, undulations of vertical S_1 in the Mt Genova (Fig. 3a) occur in between two top-to-the-NE shear zones (Fig. 10). 3) This crenulation (S_2) is associated to retrogression of blue-amphiboles, which developed in S_1 , into chlorite, suggesting that S_1 and S_2 did not develop in the same P - T conditions (Fig. 3d). For these reasons, we followed the model of Jolivet et al. (1990; 1991) and Daniel et al. (1996), according to which the crenulated structures are rather linked to D_2 .

One important observation is that these folds are also observed in the eastern limb of the Cap Corse antiform (Fig. 1; Jolivet et al., 1991). There, the NE-verging recumbent folds are observed from a few centimeters to hundreds of meters and they are later sheared by top-to-the-E shear zones and cut by the brittle detachment observed near the northernmost tip of Cap Corse (Macinaggio detachment, Fig. 1). Thus, the two antiforms, namely the Tenda massif and the Cap Corse, show the same general asymmetry with a localized east-dipping shear zone at the top with folds overturned to the east underneath. This shows that these folds are not restricted to the Tenda massif but are instead characteristic of the late architecture of Alpine Corsica, including the Tenda massif and the Schistes Lustrés nappe as two separate crustal-scale boudins formed during extension (Daniel et al., 1996).

6.1.2. Global strain distribution in the Tenda massif

In the frame of an orogenic wedge, the subduction/exhumation style and mechanism of continental material largely depends on the rheology of the subducting continental crust. To understand it, a careful structural description of pattern of strain distribution is needed. The fact that studies on those questions focused on the ETSZ leads sometimes to an apparent rigidity of the Tenda massif. Moreover, since the ETSZ is where the final increments of deformation toward the brittle field localized (Molli et al., 2006; Maggi et al., 2012; 2014), this focus leads to an overestimation of the relevance of brittle post-orogenic reactivation. Instead, our study shows that ductile markers of D_1 (S_1 , L_1 , top-to-the-SW shear criteria) and D_2 (S_2 , L_2 , top-to-the-NE shear criteria) can be encountered from the ETSZ to the Ostriconi fault, except a few domains where the protolith remains unformed (notably the Casta granodiorite). In more details, the structural analysis of the Tenda massif (Fig. 5a) reveals a finite structure characterized by a ductile NW-SE dome of D_2 features (D_2 folds, S_2 , L_2 , and

top-to-the-NE shear criteria) that rework D_1 structures (S_1 , L_1 and top-to-the-SW shear criteria). The relatively more deformed carapace mantles a less deformed core which can locally preserves undisturbed D_1 structures. From west to east, this dome is crosscut by late top-to-the-NE narrow and localized shear zones that evolve toward a brittle behavior (Figs. 5a, 5c and 5d). We then emphasize the fact that D_2 , top-to-the-NE shearing, and switch of the ETSZ from a thrust to a detachment were mostly recorded in the ductile field, and pervasively affect the finite structure of the Tenda massif.

6.1.3. Geodynamic context for the exhumation of the Tenda massif

A major controversy exists about the respective part of regional compression and extension in the exhumation of the Tenda massif (see debate in Molli et al., 2006). The rather cold path of the exhumation favors a first step of syn-orogenic exhumation (Molli and Tribuzio, 2004; Molli et al., 2006; Maggi et al., 2012), also envisaged for more external continental units (i.e. the Popolasca unit for example; Malasoma and Marroni; 2007; Di Rosa et al., 2016). Instead, the end is clearly recorded during post-orogenic extension in greenschists-facies to brittle conditions, affecting the whole nappe stack (Molli and Tribuzio, 2004; Molli et al., 2006; Malasoma and Marroni; 2007; Maggi et al., 2012; Di Rosa et al., 2016). Syn-orogenic exhumation may involve a thrust at the base and a detachment (or exhumation fault) at the top of the exhumed units (e.g. Chemenda et al., 1995; Raimbourg et al., 2005), whereas post-orogenic extension will rework the entire nappe stack. In the case of the Tenda massif, Molli et al. (2006) concluded that syn-orogenic exhumation has been responsible for most of the exhumation below the detachment situated between the Schistes Lustrés nappe and the Nappe Supérieures and accommodated by a thrust at the base of the Tenda massif. Unfortunately, this structural architecture cannot be proven because the basal

contact of the Tenda massif does not crop out anywhere (Fig. 5a). Only chronological constraints permit to discuss this problem (see Fig. 11).

A regional shift from compression to extension has been widely recognized from the Gulf of Lion to the Calabrian Arc in connection with the southeastward retreat of the Apennine slab (e.g. Malinverno and Ryan, 1986; Doglioni, 1991; Gueguen et al., 1998; Jolivet et al., 1998; 2008; 2015; Séranne, 1999; Brunet et al., 2000; Rossetti et al., 2001; Speranza et al., 2002; Faccenna et al., 2004). As attested by evidences of compression in Late Eocene-Early Oligocene sedimentary rocks of Sardinia (Faccenna et al., 2002) and syn-rift sediments of Middle Oligocene to Aquitanian age (23-20 Ma) in the Gulf of Lion (e.g. Séranne, 1999; Oudet et al., 2010; Jolivet et al., 2015), the shift from compression to extension occurred in the Oligocene probably near the Late Rupelian (34-28 Ma). Similarly, syn-rift sediments formed during Late Oligocene-Aquitanian times in Corsica and Sardinia, prior to their anticlockwise rotation as a rigid block during the spreading of the Liguro-Provençal Sea (e.g. Faccenna et al., 2002; Speranza et al., 2002; Ferrandini et al., 2003).

Top-to-the-SW shearing in the Tenda massif is constrained between early Eocene for its prograde increments (U-Pb on acmite-phengite-coatings, Maggi et al., 2012) and Early Oligocene for its waning stage (Brunet et al., 2000; Rossetti et al., 2015). The end of top-to-the-SW shearing at this time is coherent with Late Eocene-Early Oligocene *HP-LT* peak ages in the more buried levels of the Schistes Lustrés nappe (Martin et al., 2011; Vitale Brovarone and Herwartz, 2013) and with the post-Bartonian *HP-LT* thrusting in the external domains (Bonnal et al., 1973; Bézert and Caby, 1988; Jourdan, 1988; Egal and Caron, 1989; Vitale Brovarone et al., 2013) that all constrain the beginning of syn-orogenic exhumation. Recent $^{40}\text{Ar}/^{39}\text{Ar}$ ages on Alpine shear zones developing in the eastern Variscan basement argue for a major shearing during *HP-LT* peak at 37-35 Ma, before exhumation near 32 Ma (Di Vincenzo et al., 2016). As a conclusion, the syn-orogenic exhumation of the relatively internal domain

of Alpine Corsica (i.e. Schistes Lustrés nappe and Tenda massif) is narrowly constrained between ~38-32 Ma that is a maximum limit for HP-LT peak in the Schistes Lustrés nappe and for thrusting in the external domain, and ~32-30 Ma, during exhumation of the entire Alpine Corsica, synchronously with the regional compression-extension switch. This could happen when Corsica passed from an east-dipping Alpine subduction to the back-arc domain of the west-dipping Apennine subduction system near 35-30 Ma (e.g. Molli and Malavieille, 2011; Malusà et al., 2015).

