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Late Palaeozoic tectonics in Central Mediterranean: a reappraisal

Giancarlo Molli[1*], Andrea Brogi[2,3], Alfredo Caggianelli[2], Enrico Capezzuoli[4], Domenico Liotta[2,3], Amalia Spina[5] and Ivan Zibra[6]

Abstract

A revision of late Palaeozoic tectonics recorded in Tuscany, Calabria and Corsica is here presented. We propose that, in Tuscany, upper Carboniferous-Permian shallow-marine to continental sedimentary basins, characterized by unconformities and abrupt changes in sedimentary facies, coal-measures, red fanglomerate deposits and felsic magmatism, may be related with a transtensional setting where upper-crustal splay faults are linked with a mid-crustal shear zone. The remnants of the latter can be found in the deep-well logs of Pontremoli and Larderello-Travale in northern and southern Tuscany respectively. In Calabria (Sila, Serre and Aspromonte), a continuous pre-Mesozoic crustal section is exposed, where the lower-crustal portion mainly includes granulites and migmatitic paragneisses, together with subordinate marbles and metabasites. The mid-crustal section, up to 13 km-thick, includes granitoids, tonalitic to granitic in composition, emplaced between 306 and 295 Ma. They were progressively deformed during retrograde extensional shearing, with a final magmatic activity, between 295 ± 1 and 277 ± 1 Ma, when shallower dykes were emplaced in a transtensional regime. The section is completed by an upper crustal portion, mainly formed by a Palaeozoic sedimentary succession deformed as a low-grade fold and thrust belt, and locally overlaying medium-grade paragneiss units. As a whole, these features are reminiscent of the nappe zone domains of the Sardinia Variscan Orogen. In Corsica, besides the well-known effusive and intrusive Permian magmatism of the "Autochthonous" domain, the Alpine Santa Lucia Nappe exposes a kilometer-scale portion of the Permian lower to mid-crust, exhibiting many similarities to the Ivrea Zone. The distinct Mafic and Granitic complexes characterizing this crustal domain are juxtaposed through an oblique-slip shear zone named Santa Lucia Shear Zone. Structural and petrological data witness the interaction between magmatism, metamorphism and retrograde shearing during Permian, in a temperature range of c. 800–400 °C. We frame the outlined paleotectonic domains within a regional-scale, strain–partitioned, tectonic setting controlled by a first-order transcurrent/transtensional fault network that includes a westernmost fault (Santa Lucia Fault) and an easternmost one (East Tuscan Fault), with intervening crustal domains affected by extensional to transtensional deformation. As a whole, our revision allows new suggestions for a better understanding of the tectonic framework and evolution of the Central Mediterranean during the late Palaeozoic.

Keywords: Post variscan tectonics, Central mediterranean, Permian sedimentary record, Permian HT metamorphism and magmatism, Regional fault system

1 Introduction

The distribution of pre-Mesozoic rocks in the Central Mediterranean is uneven (Fig. 1), with the most relevant exposures in the Alps, and only scattered outcrops, or shallow-crustal subsurface occurences known so far, in and around the Italian peninsula (Rau and Tongiorgi 1981; Cassinis et al. 2000 2012; Vai 2001; Scisciani and Esestime 2017).

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*Correspondence: giancarlo.molli@unipi.it
[1] Università di Pisa, Via S.Maria 53, 56126 Pisa, Italy
Full list of author information is available at the end of the article

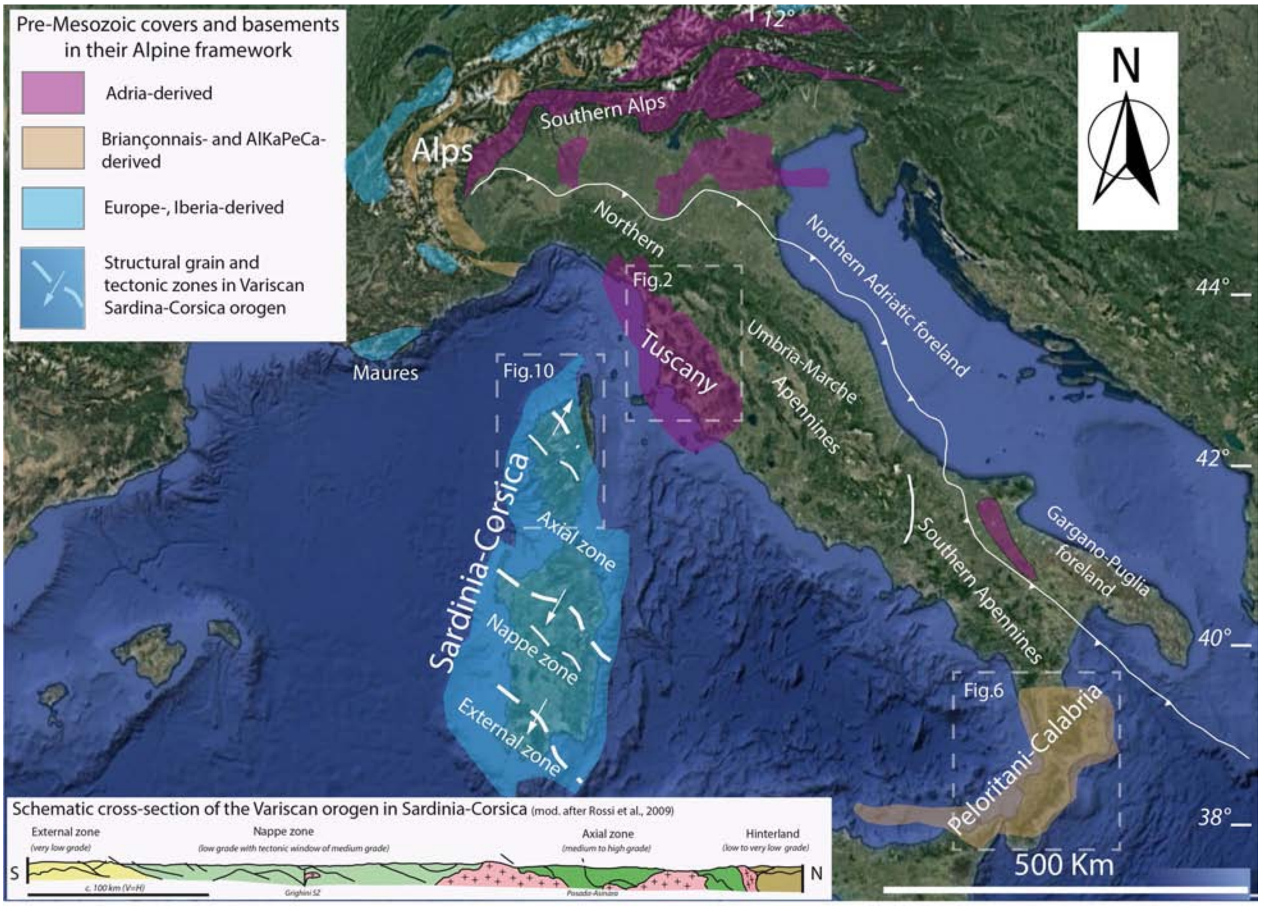


Fig. 1 Exposure or shallow subsurface (minor than 6 km) occurences of pre-Mesozoic covers and basements in Central Mediterranean. The colors in the map are related to their Alpine framework and respectively referred to the Adria, Briançonnais /AlKaPeCa and European/Iberian domains (Schmid et al. 2004; Handy et al. 2010). The main tectonic zones (with arrows indicating the tectonic transport directions) are reported for the Sardinia-Corsica Variscan orogen as also figured in the general and schematic cross-section after Rossi et al. (2009). Google image

The Variscan basement and upper Carboniferous-Permian successions are exposed in the Alpine chain (for an overall view see: Dal Piaz 1993; von Raumer and Neubauer 1993, Vai 2001; Guillot et al. 2009; Spies et al. 2010; von Raumer et al. 2013; Cassinis et al. 2018; Ballevre et al. 2018), witnessing continental crustal segments of the different Variscan paleotectonic domains, later involved and deformed, at various crustal levels, during the Alpine evolution. Considering these, recent contributions have focused on the role of upper Carboniferous-Permian tectonics, its relationships with the Variscan mountain building and collapse, and with the proto-Alpine Tethyan rifting (e.g. Schaltegger and Brack 2007; Schuster and Stüwe 2008; Froitzheim et al 2008; Cassinis et al. 2012; Kunz et al. 2017; Bergomi et al. 2017; Festa et al. 2018; Ballevre et al. 2018; Pohl et al 2018; Roda et al. 2018). Despite noteworthy exceptions (Padovano et al. 2012; Cassinis et al. 2018) the conclusions of these works overlooked the large region corresponding to the Italian peninsula and surroundings, because of the scattered occurrences of pre-Mesozoic rocks, as well as of their poor exposure and strong involvement in Central Mediterranean Tertiary tectonics.