The regional compression/extension switch is recorded by Rb-Sr (Rossetti et al., 2015) and $^{40}\text{Ar}/^{39}\text{Ar}$ (Brunet et al., 2000) ages on phengite affected by D₁ and D₂ in both the Schistes Lustrés and the Tenda units. In the Tenda massif, crystallization or re-crystallization of relatively substituted D₁ phengite ($\text{Si}^{4+} \sim 3.5$ a.p.f.u) during deformation in samples characterized by top-to-the-W kinematics are dated between 32 and 27 Ma by the Rb-Sr method while the $^{40}\text{Ar}/^{39}\text{Ar}$ one gives a minimum age of 35 Ma. The end of D₂ deformation is recorded in highly deformed samples characterized by top-to-the-NE kinematics and relatively less substituted phengite ($\text{Si}^{4+} \sim 3.3$ a.p.f.u) at 20 and 25 Ma by the Rb-Sr and $^{40}\text{Ar}/^{39}\text{Ar}$ methods, respectively. The same ages are recorded in strongly retrogressed samples of the Schistes Lustrés nappe during D₂. The systematic difference of 5 Ma observed between the two methods questioned their ability to precisely date the timing of deformation. Accordingly, the compression/extension switch is differentially constrained near 27 Ma (Rossetti et al., 2015) or 32 Ma (Brunet et al., 2000). Further investigation are needed to precisely constrain the absolute timing of deformation relative to the different events. However, the two datasets indicate that the relative time between the ends of the two events is about ~7 Ma, showing that post-orogenic movements probably accommodated a non-negligible part of the exhumation.

6.2. Styles of deformation and strain localization during burial and exhumation

Many studies on strain localization and shear zone development have focused on cases where P - T conditions remained constant through deformation. The Tenda massif offers instead an example where the distribution of deformation can be studied through variable P - T conditions during a full burial/exhumation cycle.

6.2.1. Burial (D_1 phase)

The presence of a Mesozoic to Eocene sedimentary cover resting unconformably on the Tenda massif (Jourdan, 1988; Rossi et al., 2001; Vitale Brovarone et al., 2013) shows that it was near the surface just before its burial, as the Alpine events in external units started during the Bartonian (40 Ma; Bonnal et al., 1973; Bézert and Caby, 1988; Jourdan, 1988; Egal and Caron, 1989; Vitale Brovarone et al., 2013). Then, blueschist-facies conditions are recorded during D_1 as attested by crystallization of blue-amphibole along S_1 (Figs. 3c and 3d) and top-to-the-SW shear bands (e.g. Molli et al., 2006). At large-scale, blue-amphibole occurs principally in the Mt Asto granite, within the ETSZ (Fig. 2), suggesting that its occurrence depends in part on chemical composition of the protoliths and that it is favored by deformation. Despite this observation, blue-amphibole also occurs statically in non-deformed facies of the Casta granodiorite (Fig. 7b). Within the Mt Genova granite, two occurrences have been observed (Fig. 2a), the first one along the main D81 road, in a fine-grained facies strongly localizing deformation and the second one in micaschists of the Punta di Mignola shear zone (Figs. 2a and 5g). Further south of our study area, blue-amphibole also occurs in the westernmost parts of the Tenda massif (Near Urtaca, Jourdan, 1988; Molli and Tribuzio, 2004). The most reliable peak P - T conditions constraints are provided by petrological studies

in peculiar lithologies, away from the ETSZ, in the gabbroic complex of Bocca di Tenda, south of the Tenda massif. Here, rhyolite with jadeite-bearing aegirine, Na-amphibole and celadonite-rich phengite (Tribuzio and Giacomini, 2002) and gabbro with Al-poor hornblende and celadonite-rich phengite (Molli and Tribuzio, 2004) constrain peak metamorphism near 1 GPa and 450 °C. Similar results (1.2 GPa, 400 °C) are obtained in the ETSZ with a peculiar clinopyroxene + rutile assemblage (Maggi et al., 2012). All those results suggest that the Tenda massif recorded a global blueschists-facies condition during burial, and remained a relatively coherent unit.

Even if the early brittle structures relative to burial have been completely overprinted by the later ductile deformation, we propose that first stage of deformation was probably localized along brittle thrust precursor, as observed in the Corte slices further south (Jolivet et al., 1990). The following ductile HP-LT D_1 deformation (S_1 , L_1 , top-to-the-SW shearing) is on the contrary distributed at large scale, affecting a large volume of the Tenda massif, from the core where it can preserve its original geometry, to the top of the dome where it is lately overprinted (Figs. 5a and 12). Distributed deformation is evidenced by gneissic texture during D_1 (Maggi et al., 2012; 2014). At the scale of outcrop, we do not find any evidences of structures related to the brittle-ductile transition for D_1 . However, in the less deformed volume, we found that S_1 developed only along top-to-the-SW shear bands and passes laterally to the undeformed protolith (Fig. 6a). Those observations suggest that D_1 deformation started along discrete precursors in the form of shear bands prior or during formation of S_1 . It is coherent with studies focusing on the nucleation of shear zones where the amount of later overprint is limited. They demonstrate that shear zones nucleate from discrete precursors such as brittle fractures (e.g. Mancktelow and Pennacchioni, 2005; Raimbourg et al., 2005; Goncalves et al., 2012; 2016; Sullivan et al., 2013), lithological contacts (e.g. Pennacchioni and Mancktelow, 2007) or simply microscopic heterogeneities at

mineral boundaries (e.g. Ingles et al., 1999). Finally, the top-to-the-SW shear bands are lately overprinted by brittle fractures, making the nature of the precursor uncertain. However, all those points argue for a first step of strain delocalization within the Tenda massif during burial and its first crossing through the brittle-ductile transition from surface up to blueschists-facies conditions.

6.2.2. Exhumation (D_2 phase)

Structures related to exhumation were mostly recorded in greenschist-facies conditions, while blue-amphiboles were turned into chlorite in S_2 crenulation and top-to-the-NE shear bands (Figs. 3d and 5i). It seems reasonable that they developed from blueschists-facies conditions, in a D_1 - D_2 continuum deformation. Indeed, Molli et al. (2006) described a D_2 crenulation made of fine grains of Na-amphibole showing an outward decrease of Al, compatible with a decreasing pressure, between microlithons with large Na-amphiboles. We also find blue-amphibole recrystallization in the tails of older blue-amphibole sigma clasts compatible with top-to-the-NE shearing (see white square in Fig. 5i). The exhumation path can be constrained from ~1 GPa and 450 °C to < 0.5 GPa and 300-400°C (Tribuzio and Giacomini, 2002; Molli and Tribuzio, 2004; Molli et al., 2006). The last increments of exhumation are recorded in the brittle field (Fig. 5d) in a continuum deformation, as the directions of both ductile and brittle stretching are consistent (Molli et al., 2006).

Right after the burial/exhumation switch, the first stages of D_2 deformation are characterized by pervasive folding responsible for a general verticalization of S_1 envelope (Figs. 2c, 3e, 5a and 12). Folding is expressed in low-strain D_2 domain by gentle undulations of S_1 (Fig. 3a) or inversely strongly crenulated S_1 associated to a flat S_2 cleavage (Figs. 3b, 3c, 3d), compatible with a pure shear component of deformation corresponding to a subvertical