In this paper, we revise the pre-Alpine tectonic evolution of some pre-Mesozoic crustal fragments exposed in Tuscany, Calabria and Corsica. By combining data collected by our group in the last 15 years with a reappraisal of recently published literature, we intend to frame the role and importance of upper Carboniferous-Permian tectonics and its relationship, if any, with the history of the Variscan orogen and Alpine Tethyan rifting. Differently from other regional reviews of the southern Europe-Mediterranean area, which have mainly focused on the upper crustal records (e.g. volcanics or sedimentary history of basins, Cassinis et al. 2018), we include here data from the deeper crustal levels, giving a more complete overview of the processes affecting the investigated regions during the late Palaeozoic.

Following former suggestions (e.g. Bard 1997; Vai 2001; Alvarez and Shimabukuro 2009; Patacca and Scandone 2011), we will use the Sardinia-Corsica Variscan framework to examine and compare the basement units and Permo-Carboniferous successions of Tuscany, Calabria and Corsica. This will allow us to constrain their relative paleotectonic positions and relationships, as well as the role of some late Palaeozoic structures. Moreover, our study areas include three crustal domains (Santa Lucia in Corsica, the Calabria Crystalline and the lost External Ligurian of the Northern Apennines), which share with the Ivrea Zone of the Southern Alps many first order features, including amphibolite to granulite-facies formations (kinzigites), as well as mafic underplated units (mafic complexes), therefore providing a direct record of the late Palaeozoic evolution of the lower to middle continental crust. Finally, considering that in terms of the Alpine tectonic framework our study areas may be referred to the former Europe, AlKaPeCa (Alboran-Kabilie-Peloritani-Calabria) and Adria domains respectively (Handy et al. 2010), a better knowledge of their original relationships may add further constraints to the Mesozoic and Tertiary tectonic history of the Central Mediterranean region.

2 The Variscan orogenic template and its remnants in Sardinia and Corsica

In central Mediterranean, pre-Mesozoic continental units are well known and exposed in Sardinia and Corsica. These units were, before the Tertiary history of the region i.e. in the Alpine framework, part of the Iberian-European domain (Dewey et al. 2009; Handy et al. 2010; van Hinsbergen et al. 2020). Sardinia and Corsica (with the exception of the its easternmost part, the so-called Alpine Corsica) were only slightly involved in the Alpine orogenic evolution, thus preserving the overall tectonic architecture of the southernmost segment of the Variscan orogen (Carmignani et al. 1979, 1994; Rossi et al. 2009).

2.1 Zonation of the Variscan belt in Sardinia-Corsica

Four different structural zones may be distinguished by combining data from Sardinia and Corsica (Carmignani et al. 1979, 1994; Lardeaux et al. 1994; Elter and Pandeli 2005; Helbing et al. 2006; Rossi et al. 2009; Edel et al. 2014):

1. *An external southern zone* covering the southwestern part of Sardinia and consisting of a subgreenschist facies fold-and-thrust belt, affecting a sedimentary succession ranging in age from upper Vendian to lower Carboniferous (Funedda 2009);
2. *A nappe zone* affected by greenschist facies metamorphism (Conti et al. 2001) and consisting of a continental arc-related volcanic suite of Middle Ordovician age (Oggiano et al. 2013), embedded within a thick Palaeozoic metasedimentary succession;
3. *An inner or axial zone* characterized by medium-to high-grade metamorphic rocks locally including metabasites with a MORB geochemical signature and relicts of eclogite facies metamorphism (Cappelli et al. 1992; Palmeri et al. 2004; Giacomini et al. 2008; Cruciani et al. 2015) intruded by late Variscan granitoids. Similarly to NE Sardinia, most of Corsica shows intrusive late Variscan and post Variscan plutons and volcano/sedimentary successions (see below) with relict septa of the host rocks. These rocks show similarity to those of the axial zone of Sardinia (Rossi et al. 2009; Casini et al. 2015), and are bounded, in NE Corsica, by;
4. *Hinterland* Armorica (?)-like block and northeast vergent retrowedge stack in which Panafrican micaschists and a very low-grade Palaeozoic metasedimentary cover may be recognized (Rossi et al. 1995; 2009; Faure et al. 2014).

2.2 Tectonic history of the Variscan belt in Sardina-Corsica

For the Variscan belt in Sardinia-Corsica some major tectonic issues represent still hotly debated topics, for instance the occurrence, location, and vergence of the suture zone tracking the early stages of the oceanic subduction and accretion, together with the type of the continent (Armorica microplate, Hun terrane or Brunia) involved in the collision with Gondwana (e.g. Cappelli et al. 1992; Carmignani et al. 1994; Matte 2001; Stampfli et al. 2002; Franceschelli et al. 2004; Helbing et al. 2006; Giacomini et al. 2007; Rossi et al. 2009; Guillot et al. 2009; Corsini and Rolland 2009; Elter and Padovano 2010; Oggiano et al. 2013; von Raumer et al. 2013; Faure et al. 2014; Li et al. 2014) or the connections between the inner zones and their structures and the regional scale tectonic framework, e.g. the East Variscan Shear Zone (Corsini et al. 2009; Elter and Padovano 2010; Carosi et al. 2010). Most authors, however, agree that the first-order shaping of the South Variscan belt in Sardinia and Corsica was related to the frontal to oblique continent–continent collisional processes. The continental collision produced, in Sardinia (Fig. 1): (i) SW-facing folds, (ii) top to the S/SW low angle thrusting of the high grade units of the axial zone on top of medium grade units of the northern nappes and (iii) the main fabric and structures in the low-grade nappe and external domains of the central and southern Sardinia (Arthaud 1970; Carmignani et al. 1979, 1994; Conti et al. 2001; Carosi and Oggiano

2002; Franceschelli et al. 2004; Carosi et al. 2005; Cruciani et al. 2015). The collisional structures, referred to 360–320 Ma (Di Vincenzo et al. 2004; Li et al. 2014), were reworked within regional-scale shear zones (e.g. Posada-Asinara and the Grighini Shear zones in Sardinia, Zicavo and La Vaccia in NE Corsica) interpreted as transpressional structures with dextral kinematics (Elter et al. 1990; Thevoux-Chabuel et al. 1995; Carosi and Palmeri 2002; Oggiano and Rossi 2004; Helbing et al. 2006; Iacopini et al. 2008; Frassi et al. 2009). These structures were developed during a phase of regional-scale change in transport direction (from S/SW to W/NW) in the nappe- and foreland-zones of Sardinia (Conti et al. 2001). Nevertheless, these structures and the related change in kinematics were differently interpreted by other authors (Musumeci 1992; Oggiano and Rossi 2004; Helbing et al. 2006; Casini and Oggiano 2008) as connected with a regional-scale switch from collisional contraction to post-collisional extension/transtension. This younger setting may be constrained in age between 320 and 305 Ma along the Posada-Asinara Shear zone (Di Vincenzo et al. 2004; Giacomini et al. 2008) and between 305 and 295 Ma in the southernmost Grighini Shear zone, as well as in the Zicavo and La Vaccia in NE Corsica (Musumeci 1992; Thevoux-Chabuel et al. 1995; Oggiano and Rossi 2004; Cruciani et al. 2015; Cruciani et al. 2017).

2.3 The syn- to post-orogenic magmatism: the Corsica-Sardinia Batholith

Early- and late-collisional deformation structures were associated and followed by widespread magmatism, forming the so-called Corsica-Sardinia Batholith, classically subdivided into three main magmatic suites (Orsini 1976, 1979; Ghezzo and Orsini 1982; Rossi and Cocherie 1991; Ferré and Leake 2001; Edel et al. 2014; Renna et al. 2006; Casini et al. 2015), from a petrological-geochemical point of view. The early magmatic sequence (U1), forming a small fraction of the Corsica-Sardinia Batholith, is mainly documented in northwestern Corsica and subordinately in NE Sardinia, where it is associated to anatexites and layered migmatites (Rossi et al. 2015; Casini et al. 2015). The U1 suite mainly includes high-Mg–K calc-alkaline plutons (mainly monzonite to granite-adamellites) emplaced over a short time span at c. 345–330 Ma (Paquette et al. 2003; Li et al. 2014; Rossi et al. 2015), as a result of the mixing between mantle-derived and lower-crustal melts, and interpreted as sealing the original suture zone (Rossi et al. 2009).

The late- to post-Variscan magmatism is represented by the U2 and U3 magmatic suites (Figs. 1, 10a) which form the largest part of the Corsica-Sardinia Batholith (del Moro et al. 1975; Paquette et al. 2003; Oggiano et al. 2004; Casini et al. 2008, 2015). The U2 magmatic suite is mainly characterized by granitoids with MgO content lower than the U1 suite, emplaced between 310 and 280 Ma with a climax around c.305 Ma during which calc-alkaline granitoids were emplaced (Paquette et al. 2003). The U3 magmatic suite, which was partly coeval with the latest U2 leucomonzogranites, consists of tholeiitic complexes emplaced from c. 304 Ma to c. 280 Ma (Paquette et al. 2003; Cocherie et al. 2005) and subordinate alkaline granitoids formed at c. 290 Ma to c. 280 Ma (Cocherie et al. 2005; Renna et al. 2006; Rossi et al. 2015). The early U2 plutons were emplaced at shallow depths $P \leq 0.4$ GPa forming elliptical bodies NW–SE trending characterized by a sub-horizontal foliation, and magmatic lineation (Casini et al. 2015). Their close spatial relationship with the late Carboniferous shear zones, dated at about 320–305 Ma, in northern Sardinia (Di Vincenzo et al. 2004; Carosi et al. 2012) supports a rapid, probably episodic melt migration localized along the ductile shear zones rooted in the lower crust and emplaced in a stretched upper crust during orogen-parallel extension (Casini et al. 2015). Conversely, most of the lower Permian U2 and U3 plutons show a completely different architecture, with a NE/SW trend, sharp contacts with either the metamorphic basement or older granites, weak development of an internal fabrics, and presence of stoped blocks in the pluton roof zones (Gattacceca et al. 2004; Cocherie et al. 2005).

3 The Palaeozoic rocks in the Northern Apennines

The Northern Apennines are characterized (Fig. 2) by stacked units belonging to a former accretionary wedge (Ligurian and sub-Ligurian Units) formed during the Tertiary closure stages of the Ligurian Tethys ocean, overlying the continental derived thrust-sheets and cover nappes of the Adria continental margin of Tuscan and Umbria-Marche Domains (e.g. Elter 1975; Bernoulli 2001; Butler et al. 2006; Molli 2008; Malavieille et al. 2016; Schmid et al. 2017).

3.1 Palaeozoic rocks in the Tuscan units

In the Northern Apennines, the Palaeozoic basement and covers belong to the continental units derived from the Tuscan and Umbria-Marche paleodomains, classically considered as part of the westernmost Adria-Africa continent (Fig. 2). The pre-Mesozoic basement is discontinuously exposed (e.g. Lazzarotto et al. 2003), and it has been sampled by deep boreholes (Gianelli et al. 1978; Anelli et al. 1994; Pandeli et al. 1994; Batini et al. 2003) mainly in the inner Tuscan sectors of the chain (Figs. 1, 2) and also in the external foreland domain (Vai 2001; Scisciani and Esestime 2017). For a complete bibliography on the pre-Mesozoic rocks of the Northern Apennines, the

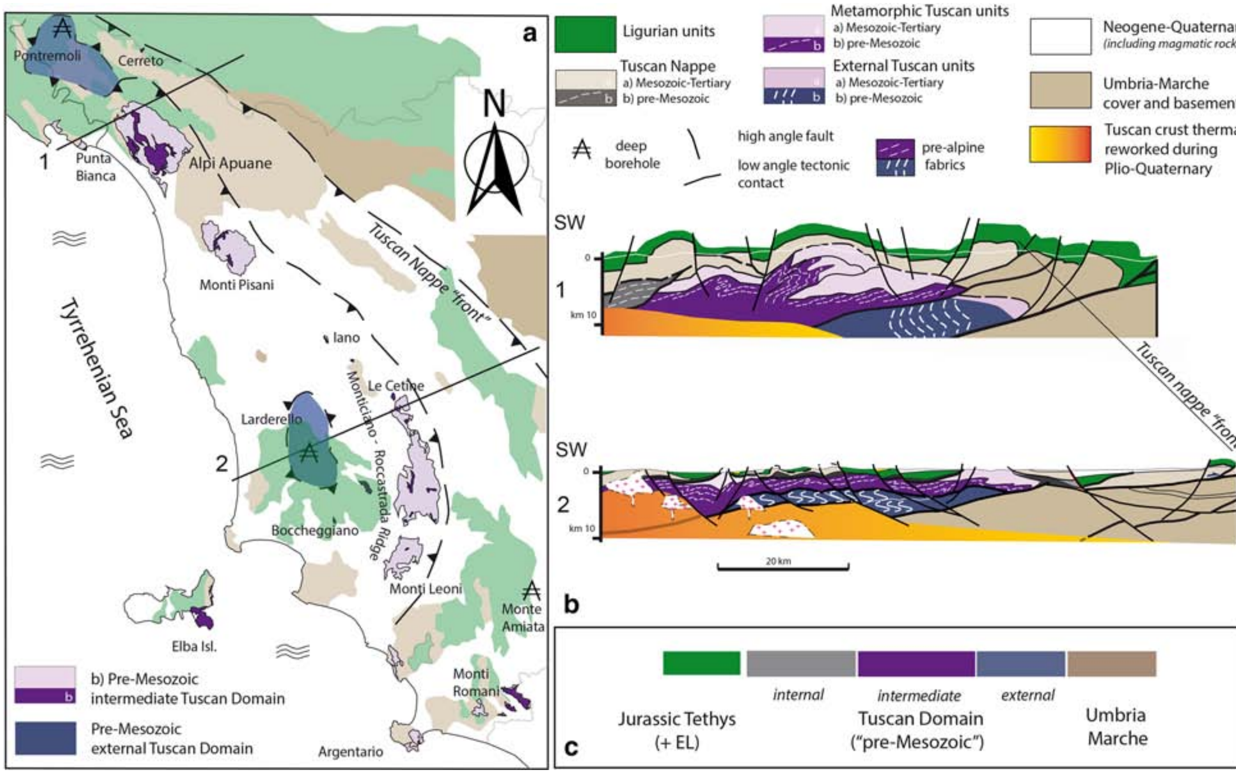


Fig. 2 **a** Tectonic map of Northern Apennines with the main pre-Mesozoic exposures of the Tuscan metamorphic units and of the external Tuscan domain in the subsurface (Pontremoli and Larderello-Travale deep wells); **b** geological cross sections of the Northern Apennines north and south of Arno, with indication of the pre-Mesozoic occurences; **c** simplified Mesozoic paleotectonic frame showing the relative pre-Mesozoic position of the Tuscan domains within Adria

reader is addressed to Rau and Tongiorgi (1974, 1981), Bagnoli et al. (1979), Tongiorgi and Bagnoli (1981), Vai and Cocozza (1986), Conti et al. (1991), Pandeli et al. (1994), Vai (2001), Pandeli et al. (2005), Aldinucci et al. (2008a, b), Cassinis et al. (2018) and references therein.