shortening. Larger-scale folds can be observed near Santo-Pietro-di-Tenda, where they affect the contact between the Tenda massif and the Schistes Lustrés nappe (Molli et al., 2006). Limbs of the Schistes Lustrés nappe incorporated into the ETSZ (see Punta di Cepo and Fornali Bay in Fig. 2) also suggest a folding of this contact. Irrespective of these variations in scale, this folding stage is observed throughout the Tenda massif affected by D₂. In agreement with Molli et al. (2006), we argue for a general strain delocalization at the time of D₂, responsible for the buckling and deactivation of the original thrust just before or after the D₁-D₂ switch (Fig. 12). Then, the strain intensity map shows that deformation is localized in narrower top-to-the-NE shear zones that locally develop thick mylonites and locally ultramylonites (Figs. 5, 7, 8 and 10). Furthermore, the transition from grade n-1 to n is marked by relatively poorly deformed boudins in a more deformed matrix (Fig. 9). Local variations in deformation grade thus appear as snapshots of a continuum of deformation that finally localized in the ETSZ and the SSZ, showing that D₂ deformation progressively localized during exhumation toward the two principal shear zones. This interpretation is further supported by the decrease of ⁴⁰Ar/³⁹Ar ages on phengites toward the ETSZ nappe from ~45 to 25 Ma (Brunet et al., 2000). Since the ductile structures are lately crosscut by NE-striking faults, the spatial and temporal relationships between the ETSZ and the SSZ are unclear. Further studies should better assess both brittle kinematics and provide radiometric chronological constraints also for the SSZ. In any case, the ETSZ and the SSZ are two different structures, the ETSZ developing at the top of the Tenda massif while the SSZ is structurally below and developed within the granitic complex (Fig. 10). Moreover, the two shear zones are characterized by different stretching directions, NE-SW for the ETSZ and NNE-SSW for the SSZ (Fig. 4a). Doming of the Tenda massif occurred during the progressive strain localization toward the ETSZ as the D₂ mylonites carapace near the top of the massif is itself domed and outcrops in the west (Fig. 5). Besides, late top-to-the-NE

narrow localized shear zones, showing a component of brittle deformation, always dip toward the northeast and are not affected by the doming. We suggest that this doming formed during regional extension near the brittle-ductile transition by a horizontal-axis rotation beneath the ETSZ where deformation progressively localized. The Tenda massif then record a re-localization of strain during its second crossing through the brittle-ductile transition zone, from strongly distributed at depth where exhumation began, to strongly localized in the upper levels of the crust, toward the brittle field (Fig. 12).

6.3. Internal factors for strain distribution during burial and exhumation

The two major shear zones of the Tenda massif are the ETSZ and the SSZ, where strain localizes toward lithological contacts between rocks of very contrasting rheology (Fig. 10). The fact that strain localizes in the ETSZ is comprehensible since the contact between the Tenda massif and the Schistes Lustrés nappe is both a zone of lithological/rheological contrast and a pathway for fluids (Maggi et al., 2012). In agreement with these authors, we find very heterogeneous second-order strain gradients in the ETSZ (see punctual observations of strain grades in Fig. 10). Maggi et al. (2012) explain it by alternating episodes of fluids-induced strain softening (feldspar-to-micas reaction) and strain hardening (neo-blastesis of K-feldspar) producing a rheological contrast that drives strain localization through a heterogeneous shear zone. But the internal lithological contacts are also preferentially used for strain localization. Indeed, most of those contacts are intensely deformed. This is true at the contact between: i) the granitic complex and the volcano-sedimentary sequence in the Punta di Mignola (Fig. 5g), ii) the Mt Genova and the Saleccia rhyolite in the SSZ (Figs. 5b and 10), or iii) the Mt Genova and the Casta granodiorite, east of Mt Genova (Fig. 10). This is in agreement with studies showing that shear zones nucleate on rheological heterogeneities such as brittle structures

(e.g. Mancktelow and Pennacchioni, 2005; Goncalves et al., 2012; 2016; Sullivan et al., 2013) or lithological contacts (e.g. Pennacchioni and Mancktelow, 2007; Rennie et al., 2013) where strain remained localized.

S_2 crenulation is well developed in these preferential zones for strain localization, particularly in the vicinity of the SSZ and the ETSZ, where S_2 fabrics are visible at landscape scales (Figs. 3f and 3g). Inversely, in less deformed zone, D_2 folds are rather characterized by S_1 undulations without crenulation development (Fig. 3a). The fact that top-to-the-SW shear bands are privileged structures where S_2 develops (see section 5 about strain grades) suggests that penetrative S_2 fabrics visible in the major shear zones of the ETSZ are domain where top-to-the-SW D_1 shear zones initially localized during burial. Lithological contact thus is a primordial factor driving strain localization, directly for burial and D_1 structures, and indirectly for exhumation and D_2 structures, through a structural control of primarily deformation zone born during D_1 .

If lithological contact is a driving structure for strain localization, accommodation of deformation largely varies in function of the protolith. The best example is the Casta granodiorite that remained globally undeformed as it resisted in a large scale strain shadow. The Mt Genova also remained relatively poorly deformed with only few narrow shear zones situated structurally below and above. Inversely, deformation pervasively and intensely affected the Saleccia rhyolite and the Mt Asto monzogranite. Maggi et al. (2012; 2014) described the Casta granodiorite as the protolith of the ETSZ. However, we rather prefer the geological outline of Rossi et al. (1994) that shows that the ETSZ developed in the Mt Asto monzogranite. Indeed, when poorly deformed, it is clear that the ETSZ developed in a relatively felsic facies (Fig. 6a) compared to the Casta granodiorite that is more basic and characterized by a higher amount of hornblendes (Fig. 7a). Except for the Saleccia rhyolite that can deform easily due to its rather fine-grained matrix, the fact that plutonic granitoids

behaved differently during deformation suggests that deformation mechanisms depend on the composition of the protolith (X) as an internal parameter together with external parameters such as fluids, *P-T* conditions, associated with mineral transformations, as suggested for the Tenda massif (Gueydan et al., 2003; Maggi et al., 2014) or other granites-related shear zone (e.g. Goncalves et al., 2012; 2016; Oliot et al., 2010; 2014).

7. Conclusion

This structural study is based on a large dataset collected over most of the Tenda massif, providing new insights into its history from burial (D_1) to exhumation (D_2):

- The first order finite structure of the Tenda massif is characterized by a NW-SE dome of D_2 related structures (D_2 folds, L_2 , S_2 and top-to-the-NE shear criteria) mantling a less deformed core where D_1 structures (L_1 , S_1 and top-to-the-SW shear criteria) remain preserved in their initial position.

- Field evidence demonstrates that both D_1 and D_2 structures affect the whole massif. Although D_1 structures can still be encountered in their initial position, they are generally reworked by D_2 deformation.

- Burial occurred from surface, where localized deformation was concentrated along thrusts at the top of continental material (visible further south in Corte slices), until blueschist-facies conditions where deformation was rather distributed, affecting both continental and oceanic material.

- Exhumation probably started in a syn-orogenic context or simply at peak-metamorphic conditions, with delocalization of thrusting from the ETSZ toward the more external domains. D_2 deformation then occurred firstly distributed through a general folding of previous structures, even the original thrust between the Tenda massif and the overlying

Schistes Lustrés nappe. Through the switch from syn-orogenic to post-orogenic exhumation, D₂ localized in a set of large-scale shear zones including notably the ETSZ, located at the roof of the dome between continental and oceanic material, and the SSZ, an internal shear zone described for the first time. The general doming of D₂ structures occurred during this strain localization. The final increments are localized in the ETSZ and scattered in the rest of the massif through narrow semi-brittle shear zones, not affected by doming.

- Lithological contacts were site of preferred localization, directly for burial and D₁ structures, and indirectly for exhumation and D₂ structures, through a structural control of primarily deformation zone born during D₁. The deformation style and the tectonic coupling between continental and oceanic material seem to have been mostly controlled by external parameters, notably depth (i.e. *P-T* conditions) in the wedge. However, the strain intensity map shows that the amount of deformation depends at least in part on the protolith nature that can, through an interaction with *P-T*-fluids conditions, have a strong impact on the deformation style and the rheology of continental material.