According to our recent research in northern Tuscany (Molli et al. 2018), the pre-Mesozoic rocks of the Northern Apennines may be referred to three different tectonic zones (Fig. 2), from west to east:

1. *Internal zone* (former basement of the Tuscan Nappe). This internal zone is characterized by medium grade units mainly made up of micaschists with minor bodies of amphibolites (Ricci 1968; Di Sabatino et al. 1979). Recent petrological studies (Molli et al. 2002; Franceschelli et al. 2004; Pandeli et al. 2005; Elter and Pandeli 2005; Lo Pò et al. 2017) in the Cerreto area, north of the Alpi Apuane area (Figs. 2, 3 and 4), defined a peak pressure exceeding 1.1 GPa, followed by a peak temperature of 550–590 °C at 0.9–1 GPa. The post-peak evolution occurred at 328–312 Ma (Molli et al. 2002) in micaschists and embedded amphibolites, both evolving in similar P–T conditions and final retrogression stage, being constrained at T < 475 °C, and P < 0.7 GPa, associated with well developed mylonitic fabrics (Figs. 3, 4);

2. *Intermediate zone* to which most of the pre-Mesozoic units may be referred (Figs. 2, and 5). The latter belong to the Tuscan metamorphic units and are exposed in the so-called Mid-Tuscan Ridge (Alpi Apuane, Monti Pisani, Iano, Montagnola Senese-Monti Leoni, Monti Romani) and in the Tuscan Archipelago (Elba island). These units were affected by a regional low-grade (or intermediate to high pressure) and a locally high grade metamorphism during the Apennine orogenesis and late magmatism; different *P*–*T* values have been reconstructed for the different units (Duranti et al. 1992; Theye et al. 1997; Giorgetti et al. 1998; Jolivet et al. 1998; Brunet et al. 2000; Brogi and Giorgetti 2012; Bianco et al. 2015, 2019; Caggianelli et al. 2018). A synthesis of the whole data in the framework of the Northern Apennines evolution is given by Jolivet et al. (1998), Brunet et al. (2000), Rossetti et al. (2002), Franceschelli et al. (2004); Molli (2008), Rossetti

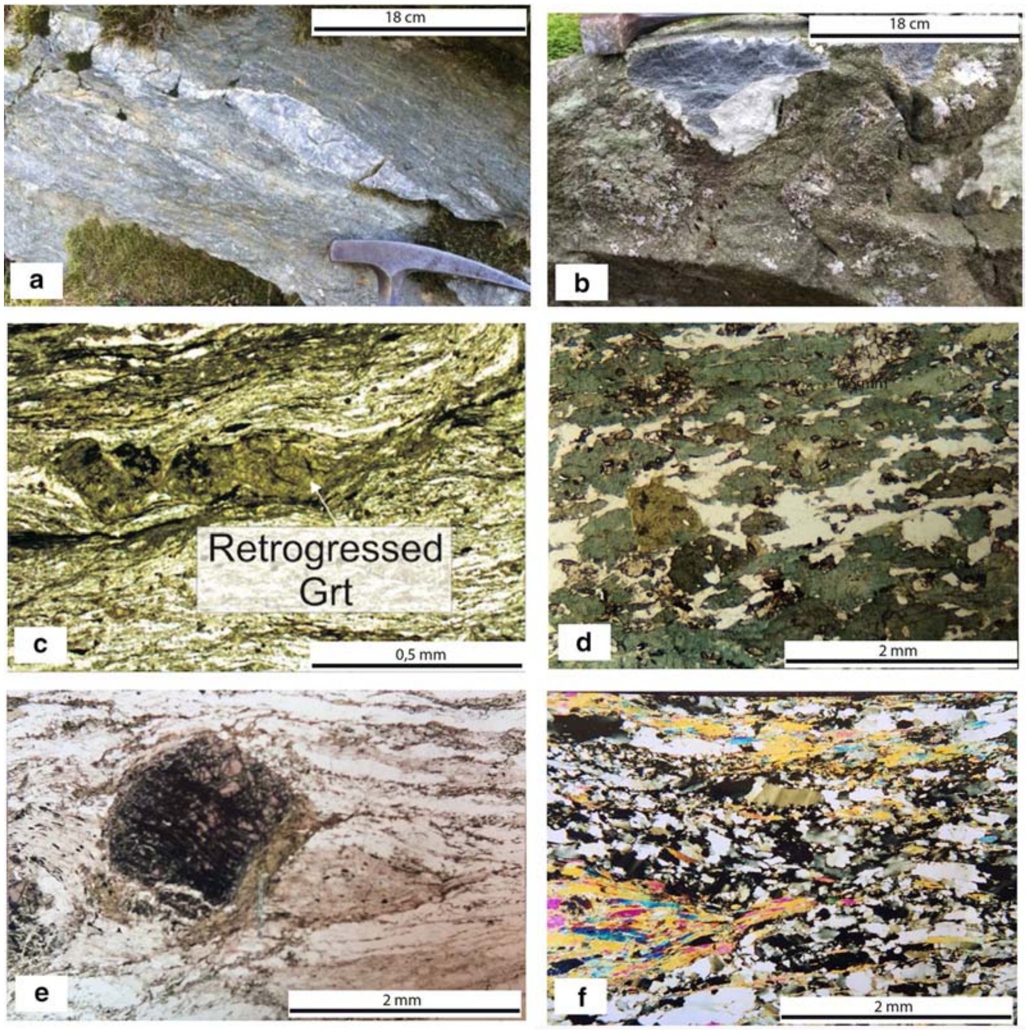


Fig. 3 Mesoscopic appearence of the Cerreto medium grade rocks belonging to the former inner pre-Mesozoic "Tuscan" domain **a** mylonitic micaschist; **b** Morb-derived amphibolite. Plane-polarized light microphotographs of **c** micaschist with fractured and retrogressed garnet porphyroclast (from Lo Po' et al. 2017) and **d** amphibolites (details in Molli et al. 2002; Lo Po' et al. 2017); **e** mylonitic micaschist and **f** quartz-feldspatic mylonites from Pontremoli well (former pre-Mesozoic external "Tuscan" domain)

et al. (2008). The intermediate domain is well documented in the tectonic windows (Punta Bianca, Alpi Apuane and Monti Pisani) to the north of the Arno River (Fig. 2), where it includes low-grade Variscan units (Conti et al. 1993). These mainly contain albite-bearing chloritic phyllites and quarzites including lenses of mafic metavolcanics and calcareous schists (Lower Phyllites), felsic metavolcanics and metavolcanoclastic rocks (Porphyroids and porphyritic schists), metasandstones, quarzites and phyllites (Upper Phyllites), graphitic schists and Orthoceras-bearing dolostone. As a whole, these units are considered as part of a Cambrian? to Silurian succession (Vai 1972; Bagnoli et al. 1979), classically correlated with the Palaeozoic units of central Sardinia (Carmignani et al. 1979, 1994; Conti et al. 1993; Pandeli et al. 1994). This low-grade basement is locally covered by Carboniferous-Permian deposits (Rau and Tongiorgi 1972; Bagnoli et al. 1979; Pandeli et al. 1994; Spina et al. 2019; Figs. 2, 5). In southern Tuscany, the oldest Carboniferous-Permian succession is discontinuously exposed along the Monticiano-Roccastrada Ridge (Figs. 2 and 5). This succession consists of Moscovian bioclastic limestone (Calcare di Sant'Antonio Formation—early Pennsylvanian; Pasini 1991; Lazzarotto et al. 2003; Engelbrecht 2008), related to a carbonatic platform (Cocozza et al. 1987; Lazzarotto et al. 2003), unconformably overlain by

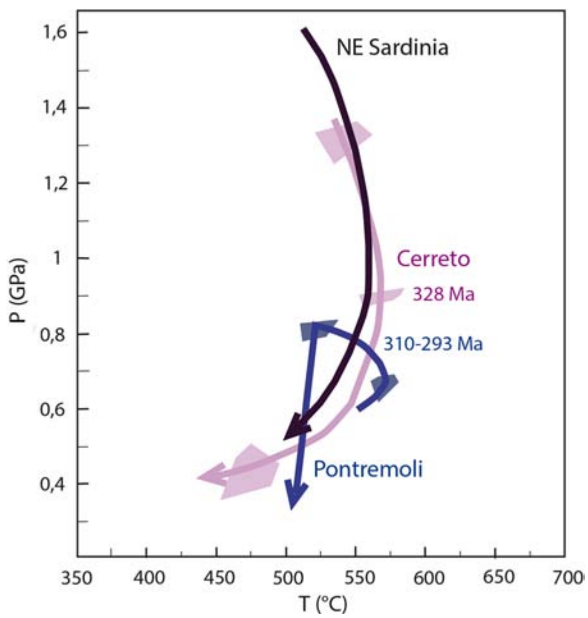


Fig. 4 P–T–t path for inner (pink line) and external (blue line) Tuscan basement compared with that proposed for the axial zone of the Sardinian Variscan belt (Cruciani et al. 2015) in black continuous line (after Lo Po' et al. 2017)