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References

- Agard, P. & Vitale-Brovarone, A. (2013). Thermal regime of continental subduction: The record from exhumed HP-LT terranes (New Caledonia, Oman, Corsica). *Tectonophysics* 601, 206–215.
- Beccaluva, L., Ohnenstetter, D. & Ohnenstetter, M. (1981). K-Ar age determination on some Tethyan ophiolites. *Rendiconti della Societa Italiana di Mineralogia e Petrologia* 37, 869–880.
- Bellahsen, N., Jolivet, L., Lacombe, O., Bellanger, M., Boutoux, A., Garcia, S., Mouthereau, F., Le Pourhiet, L. & Gumiaux, C. (2012). Mechanisms of margin inversion in the external Western Alps: Implications for crustal rheology. *Tectonophysics* 560–561, 62–83.
- Bellanger, M., Bellahsen, N., Jolivet, L., Baudin, T., Augier, R. & Boutoux, A. (2014). Basement shear zones development and shortening kinematics in the Ecrins Massif, Western Alps. *Tectonics* 33, 84–111.
- Bézert, P. & Caby, R. (1988). Sur l'âge post-bartonien des événements tectono-métamorphiques alpins en bordure orientale de la Corse cristalline (Nord de Corte). *Bulletin de la Societe Geologique de France* 4, 965–971.
- Bonnal, M., Parsy, A., Priou-Lacazedieu, A. & Durand-Delga, M. (1973). Sur la structure de la Balagne sédimentaire (Corse). *Comptes Rendus de l'Academie des Sciences de Paris* 276, 1949–1952.

- 898
- 899 Borradaile, G. J. & Jackson, M. (2010). Structural geology, petrofabrics and magnetic fabrics
900 (AMS, AARM, AIRM). *Journal of Structural Geology* 32, 1519–1551.
- 901
- 902 Brunet, C., Monié, P., Jolivet, L. & Cadet, J. P. (2000). Migration of compression and
903 extension in the Tyrrhenian Sea, insights from Ar-40/Ar-39 ages on micas along a transect
904 from Corsica to Tuscany. *Tectonophysics* 321, 127–155.
- 905
- 906 Caron, J. M. (1994). Metamorphism and deformation in Alpine Corsica. *Schweizerische*
907 *Mineralogische Und Petrographische Mitteilungen* 74, 105–114.
- 908
- 909 Caron, J. M., Kienast, J. R. & Triboulet, C. (1981). High-pressure-low-temperature
910 metamorphism and polyphase alpine deformation at Sant'Andrea di Cotone (eastern Corsica,
911 France). *Tectonophysics* 78, 419–451.
- 912
- 913 Caron, J. M. & Péquignot, G. (1986). The transition between blue-schist and lawsonite
914 bearing eclogites on the example of corsican metabasalt. *Lithos* 19, 205–218.
- 915
- 916 Cavazza, W., DeCelles, P. G., Fellin, M. G. & Paganellin, L. (2007). The Miocene Saint-
917 Florent Basin in northern Corsica: stratigraphy, sedimentology, and tectonic implications.
918 *Basin Research* 19, 507–527.
- 919
- 920 Charles, N., Gumiaux, C., Augier, R., Chen, Y., Zhu, R. & Lin, W. (2011). Metamorphic
921 Core Complexes vs. synkinematic plutons in continental extension setting: Insights from key
922 structures (Shandong Province, eastern China). *Journal of Asian Earth Sciences* 40, 261–278.

- 923
- 924 Chemenda, A. I., Mattauer, M., Malavieille, J. & Bokun, A. N. (1995). A mechanism for syn-
- 925 collisional rock exhumation and associated normal faulting: Results from physical modelling.
- 926 Earth and Planetary Science Letters 132, 225–232.
- 927
- 928 Cohen, C. R., Schweickert, R. A. & Odom, A. L. (1981). Age of emplacement of the schistes
- 929 Lustrés nappe, Alpine Corsica. Tectonophysics 73, 267–283.
- 930
- 931 Daniel, J. M., Jolivet, L., Goffé, B. & Poinssot, C. (1996). Crustal-scale strain partitioning:
- 932 footwall deformation below the Alpine Oligo-Miocene detachment of Corsica. Journal of
- 933 Structural Geology 18, 41–59.
- 934
- 935 De Wever, P., Danelian, T., Durand-Delga, M., Cordey, F. & Kito, N. (1987). Datations des
- 936 radiolarites post-ophiolitiques de Corse alpine à l’aide des radiolaires. Comptes Rendus de
- 937 l’Academie des Sciences de Paris 305, 893–900.
- 938
- 939 Di Rosa, M., De Giorgi, A., Marroni, M. & Vidal, O. (2016). Syn-convergence exhumation of
- 940 continental crust: evidence from structural and metamorphic analysis of the Monte Cecu area,
- 941 Alpine Corsica (Northern Corsica, France). Geological Journal.
- 942
- 943 Di Vincenzo, G., Grande, A., Prosser, G., Cavazza, W. & DeCelles, P. G. (2016). ^{40}Ar – ^{39}Ar
- 944 laser dating of ductile shear zones from central Corsica (France): Evidence of Alpine (middle
- 945 to late Eocene) syn-burial shearing in Variscan granitoids. Lithos 262, 369–383.
- 946

- Doglioni, C. (1991). A proposal for the kinematic modelling of W-dipping subductions - possible applications to the Tyrrhenian-Apennines system. *Terra Nova* 3, 423–434.
- Durand-Delga, M. (1984). Principaux traits de la Corse Alpine et corrélations avec les Alpes Ligures. *Memorie della Societa Geologica Italiana* 28, 285–329.
- Egal, E. & Caron, J. M. (1989). Structures de l'Eocène autochtone en Corse. *Comptes Rendus de l'Academie des Sciences de Paris* 309, 1431–1436.
- Faccenna, C., Piromallo, C., Crespo-Blanc, A., Jolivet, L. & Rossetti, F. (2004). Lateral slab deformation and the origin of the western Mediterranean arcs. *Tectonics* 23, TC1012.
- Faccenna, C., Speranza, F., D'Ajello Caracciolo, F., Mattei, M. & Giacomo, O. (2002). Extensional tectonics on Sardinia (Italy): insights into the arc–back-arc transitional regime. *Tectonophysics* 356, 213–232.
- Faure, M. & Malavieille, J. (1980). Les plis en fourreau du substratum de la Nappe des Schistes Lustrés de Corse. Signification cinématique. *Comptes Rendus de l'Academie des Sciences de Paris* 290, 1349–1352.
- Faure, M. & Malavieille, J. (1981). Etude structurale d'un cisaillement ductile : le charriage ophiolitique corse dans la région de Bastia. *Bulletin de la Societe Geologique de France* 23, 335–345.