dark phyllites and metasandstones (Scisti a *Spirifer* Fm). The latter were attributed to late Carboniferous-early Permian on the basis of brachiopods (Cocozza 1965), and to late Moscovian-Kasimovian by fusulinid assemblages (Pasini 1991). Other exposures of the upper Carboniferous?-lower Permian succession (i.e. upper Pennsylvanian-Cisuralian *p.p.)* are in the Monti Pisani area (Scisti di San Lorenzo Fm, Rau and Tongiorgi 1974; Bagnoli et al. 1979; Pandeli et al. 2008; Landi degl'Innocenti et al. 2008). Here, they consist of a low-grade metamorphic succession formed by black silty-phyllites and metasandstones. Differently, in the Iano area (Figs. 2 and 5) the upper Carboniferous?-lower Permian succession consists of metaconglomerate with lenses of metasandstone, passing upwards to metasandstone and organic-rich metasiltstones and phyllites (Scisti di Iano Fm—Vai and Francavilla 1974; Costantini et al. 1998; Lazzarotto et al. 2003). These organic-rich siliciclastic sediments were attributed to different palaeoenvironments: from continental (Scisti di San Lorenzo Fm) to coastal-neritic (Scisti di Iano and Scisti a *Spirifer* formations), deposited in an equatorial climate (Rau and Tongiorgi 1974; Costantini et al. 1998; Lazzarotto et al. 2003). A second sedimentary cycle, referred to the middle-late Permian (i.e. Guadalupian–Lopingian), has been constrained by using data from different localities of Tuscany. This cycle includes: (i) middle?-Permian coarse grained metaconglomerate (Breccia di Asciano Fm, in Rau and Tongiorgi 1974; Bagnoli et al. 1979; Pandeli et al. 2008; Landi degl'Innocenti et al. 2008), exposed in the Mt. Pisani area; (ii) middle Permian phyllitic quartzites enriched in volcanic felsic components (Scisti Porfirici di Iano Fm); (iii) middle Permian metarudites, metasandstones and phyllites (Breccia e Conglomerati di Torri Fm), deposited in alluvial fans, covered by siltstones and phyllites with volcanic-rich quartzitic sandstones and conglomerates (Siltiti del Fregione Fm, Costantini et al. 1998; Pandeli 1998). On the other hand, the deposits of Guadalupian–Lopingian age are characterized by continental to marine, locally organic-rich successions and represented by metasandstones, metaconglomerates and metasiltstones that are known with different names in different areas: (i) Montignoso Formation in the Alpi Apuane (Massa Unit) area; (ii) Arenarie di Poggio al Carpino, Le Cetine, Farma, Carpineta, Falsacqua, Quarziti di Poggio alle Pigne and Conglomerato di Fosso Pianaccia formations in the mid-Tuscan range of southern Tuscany (see Fig. 5); (iii) Arenarie rosse di Castelnuovo Fm, as recognized in the subsurface of the Larderello geothermal area; (iv) Mt Calamita and Rio Marina formations in the Elba Island; (v) Arenarie del Monte Argentario Fm, in the Argentario Promontory; (vi) "C" Fm, in subsurface of the the Monte Amiata volcano-geothermal area; (vii) Arenarie di Ponte San Pietro, Quarzite e Fillade di Roccaccia di Montauto, Metarenarie di Botro del Lecceto, Calcescisti di Valle Tegolaia Fm, in the Monti Romani area (Gianelli et al. 1978; Pandeli et al. 1988; Pandeli and Pasini 1990; Moretti et al. 1990; Elter and Pandeli 1991; Cirilli et al. 2002, 2004; Lazzarotto et al. 2003; Aldinucci et al. 2005; 2008a, b; Brogi 2008; Patacca et al. 2011; Spina et al. 2019).

Permian magmatism has been recognized, since the late sixties (Barberi 1966; Rau and Tongiorgi 1974; Bagnoli et al. 1979; Costantini et al. 1998) within volcanosedimentary layers in the Iano exposures, as well as in clasts within Mid-Triassic Verrucano deposits (Rau and Tongiorgi 1974; Franceschelli et al. 2004). More recently, sub-intrusive bodies cross-cutting the Variscan foliation and named Metarhyolite di Fornovolasco Fm. (Vezzoni et al. 2018), have been documented within the Palaeozoic basement of the Alpi Apuane (Pieruccioni etal. 2018). U–Pb zircon dating suggests a 292–271 Ma crystallization age (Fig. 4b, Vezzoni et al. 2018) in the range of Permian magmatism of the Central Mediterranean area (Buzzi and Gaggero 2008; Buzzi et al. 2008; Rossi et al. 2009);