- 971 Fellin, M. G., Vance, J. A., Garver, J. I. & Zattin, M. (2006). The thermal evolution of
 972 Corsica as recorded by zircon fission-tracks. *Tectonophysics* 421, 299–317.
 973
- 974 Ferrandini, J., Gattacceca, J., Ferrandini, M., Deino, A. & Janin, M.-C. (2003).
 975 Chronostratigraphie et paléomagnétisme des dépôts oligo-miocènes de Corse: implications
 976 géodynamiques pour l'ouverture du bassin liguro-provençal. *Bulletin de la Societe*
 977 *Geologique de France* 174, 357–371.
 978
- 979 Ferrandini, M., Ferrandini, J., Loÿe-Pilot, M.-D., Butterlin, J., Cravatte, J. & Janin, M.-C.
 980 (1998). Le Miocène du bassin de Saint-Florent (Corse): Modalités de la transgression du
 981 Burdigalien supérieur et mise en évidence du Serravallien. *Geobios* 31, 125–137.
 982
- 983 Fournier, M., Jolivet, L., Goffé, B. & Dubois, R. (1991). Alpine Corsica Metamorphic Core
 984 Complex. *Tectonics* 10, 1173–1186.
 985
- 986 Garfagnoli, F., Menna, F., Pandeli, E. & Principi, G. (2009). Alpine metamorphic and tectonic
 987 evolution of the Inzecca-Ghisoni area (southern Alpine Corsica, France). *Geological Journal*
 988 44, 191–210.
 989
- 990 Gibbons, W. & Horak, J. (1984). Alpine metamorphism of Hercynian hornblende granodiorite
 991 beneath the blueschist facies schistes-lustrés nappe of NE Corsica. *Journal of Metamorphic*
 992 *Geology* 2, 95–113.
 993

Goncalves, P., Oliot, E., Marquer, D. & Connolly, J. A. D. (2012). Role of chemical processes on shear zone formation: an example from the Grimsel metagranodiorite (Aar massif, Central Alps). *Journal of Metamorphic Geology* 30.

Goncalves, P., Poilvet, J.-C., Oliot, E., Trap, P. & Marquer, D. (2016). How does shear zone nucleate? An example from the Suretta nappe (Swiss Eastern Alps). *Journal of Structural Geology* 86, 166–180.

Gueguen, E., Doglioni, C. & Fernandez, M. (1998). On the post-25 Ma geodynamic evolution of the western Mediterranean. *Tectonophysics* 298, 259–269.

Gueydan, F., Leroy, Y. M., Jolivet, L. & Agard, P. (2003). Analysis of continental midcrustal strain localization induced by microfracturing and reaction-softening. *Journal of Geophysical Research* 108, 2064.

Huet, B., Labrousse, L. & Jolivet, L. (2009). Thrust or detachment? Exhumation processes in the Aegean: Insight from a field study on Ios (Cyclades, Greece). *Tectonics* 28, TC3007.

Ingles, J., Lamouroux, C., Soula, J.-C., Guerrero, N. & Debat, P. (1999). Nucleation of ductile shear zones in a granodiorite under greenschist facies conditions, Néouvielle massif, Pyrenees, France. *Journal of Structural Geology* 21, 555–576.

Jakni, B., Poupeau, G., Sosson, M., Rossi, P., Ferrandini, J. & Guennoc, P. (2000). Dénudations cénozoïques en Corse: une analyse thermochronologique par traces de fission sur apatites. *Comptes Rendus de l'Academie des Sciences de Paris* 331, 775–782.

- 1019
- 1020 Jolivet, L. et al. (1998). Midcrustal shear zones in postorogenic extension: Example from the
- 1021 northern Tyrrhenian Sea. *Journal of Geophysical Research* 103, 12123–12160.
- 1022
- 1023 Jolivet, L., Augier, R., Faccenna, C., Negro, F., Rimmelé, G., Agard, P., Robin, C., Rossetti,
- 1024 F. & Crespo-Blanc, A. (2008). Subduction, convergence and the mode of backarc extension in
- 1025 the Mediterranean region. *Bulletin de la Societe Geologique de France* 179, 525–550.
- 1026
- 1027 Jolivet, L., Daniel, J. M. & Fournier, M. (1991). Geometry and kinematics of extension in
- 1028 Alpine Corsica. *Earth and Planetary Science Letters* 104, 278–291.
- 1029
- 1030 Jolivet, L., Dubois, R., Fournier, M., Goffé, B., Michard, A. & Jourdan, C. (1990). Ductile
- 1031 extension in alpine Corsica. *Geology* 18, 1007–1010.
- 1032
- 1033 Jolivet, L., Gorini, C., Smit, J. & Leroy, S. (2015). Continental breakup and the dynamics of
- 1034 rifting in back-arc basins: The Gulf of Lion margin. *Tectonics* 34, 662–679.
- 1035
- 1036 Jourdan, C. (1988). Balagne orientale et massif du Tenda (Corse septentrionale): étude
- 1037 structurale, interprétation des accidents et des déformations, reconstitutions géodynamiques.
- 1038 Thèse de Doctorat, Orsay, Université de Paris-Sud.
- 1039
- 1040 Lacombe, O. & Jolivet, L. (2005). Structural and kinematic relationships between Corsica and
- 1041 the Pyrenees-Provence domain at the time of the Pyrenean orogeny. *Tectonics* 24, TC1003.
- 1042

- 1043 Lahondère, D. (1988). Le métamorphisme éclogitique dans les orthogneiss et les metabasites
1044 ophiolitiques de la région de Farinole (Corse). Bulletin de la Societe Geologique de France 4,
1045 579–585.
- 1046
- 1047 Lahondère, D. & Guerrot, C. (1997). Datation Sm-Nd du métamorphisme éclogitique en
1048 Corse alpine□: un argument pour l'existence au Crétacé supérieur d'une zone de subduction
1049 active localisée sous le bloc corso-sarde. Géologie de la France 3, 3–11.
- 1050
- 1051 Launeau, P. & Robin, P.-Y. F. (2005). Determination of fabric and strain ellipsoids from
1052 measured sectional ellipses—implementation and applications. Journal of Structural Geology
1053 27, 2223–2233.
- 1054
- 1055 Laurent, V., Beaudoin, A., Jolivet, L., Arbaret, L., Augier, R., Rabillard, A. & Menant, A.
1056 (2015). Interrelations between extensional shear zones and synkinematic intrusions: The
1057 example of Ikaria Island (NE Cyclades, Greece). Tectonophysics 651–652, 152–171.
- 1058
- 1059 Leloup, P. H., Arnaud, N., Sobel, E. R. & Lacassin, R. (2005). Alpine thermal and structural
1060 evolution of the highest external crystalline massif: The Mont Blanc. Tectonics 24, TC4002.
- 1061
- 1062 Maffione, M., Speranza, F., Faccenna, C., Cascella, A., Vignaroli, G. & Sagnotti, L. (2008). A
1063 synchronous Alpine and Corsica-Sardinia rotation. Journal of Geophysical Research 113,
1064 B03104.
- 1065
- 1066 Maggi, M., Rossetti, F., Corfu, F., Theye, T., Andersen, T. B. & Faccenna, C. (2012).
1067 Clinopyroxene-rutile phyllonites from the East Tenda Shear Zone (Alpine Corsica, France):

pressure-temperature-time constraints to the Alpine reworking of Variscan Corsica. *Journal of the Geological Society* 169, 723–732.

Maggi, M., Rossetti, F., Ranalli, G. & Theye, T. (2014). Feedback between fluid infiltration and rheology along a regional ductile-to-brittle shear zone: The East Tenda Shear Zone (Alpine Corsica). *Tectonics* 33, 253–280.

Malasoma, A. & Marroni, M. (2007). HP/LT metamorphism in the Volparone Breccia (Northern Corsica, France): evidence for involvement of the Europe/Corsica continental margin in the Alpine subduction zone. *Journal of Metamorphic Geology* 25, 529–545.

Malasoma, A., Marroni, M., Musumeci, G. & Pandolfi, L. (2006). High-pressure mineral assemblage in granitic rocks from continental units, Alpine Corsica, France. *Geological Journal* 41, 49–59.

Malavieille, J., Chemenda, A. & Larroque, C. (1998). Evolutionary model for Alpine Corsica: mechanism for ophiolite emplacement and exhumation of high-pressure rocks. *Terra Nova* 10, 317–322.

Malinverno, A. & Ryan, W. B. F. (1986). Extension in the Tyrrhenian Sea and shortening in the Apennines as result of arc migration driven by sinking of the lithosphere. *Tectonics* 5, 227–245.