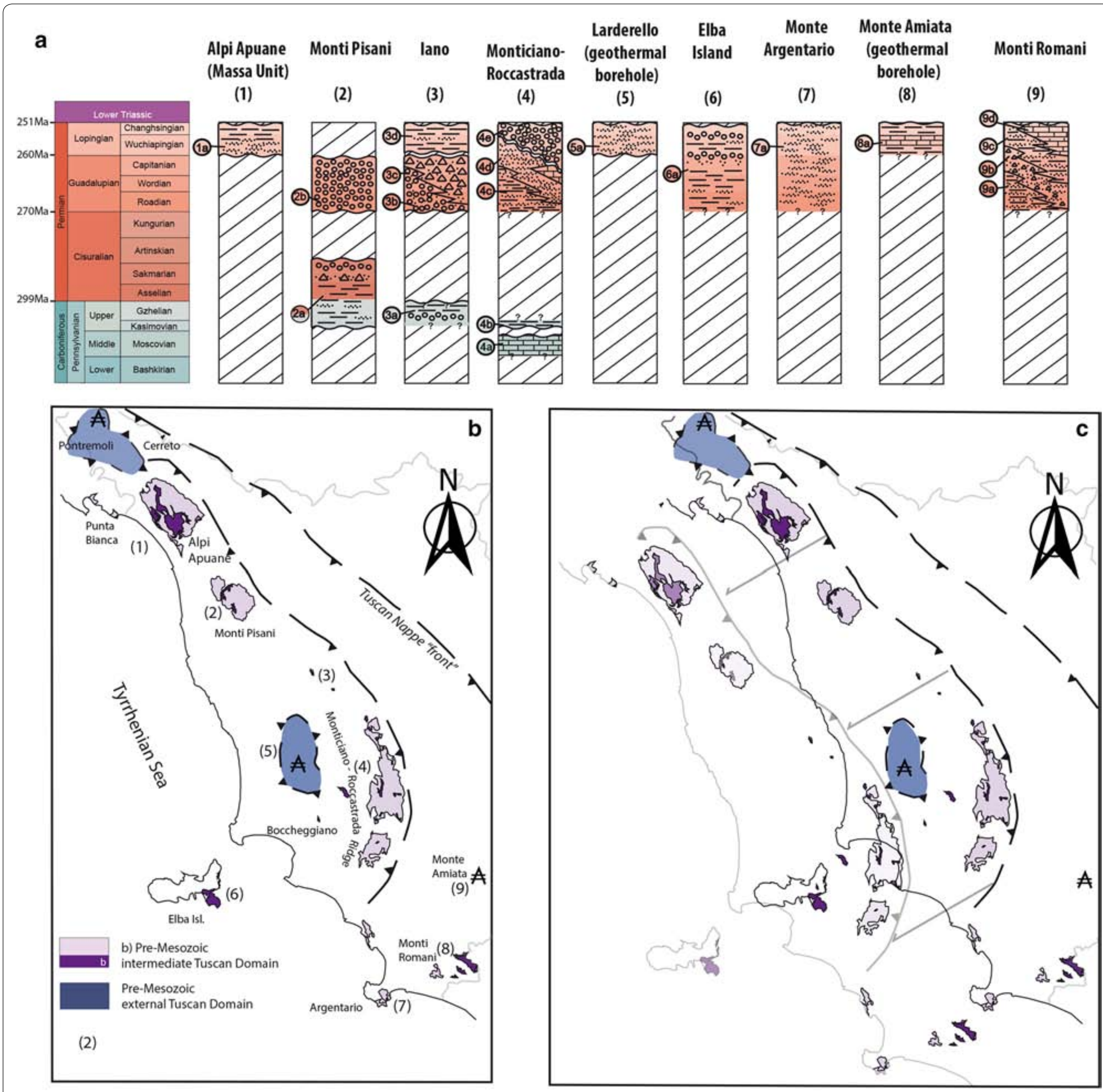


Fig. 5 **a** Upper Carboniferous-Permian successions of the intermediate pre-Mesozoic "Tuscan" domain. Alpi Apuane: 1a—Montignoso Fm; Monti Pisani: 2a – Scisti di San Lorenzo Fm; 2b—Breccia di Asciano Fm; Middle Tuscan Ridge: 3a – Scisti di Iano Fm; 3b—Breccia e Conglomerati di Torri Fm; 3c – Scisti Porfirici Fm; 3d—Fosso del Fregione Fm; 4a—Calcare di Sant'Antonio Fm; 4b—Scisti a Spirifer Fm; 4c—Farma Fm—Falsacqua Fm; 4d—Carpineta Fm—Quarziti di Poggio alle Pigne Fm; 4e—Arenarie di Poggio al Carpino Fm—Le Cetine Fm.- Conglomerato di Fosso Pianacce Fm; Larderello (geothermal borehole): 5a—Arenarie Rosse di Castelnuovo Fm; Elba Island: 6a—Mt Calamita Fm—Rio Marina Fm.; Monte Argentario: 7a—Arenarie del Mt. Argentario Fm; Monte Amiata (geothermal borehole): 8a—C Fm; Monti Romani: 9a—Arenarie di Ponte San Pietro Fm., 9b—Quarzite e Fillade di Roccaccia di Montauto Fm; 9c—Metarenarie di Botro del Lecceto; 9d—Calcescisti di Valle Tegolaia Fm. See the text for related bibliography; **b** location of the different Palaeozoic sequences of the intermediate pre-Mesozoic "Tuscan" domain; **c** retrodeformation of a major Apenninic thrusting of the "Tuscan Metamorphic units" on top of the external Tuscan units (transport direction and kinematic tectonic frame in Molli 2008; Le Breton et al. 2017) useful to infer the relative positions of pre-Mesozoic "intermediate" and "external" domains

3. *External or easternmost zone,* only reached at depth by the Pontremoli well (Eni) and by the boreholes in the Larderello-Travale geothermal areas (Enel), where garnet micaschists and gneisses were encountered (Batini et al. 1983; Elter and Pandeli 1990; Anelli et al. 1994; Pandeli et al. 1994, 2005). In the

Pontremoli borehole (Lo Po' et al. 2016; Molli et al. 2018) medium-grade garnet micaschists associated with quartz-feldspatic mylonites (Fig. 3e, f) show (Fig. 4) thermal metamorphic peak at 575 °C and 0.7 GPa, followed by the peak pressure stage occurring at 520 °C and 0.8 GPa, and then by a nearly isothermal decompression at 475–520 °C characterizing the late retrograde stage (Lo Pò et al. 2016). The peak conditions were dated by monazite geochronology at 310–293 Ma (Lo Pò et al. 2016). In the Larderello-Travale geothermal area, in the same structural position of the Pontremoli well, i.e. below a stack of Tuscan units including the metamorphic Monticiano-Roccastrada unit, micaschists (e.g. Franceschini 1994; Musumeci et al. 2002 for a different interpretation) and gneisses were recognized (Gianelli et al. 1978; Elter and Pandeli 1990; Pandeli et al. 1994; Franceschelli et al. 2004; Pandeli et al. 2005). A Rb/Sr radiometric age of 285 ± 11 Ma was obtained for a micaschist sampled at Larderello (reported in Del Moro et al. 1982). Moreover, through classical geothermobarometry (garnet-biotite and plagioclase-garnet-muscovite-biotite-quartz) and the stability fields of staurolite, muscovite, andalusite and cordierite assemblages, Bertini et al. (1994) estimated peak conditions of 500–600 °C and 0.7 GPa, in the gneiss complex. The peak was followed by an isothermal decompression to 0.2–0.35 GPa at 500–600 °C, developed during the pre-Alpine history with a *P–T* path referred to a fast tectonic exhumation after the Variscan collision (Franceschelli et al. 2004). Other subsurface occurences in the geothermal fields of the Monte Amiata (Pandeli and Pasini 1990; Batini et al. 2003; Brogi 2008) sampled Permian metasediments only (Figs. 2 and 5).

3.2 Palaeozoic rocks in the External Ligurian Units

To complete the catalogue of the pre-Mesozoic rocks of the Northern Apennines, we mention here the occurence of Palaeozoic rocks within the so-called External Ligurian units (Elter et al. 1966; Molli 2008; Malavieille et al. 2016). These have been interpreted as derived from the Late Cretaceous tectono-sedimentary reworking the former Ligurian-Tethys Ocean Continent Transition (OCT) crust (Sturani 1973; Elter 1975; Molli 1996; Marroni et al. 1998), where lower and upper continental crust rocks were included. The lower crust is represented by gabbro-derived mafic granulite, derived from original gabbroic rocks of tholeiitic affinities with evidence of crustal contamination (Meli et al. 1996; Marroni and Tribuzio 1996; Montanini 1997). Sm/Nd mineral whole-rock isochron age at 291 ± 9 Ma dated the emplacement of gabbro at intermediate crustal levels, reflecting a close age, paragenetic and compositional resemblance with the gabbro-derived granulite of the Ivrea Zone (Marroni et al. 1998). The intrusive mafic complex underwent subsolidus re-equilibration under granulite-facies conditions (P = 0.7–0.8 GPa, 800–900 °C), with an evolution characterized by temperature and pressure decrease (Marroni et al. 1998). The felsic granulites have a quartz-feldspathic composition and consist of mesoperthitic to perthitic feldspar, quartz and garnet (up to 15%) with isotopic compositions approaching those of the granulite-facies basement metasediments from the Ivrea Zone (Voshage et al. 1987). Other lower-crustal rocks may be found within the Cretaceous coarse-grained deposits called Salti del Diavolo Conglomerate Fm. (Elter et al. 1966; Marroni et al. 2001) where two-mica gneiss and biotite-sillimanite kinzigitic paragneiss were described. Finally, the upper continental crust is mainly documented by granitoids with a wide variety of rock-types, ranging from two-mica leucogranite (volumetrically dominant), to biotite-bearing granodiorite and rare biotite-bearing tonalite to diorite. The two-mica leucogranites were emplaced at 310–280 Ma based on K/Ar and Rb/Sr muscovite ages (Ferrara and Tonarini 1985).

4 The Palaeozoic basement in Calabria and Southern Apennines

The southern Apennines-Calabria-Peloritani chain comprises oceanic and continental-derived cover and basement units overthrusted upon the Adriatic continental crust (Amodio-Morelli et al. 1976; Dewey et al. 1989; Bonardi et al. 2001; Rossetti et al. 2004; Iannace et al. 2007; Carminati et al. 2012; Turco et al. 2012; Vitale and Ciarcia 2013).

The different units of the Apennines-Calabria-Peloritani chain may be grouped (Fig. 6) from top to bottom in:

1. *Continental units belonging to the Calabria-Peloritani terrane,* part of the former AlKaPeCa microplate (Bouillin 1984; Michard et al. 2002; Handy et al. 2010; Vitale and Ciarcia 2013; Cirrincione et al. 2015; Critelli 2018). These units are made up of a Palaeozoic basement and locally Mesozoic cover (e.g. Innamorati and Santantonio 2018), showing different degrees of Tertiary-age deformation and metamorphism which ranges from very low grade to HP/LT peak conditions (Serre, Sila, Aspromonte, Peloritani, Castagna; Bagni, Africo-Polsi Bonardi et al. 2001; Somma et al. 2001; Langone et al. 2006; Heymes et al. 2008);
2. *Oceanic and OCT-derived units* showing different degrees of Tertiary-age metamorphism, ranging from very low grade to HP/LT peak conditions (Diamante-Terranova, Gimigliano, Malvito and Frido units, e.g.