- 1091 Malusà, M. G. et al. (2015). Contrasting styles of (U)HP rock exhumation along the Cenozoic
1092 Adria-Europe plate boundary (Western Alps, Calabria, Corsica). *Geochemistry Geophysics*
1093 *Geosystems* 16, 1786–7824.
- 1094
- 1095 Maluski, H. (1977). Application de la méthode $^{40}\text{Ar}/^{39}\text{Ar}$ aux minéraux des roches
1096 cristallines perturbées par les événements thermiques et tectoniques en Corse. *Bulletin de la*
1097 *Société Géologique de France* 19, 849–855.
- 1098
- 1099 Mancktelow, N. S. & Pennacchioni, G. (2005). The control of precursor brittle fracture and
1100 fluid-rock interaction on the development of single and paired ductile shear zones. *Journal of*
1101 *Structural Geology* 27, 645–661.
- 1102
- 1103 Martin, L. A. J., Rubatto, D., Brovarone, A. V. & Hermann, J. (2011). Late Eocene lawsonite-
1104 eclogite facies metasomatism of a granulite sliver associated to ophiolites in Alpine Corsica.
1105 *Lithos* 125, 620–640.
- 1106
- 1107 Mattauer, M., Faure, M. & Malavieille, J. (1981). Transverse lineation and large-scale
1108 structures related to Alpine obduction in Corsica. *Journal of Structural Geology* 3, 401–409.
- 1109
- 1110 Meresse, F., Lagabrielle, Y., Malavieille, J. & Ildefonse, B. (2012). A fossil Ocean–Continent
1111 Transition of the Mesozoic Tethys preserved in the Schistes Lustrés nappe of northern
1112 Corsica. *Tectonophysics* 579, 4–16.
- 1113

- 1114 Molli, G. (2008). Northern Apennine-Corsica orogenic system: an updated overview. In:
1115 Siegesmund, S., Fügenschuh, B., Froitzheim, N. (Eds.), Tectonic Aspects of the Alpine-
1116 Dinaride-Carpathian System. Geological Society, London, Special Publications 298, 413–442.
1117
- 1118 Molli, G. & Malavieille, J. (2011). Orogenic processes and the Corsica/Apennines
1119 geodynamic evolution: insights from Taiwan. *International Journal of Earth Sciences* 100,
1120 1207–1224.
1121
- 1122 Molli, G. & Tribuzio, R. (2004). Shear zones and metamorphic signature of subducted
1123 continental crust as tracers of the evolution of the Corsica/Northern Apennine orogenic
1124 system. Geological Society, London, Special Publications 224, 321–335.
1125
- 1126 Molli, G., Tribuzio, R. & Marquer, D. (2006). Deformation and metamorphism at the eastern
1127 border of the Tenda Massif (NE Corsica): a record of subduction and exhumation of
1128 continental crust. *Journal of Structural Geology* 28, 1748–1766.
1129
- 1130 Mouthereau, F., Tensi, J., Bellahsen, N., Lacombe, O., De Boisgrollier, T. & Kargar, S.
1131 (2007). Tertiary sequence of deformation in a thin-skinned/thick-skinned collision belt: The
1132 Zagros Folded Belt (Fars, Iran). *Tectonics* 26, TC5006.
1133
- 1134 Mouthereau, F., Watts, A. B. & Burov, E. (2013). Structure of orogenic belts controlled by
1135 lithosphere age. *Nature Geoscience* 6, 785–789.
1136

- 1137 Ohnenstetter, M., Ohnenstetter, D., Vidal, P., Cornichet, J., Hermitte, D. & Mace, J. (1981).
 1138 Crystallization and age of zircon from Corsican ophiolitic albitites: consequences for oceanic
 1139 expansion in Jurassic times. *Earth and Planetary Science Letters* 54, 397–408.
 1140
- 1141 Oliot, E., Goncalves, P. & Marquer, D. (2010). Role of plagioclase and reaction softening in a
 1142 metagranite shear zone at mid-crustal conditions (Gotthard Massif, Swiss Central Alps).
 1143 *Journal of Metamorphic Geology* 28, 849–871.
 1144
- 1145 Oliot, E., Goncalves, P., Schulmann, K., Marquer, D. & Lexa, O. (2014). Mid-crustal shear
 1146 zone formation in granitic rocks: Constraints from quantitative textural and crystallographic
 1147 preferred orientations analyses. *Tectonophysics* 612–613, 63–80.
 1148
- 1149 Oudet, J., Münch, P., Borgomano, J., Quillevère, F., Melinte-Dobrinescu, M. C., Demory, F.,
 1150 Viseur, S. & Corneé, J.-J. (2010). Land and sea study of the northeastern golfe du Lion rifted
 1151 margin: the Oligocene–Miocene of southern Provence (Nerthe area, SE France). *Bulletin de la*
 1152 *Société géologique de France* 181, 591–607.
 1153
- 1154 Pennacchioni, G. & Mancktelow, N. S. (2007). Nucleation and initial growth of a shear zone
 1155 network within compositionally and structurally heterogeneous granitoids under amphibolite
 1156 facies conditions. *Journal of Structural Geology* 29, 1757–1780.
 1157
- 1158 Raimbourg, H., Jolivet, L., Labrousse, L., Leroy, Y. & Avigad, D. (2005). Kinematics of
 1159 syneclogite deformation in the Bergen Arcs, Norway: implications for exhumation
 1160 mechanisms. In: Gapais, D., Brun, J. P., Cobbold, P. R. (Eds.), *Deformation Mechanisms,*

- 1161 Rheology and Tectonics: from Minerals to the Lithosphere. Geological Society, London,
1162 Special Publications 243, 175–192.
- 1163
- 1164 Ravna, E. J. K., Andersen, T. B., Jolivet, L. & De Capitani, C. (2010). Cold subduction and
1165 the formation of lawsonite eclogite – constraints from prograde evolution of eclogitized
1166 pillow lava from Corsica. *Journal of Metamorphic Geology* 28, 381–395.
- 1167
- 1168 Rehault, J.-P., Boillot, G. & Mauffret, A. (1984). The Western Mediterranean Basin
1169 geological evolution. *Marine Geology* 55, 447–477.
- 1170
- 1171 Rennie, S. F., Fagereng, A. & Diener, J. F. A. (2013). Strain distribution within a km-scale,
1172 mid-crustal shear zone: The Kuckaus Mylonite Zone, Namibia. *Journal of Structural Geology*
1173 56, 57–69.
- 1174
- 1175 Rossetti, F., Faccenna, C., Goffé, B., Monié, P., Argentieri, A., Funicello, R. & Mattei, M.
1176 (2001). Alpine structural and metamorphic signature of the Sila Piccola Massif nappe stack
1177 (Calabria, Italy): Insights for the tectonic evolution of the Calabrian Arc. *Tectonics* 20, 112–
1178 133.
- 1179
- 1180 Rossetti, F., Glodny, J., Theye, T. & Maggi, M. (2015). Pressure-temperature-deformation-
1181 time of the ductile Alpine shearing in Corsica: From orogenic construction to collapse. *Lithos*
1182 218–219, 99–116.
- 1183
- 1184 Rossi, P. et al. (2001). Carte Géologique de France (1/50 000), feuille de Santo-Pietro-Di-
1185 Tenda (1106). Bureau des Recherches Géologiques et Minières.