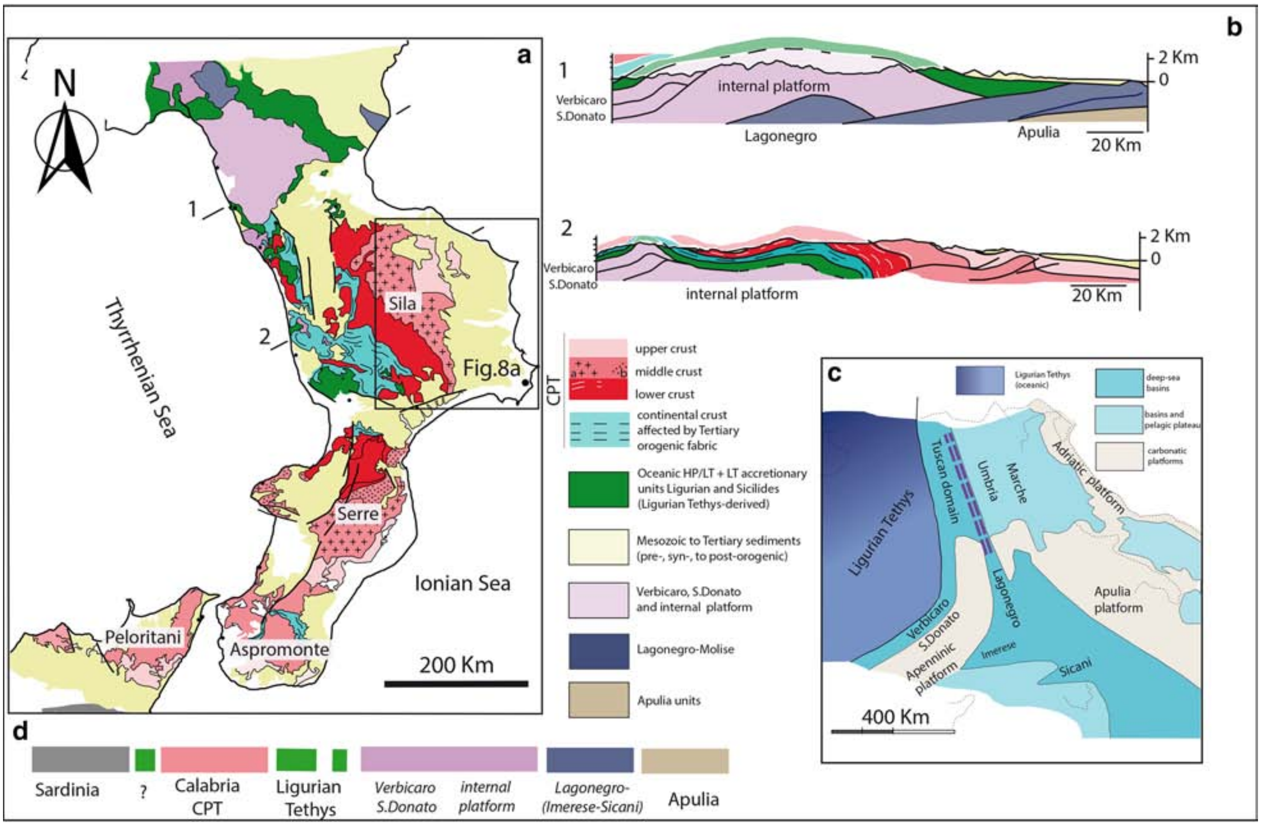


Fig. 6 **a** Schematic geological map of Southern Apennines-Calabria-Peloritani showing the main groups of tectonic units (based onAmodio-Morelli et al. 1976; Bonardi et al. 2001; Vitale and Ciarcia 2013); **b** regional cross-section of the northern Calabria (based on Elter, Scandone 1980); **d** restoration of the Central Mediterranean region showing the distribution of the Central and Southern Apennine platforms and basins (after Patacca and Scandone 2007). The palinspastic reconstruction shows the position of Lagonegro-Imerese-Sicani and the northward prolongation between Tuscan and Umbria-Marche domains. This Mesozoic setting includes inheritance of former Permian precursors related to a major regional fault system, whose location is indicated as violet dashed line see text; **e** simplified paleotectonic frame showing the main domains involved in Southern Apennines-Calabria-Peloritani orogenic building and their relative pre-Mesozoic position

Cello et al. 1996; Liberi and Piluso 2009; Cavalcante et al 2012; Vitale et al. 2013, 2019; Laurita et al. 2014; Fedele et al. 2018; Tursi et al. 2020);

3. *Distal to proximal Adria-derived continental cover units* (Lungro-Verbicaro, Cetraro, Apenninic carbonate platforms and Lagonegro-Molise basin-derived units) forming the southern Apennines to the E, and the Sicily-Maghrebian Chain to the W (Mazzoli et al. 2001; Iannace et al. 2007; Vitale and Ciarcia 2013; Vitale et al. 2019), respectively. The most internal part of this group of units (Lungro-Verbicaro and Cetraro) shows HP/LT metamorphism. Differently, the others (Apenninic carbonate platforms and Lagonegro-Molise basin-derived units), in a lowermost position, were deformed at shallow crustal depths (T less than c. 250 °C);
4. *Foreland units*, represented in the Apulia region and southern Sicily (Patacca and Scandone 2007; Bernoulli 2001; Catalano et al. 1991, 1995).

Pre-Mesozoic rocks find their best exposures and the minor degree of orogenic reworking in the uppermost unit of the Calabria-Peloritani terrane (Bonardi et al. 2001; Appel et al. 2011) and in particular in the Sila and Serre massifs (Figs. 6, 7, 8 and 9), where a nearly complete crustal section with an estimated total thickness of 20–25 km has been documented (Schenk 1990; Grässner and Schenk 2001; Caggianelli and Prosser 2001; Caggianelli et al. 2013 and ref. therein). The crustal section can be broadly subdivided into three crustal levels (Figs. 7, 8 and 9).

4.1 The lower crustal units

The lower crust (up to 8 km thick) mainly includes granulites and migmatitic paragneisses with interleaved marbles and metabasites (Caggianelli et al. 1991; Kruhl and Huntemann 1991). Mafic granulites (Fig. 7a) occur at the lowermost levels, representing former gabbros that underplated the Calabria