- 1186
- 1187 Rossi, P., Durand-Delga, M. & Cocherie, A. (1993). Caractère volcano-plutonique du
- 1188 magmatisme calco-alcalin composite d'âge Stéphanien supérieur-Permien inférieur en Corse.
- 1189 Comptes Rendus de l'Académie des Sciences de Paris 316, 1779–1788.
- 1190
- 1191 Rossi, P., Lahondère, J.-C., Lluch, D., Loÿe-Pilot, M.-D. & Jacquet, M. (1994). Carte
- 1192 Géologique de France (1/50 000), feuille de Saint-Florent (1103). Bureau des Recherches
- 1193 Géologiques et Minières.
- 1194
- 1195 Séranne, M. (1999). The Gulf of Lion continental margin (NW Mediterranean) revisited by
- 1196 IBS: an overview. In: Durand, B., Jolivet, L., Horvath, F., Séranne, M. (Eds.), Tertiary
- 1197 Extension within the Alpine Orogen. Geological Society, London, Special Publications 156,
- 1198 15–36.
- 1199
- 1200 Speranza, F., Villa, I. M., Sagnotti, L., Florindo, F., Cosentino, D., Cipollari, P. & Mattei, M.
- 1201 (2002). Age of the Corsica–Sardinia rotation and Liguro–Provençal Basin spreading: new
- 1202 paleomagnetic and Ar/Ar evidence. Tectonophysics 347, 231–251.
- 1203
- 1204 Sullivan, W. A., Boyd, A. S. & Monz, M. E. (2013). Strain localization in homogeneous
- 1205 granite near the brittle–ductile transition: A case study of the Kellyland fault zone, Maine,
- 1206 USA. Journal of Structural Geology 56, 70–88.
- 1207
- 1208 Tribuzio, R. & Giacomini, F. (2002). Blueschist facies metamorphism of peralkaline rhyolites
- 1209 from the Tenda crystalline massif (northern Corsica): evidence for involvement in the Alpine
- 1210 subduction event? Journal of Metamorphic Geology 20, 513–526.

- 1211
- 1212 Vitale Brovarone, A., Beltrando, M., Malavieille, J., Giuntoli, F., Tondella, E., Groppo, C.,
- 1213 Beyssac, O. & Compagnoni, R. (2011a). Inherited Ocean-Continent Transition zones in
- 1214 deeply subducted terranes: Insights from Alpine Corsica. *Lithos* 124, 273–290.
- 1215
- 1216 Vitale Brovarone, A., Beyssac, O., Malavieille, J., Molli, G., Beltrando, M. & Compagnoni,
- 1217 R. (2013). Stacking and metamorphism of continuous segments of subducted lithosphere in a
- 1218 high-pressure wedge: The example of Alpine Corsica (France). *Earth-Science Reviews* 116,
- 1219 35–56.
- 1220
- 1221 Vitale Brovarone, A., Groppo, C., Hetényi, G., Compagnoni, R. & Malavieille, J. (2011b).
- 1222 Coexistence of lawsonite-bearing eclogite and blueschist: phase equilibria modelling of
- 1223 Alpine Corsica metabasalts and petrological evolution of subducting slabs. *Journal of*
- 1224 *Metamorphic Geology* 29, 583–600.
- 1225
- 1226 Vitale Brovarone, A. V. & Herwartz, D. (2013). Timing of HP metamorphism in the Schistes
- 1227 Lustres of Alpine Corsica: New Lu-Hf garnet and lawsonite ages. *Lithos* 172, 175–191.
- 1228
- 1229 Zarki-Jakni, B., van der Beek, P., Poupeau, G., Sosson, M., Labrin, E., Rossi, P. & Ferrandini,
- 1230 J. (2004). Cenozoic denudation of Corsica in response to Ligurian and Tyrrhenian extension:
- 1231 Results from apatite fission track thermochronology. *Tectonics* 23, TC1003.
- 1232
- 1233 Figures Captions
- 1234

Figure 1. Geological map of the studied area. (a) Tectonic map of western Mediterranean region. (b) Geological map of Alpine Corsica associated with a representative cross-section (modified after Vitale Brovarone et al., 2013). Ce: Centuri continental unit; Fa: Farinole continental unit; In: Inzecca unit; Ma: Macinaggio unit; Ne: Nebbio unit; Sa: Sampolo unit; SDP: Serra di Pigno continental unit; SL: Santa Lucia; Pi: Pineto ophiolitic unit; Po: Popolasca unit.

Figure 2. Simplified geological map (modified after Rossi et al., 1994; 2001) with main planar fabrics of the Tenda massif. (a) Foliation map. (b) S_2 trajectories map. Note that S_2 depicts anastomosing structures around grey zones that represent volumes where S_1 or even undeformed rocks are preserved. (c) Statistics of the foliations geometries. Foliation poles are presented with density contours in Schmidt's lower hemisphere equal-area projection and strike preferred orientations are given by rose-diagrams. Geometries of the large-scale were retrieved by the SW-NE elongation of the clouds showing a NW-SE dome. S_1 and S_2 strikes are characterized by different preferred orientations from N170°E to N140°E, respectively. Note that the S_2 dip is relatively flat comparing to S_1 that displays more dispersed dips from flat to vertical.

Figure 3. Characterization of the main planar fabrics and folds at different scales. (a) Subvertical S_1 of the Mt Genova (view from the south). (b) Close-up view of D_2 folds affecting S_1 and associated with the development of a S_2 crenulation cleavage in the center of the Tenda massif (N 42.654838° E 9.190533°). (c) Close-up view of D_2 folds and folded S_1 associated with the development of a S_2 crenulation cleavage in the lowermost structural levels of the ETSZ (N 42.640269° E 9.264771°). Note that S_1 carries blue-amphibole formed during D_1 . (d) Microscopic view from the previous outcrop (insert c). Note that S_1 carries

blue-amphiboles that are retrogressed into chlorite in S_2 cleavage. (e) Equal-area projection plots (Schmidt's lower hemisphere) of structural features characterizing D_2 folds from typical outcrops over the Tenda massif: Envelopes of folded S_1 are steeply dipping while S_2 is sub-horizontal, both striking NNW-SSE, just as the D_2 fold axis trends.

Figure 3 (continued). (f and g) Landscape views of the Saleccia Shear Zone (SSZ) (f: view from the main D81 road) and the ETSZ (g: view from the Punta Mortella) both marked by the development of a penetrative S_2 fabric. S_2 planes from the SSZ and the ETSZ are plotted. (h) Late D_2 folds affecting S_2 in the Punta Mortella (N 42.718401° E 9.250547°). (i) Equal-area projection plots (Schmidt's lower hemisphere) of late D_2 folds axis of the ETSZ from Punta di Cepo to Punta Mortella. Note that axis trends evolve from NNW-SSE away from the ETSZ (Fig. 3e) to ENE-WSW near its most deformed parts.

Figure 4. Stretching lineations of the Tenda massif. (a) Stretching lineation map and associated sense of shear. (b) Statistics of the stretching lineations geometries. Lineation poles are presented with density contours in Schmidt's lower hemisphere equal-area projection and trends preferred orientations are given by rose-diagrams. Note that L_1 and L_2 trends are characterized by slightly different preferred orientations from N063°E to N052°E, respectively.

Figure 5. Kinematics of D_2 deformation in the whole Tenda massif. (a) Representative cross section of the Tenda massif from west to east showing the large-scale structures and the geometries of D_2 top-to-the-NE (green) and D_1 top-to-the-SW shearing (blue). Note that both D_1 and D_2 affect the whole massif; the undeformed Variscan protolith being locally preserved. Also showed are stereographic projections of structures along the cross-section. Letters

highlight the position of pictures presented in the rest of figure 5. (b) Top-to-the-NE mylonites in the Saleccia rhyolite, west of Saleccia. (c) Localized top-to-the-NE narrow shear zone in the west of the Tenda massif (N 42.667220° E 9.148613°). (d) Localized top-to-the-NE steep semi-brittle shear zone affecting an isotropic protolith in the ETSZ (N 42.650459° E 9.263404°).

Figure 5 (continued). (e) Rhyolite intrusion (top of a small-scale dome) asymmetrically boudinaged in a pelitic matrix, west of the Tenda massif (N 42.661448° E 9.130829°). (f) Zoom on the previous outcrop: D₂ folds with folded S₁ associated with S₂ crenulation. Note, from e and f, that S₂ is sigmoidal in between top-to-the-NE shear bands. (g) Top-to-the-NE shearing of a granitic dyke within a micaschists matrix (Punta di Mignola shear zone; contact between the volcano-sedimentary sequence and the granitic complex). (h) Blue-amphibole forming sigma-clasts compatible with top-to-the-NE shearing in the ETSZ (N 42.658489° E 9.272518°). (i) Microscopic view in thin section from the previous outcrop. Top-to-the-NE shearing is associated with the boudinage and truncation of blue-amphibole that is retrogressed into chlorite. Note that crystallization of small blue-amphiboles in tails of sigma-clasts also occurs (see white square). (j) Mylonites in the uppermost levels of the ETSZ. Mineral association is made of quartz, albite, phengite, chlorite and pyrite (N 42.700461° E 9.26382°).

Figure 6. Examples of D₁ top-to-the-SW criteria. (a, b, c) Typical outcrop showing preserved top-to-the-SW criteria in the lowermost structural levels of the ETSZ (N 42.649840° E 9.266655°). (a) In poorly deformed zone, gently dipping top-to-the-SW localized shear bands affect an undeformed granite. Note that S₁ only develops near shear bands. (b) With increasing deformation, the outcrop is characterized by distributed strain with mylonites. Top-

to-the-SW shear bands dip to the west when S_1 appears flat. (c) Top-to-the-SW shearing appears sometimes localized through high strain bands. Note that this band widens from the left to the right of the picture. Note also that shear bands in (a) and the high strain band in (c) have the same orientation. (d) Toward the top of the ETSZ, both S_1 and shear bands dip to the east, compatibly with the general trend of the ETSZ (N 42.654916° E 9.267789°). Top-to-the-SW shearing is clearly recorded in the blue-amphibole stability field. (e) Cross-cutting relationships between top-to-the-NE and top-to-the-SW deformation in the ETSZ (N 42.656507° E 9.272097°).

Figure 7. D_2 strain intensity scale in the granitic protolith. (a) Grade 0s.s: outcrop picture of a typical isotropic granite (N 42.67536° E 9.22939°). The magmatic parageneses is equant and apparently metastable. (b) Grade 0s.s: magmatic hornblendes are incipiently statically replaced by blue-amphiboles involved in a coronitic reaction (N 42.66067° E 9.23987°). (c) Grade 0W: top-to-the-SW shear bands and sigmoidal S_1 in between (N 42.649840° E 9.266655°). (d) Grade 1: D_2 folds with crenulated still steeply dipping S_1 and formation of S_2 crenulation cleavage (N 42.655945° E 9.272113°). Note that deformation is characterized by only a pure shear component. (e) Grade 2: appearance of simple shear components on S_2 cleavage (N 42.700461° E 9.26382°). S_1 in between shows sigmoidal feature indicating top-to-the-NE shearing. Angle between shear bands (S_2) and foliation (S_1) is about 60°. (f) Transition between grades 2 and 3: top-to-the-NE shear bands with sigmoidal S_2 in between (N 42.70249° E 9.25855°). Spacing between shear bands is around 5-10 cm. Angle between shear bands and S_2 is between 60 and 30°. D_2 folds remain visible. (g) Transition between grades 3 and 4: penetrative mylonitic fabric (N 42.700461° E 9.26382°). Spacing between shear bands is close to 3-2 cm, and angle between shear bands and S_2 is about 30°. (h) Grade

4: development of ultramylonite bodies within strongly flattened mylonitic fabric (N 42.71928° E 9.23598°). See text for further explanations.

Figure 8. Synoptic view of all strain grades described in figure 7 along a virtual D_2 strain gradient. This gradient is characterized by an increase of D_2 fabrics resulting in the progressive overprint of D_1 structures. Note that both spacing between shear bands and angle between S_2 and top-to-the-NE shear bands decrease upward, from grade 2 to grade 4. Note that D_1 fabrics have always been observed at the base of the D_2 strain gradient in close association with undeformed rocks.

Figure 9. Examples of structural evidence of the time-sequence relative to the increasing strain intensity. Volumes characterized by grade 2 are asymmetrically boudinaged in volumes characterized by grade 3 at outcrop (a: N 42.659128° E 9.273212°) and landscape-scales (b: N 42.71167° E 9.255127°).

Figure 10. Strain grades map of the Tenda massif within granitic protoliths. One of the five grades (see Figs. 7 and 8) has been attributed for each of the 617 visited outcrops. Interpolation highlights either the overall increase of finite deformation toward the ETSZ and the SSZ or more local strain localization. (b) Cross-sections normal to the regional foliation and parallel to the regional stretching across the main gradients of deformation. SSZ: Saleccia Shear zone; ETSZ: East Tenda Shear Zone.

Figure 11. Available biostratigraphic and radiochronologic age constraints of Alpine events for the Gulf of Lion basin and Corsica-Sardinia continental portions. Data are from (1) Séranne, 1999; Oudet et al., 2010; Jolivet et al., 2015; (2) Faccenna et al., 2002; (3) Speranza

et al., 2002; (4) Bonnal et al., 1973; Bezert and Caby, 1988; Jourdan, 1988; Egal and Caron, 1989; (5) Ferrandini et al., 2003; (6) Cavazza et al., 2007; (7) Brunet et al., 2000; (8) Martin et al., 2011; (9) Vitale Brovarone and Herwartz, 2013; (10) Rossetti et al., 2015; (11) Fellin et al., 2006; (12) Jakni et al., 2000; Zarkni-Jakni et al., 2004; (13) Lahondère and Guerrot, 1997; (14) Maggi et al., 2012; (15) Di Vincenzo et al., 2016.

Figure 12. Tectonic model of Alpine Corsica orogenic system from Eocene to Miocene (inspired from Ravna et al., 2010; Molli and Malavieille, 2011; Vitale Brovarone et al., 2013 and references on chronological constraints presented in Fig. 11) showing the evolution of deformation in the Tenda massif from burial to exhumation. During Late Eocene, the Tenda massif was buried underneath the Ligurian accretionary complex until its juxtaposition below the Schistes Lustrés nappe, and affected by top-to-the-SW deformation (D_1) firstly localized then pervasive at maximal depth. There, the two units are tectonically coupled (see also Molli et al., 2006). From Middle Oligocene to Miocene, syn-to post-orogenic exhumation was marked by a progressive localization of strain from distributed (pervasive D_2 folding of previous structures) to localized notably near the ETSZ. All sketches are oriented SW-NE.

Figure 1

Figure 3-1

Figure 3-2

Figure 4

Figure 5-1

Figure 5-2

Figure 6

Figure 7

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1415 Figure 8

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Figure 9

Figure 10

Figure 11

Figure 12

1433 Highlights

1434 The Tenda massif was not rigid, as deformation was pervasive when exhumation started.

1435 Strain mapping shows a strain localization toward the roof and internal shear zones.

1436 The large-scale finite architecture is a dome of extensional-related structures.

1437 Lithological contacts drive strain localization.

1438 Style of deformation was controlled by protolith nature and P-T-fluids conditions.