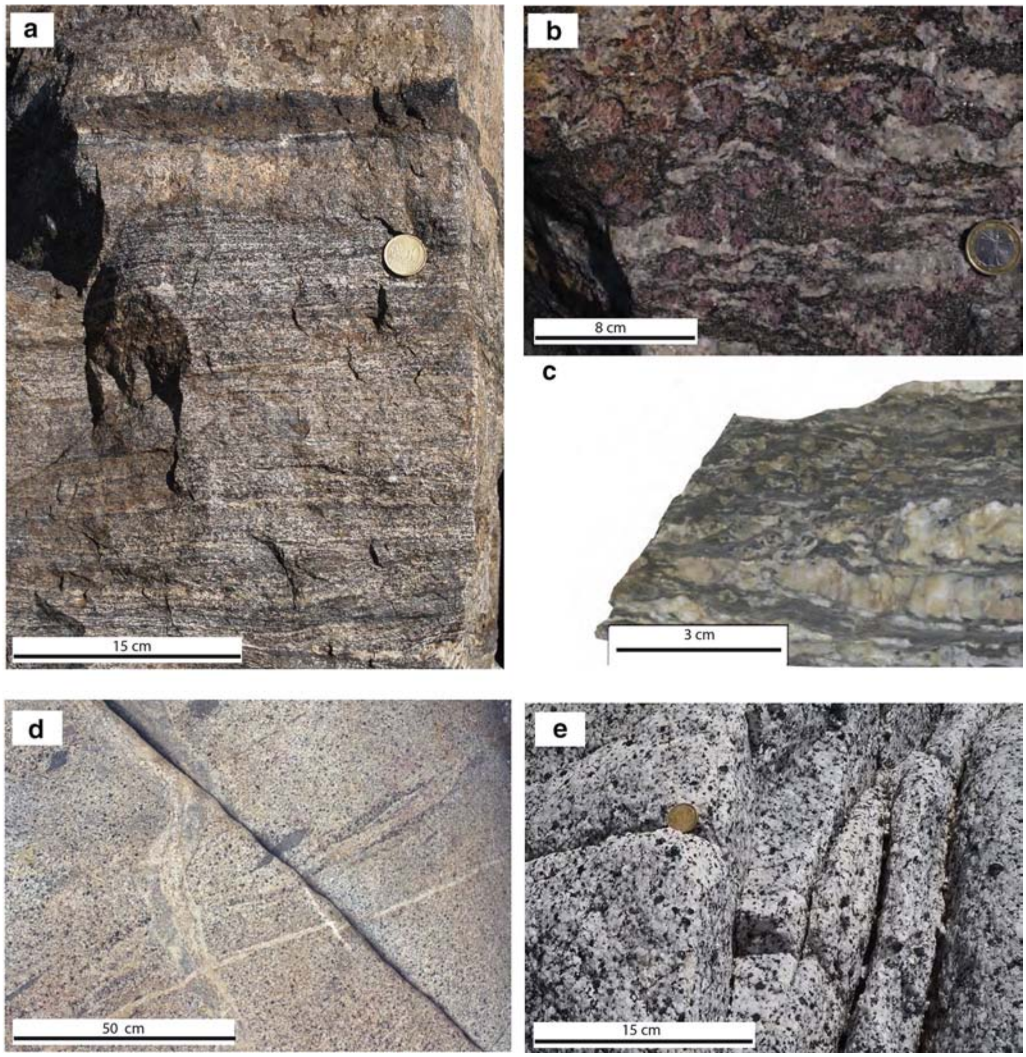


Fig. 7 Calabria lower (**a**–**c**) and intermediate crust (**d**, **e**): **a** granulitic metagabbro (Curinga); **b** felsic granulite, garnet-sillimanite rich fels (Marina di Vibo); **c** Bt + Pl + Kfs + Sil + Grt ± Ms bearing migmatitic paragneiss (kinzigite Auct.); **d** tonalite emplaced at middle crust level and characterized by a dominant magmatic fabric (Ioppolo); **e** foliated peraluminous granodiorite (Parghelia)

continental crust (Fiannacca et al. 2019). Felsic granulites (Fig. 7b, c) derive from original arenites whereas migmatitic paragneisses derive from a pelitic protolith (Caggianelli et al. 1991). According to Schenk (1989), peak metamorphic conditions of 790 ± 30 °C and c. 0.75 GPa were attained at the bottom of the Serre massif crust section at 300 ± 10 Ma. As an effect of the intense thermal perturbation, the fertile metapelitic rocks underwent widespread partial melting, mostly under water undersaturated conditions by muscovite and biotite breakdown reactions. The partial melting, estimated to a maximum degree of 60%, was responsible for the genesis of peraluminous granitic melts (Caggianelli et al. 1991). After the metamorphic peak the lower-crustal rocks recorded isothermal decompression of c. 200 MPa (Schenk 1989). From 290 Ma to Oligocene, a slow isobaric cooling occurred, when the final exhumation took place by extension and erosion (Thomson 1994; Festa et al. 2003).

4.2 The middle crustal units

The intermediate crust (Fig. 7d, e) essentially includes a succession of dominant calc-alkaline and minor strongly peraluminous granitoids (Rottura et al. 1990). These range in composition from tonalite to monzogranite with only minor mafic bodies of amphibole gabbro. In the Serre massif, the cumulative thickeness of the granitoids amounts to c. 13 km. Here the contact

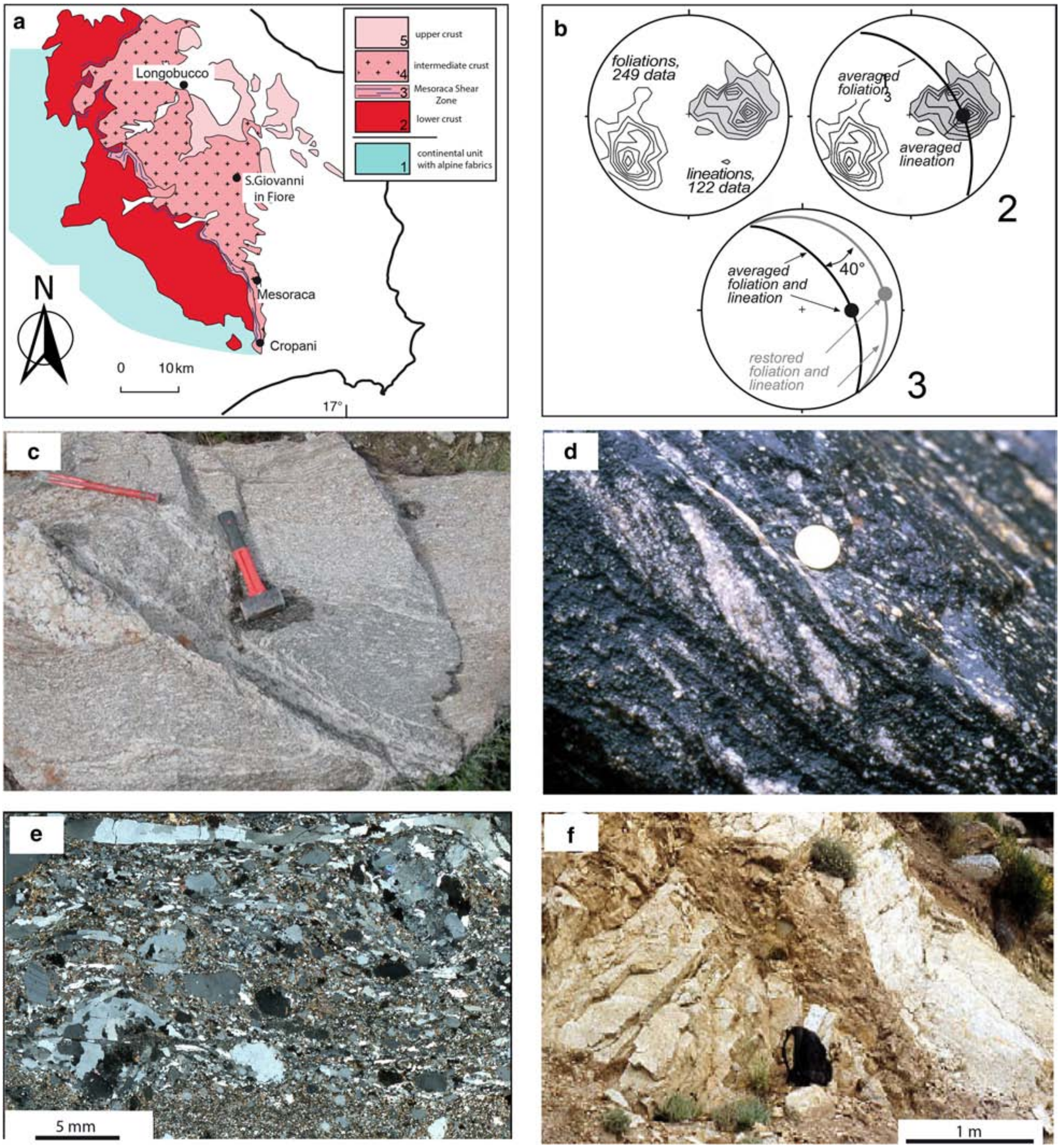


Fig. 8 **a** Sketch map of the Sila Massif: (1) continental unit affected by pervasive Tertiary orogenic fabrics; (2) lower crustal section (high grade metamorphic rocks), (3) Mesoraca Shear Zone, (4) intermediate crustal section (leucogranite, gabbro-diorite; metaluminous to strongly peraluminous granite and granodiorite); (5) upper crustal section of low grade metamorphic rocks. **b** Equal area, lower hemisphere stereonets showing structural data of the Mesoraca Shear Zone. (1) Poles of the main foliation and mineral and/or stretching lineation in migmatitic paragneiss, foliated granodiorite and mylonites. Contouring of lineation and foliation data indicated by grey and white areas, respectively; contouring interval (2%) equals the maximum of the data distribution contoured at > 16%; (2) best fit of foliation and lineation data; (3) back-rotation of the averaged mean foliation and lineation obtained by assuming a horizontal rotation axis parallel to the strike of the mean foliation (details in Liotta et al. 2008). **c** Melt-present deformation structures in granodiorites representing the early stages of Mesoraca Shear Zone deformation; **d** S/C structures indicating a top-to-the-west sense of shear in mylonites of the Mesoraca Shear Zone; **e** thin section scan (crossed polars) of a granodiorite involved in the Mesoraca shear zone. Quartz ribbons and porphyroclasts of quartz and feldspars locally showing core and mantle structure can be observed; **f** undeformed pegmatite and porphyritic dyke, intruding wall-rocks and the previously emplaced granitoids, respectively. The dyke (dated between c.290–280 Ma in Liotta et al. 2008) is exposed close to the Arvo Lake

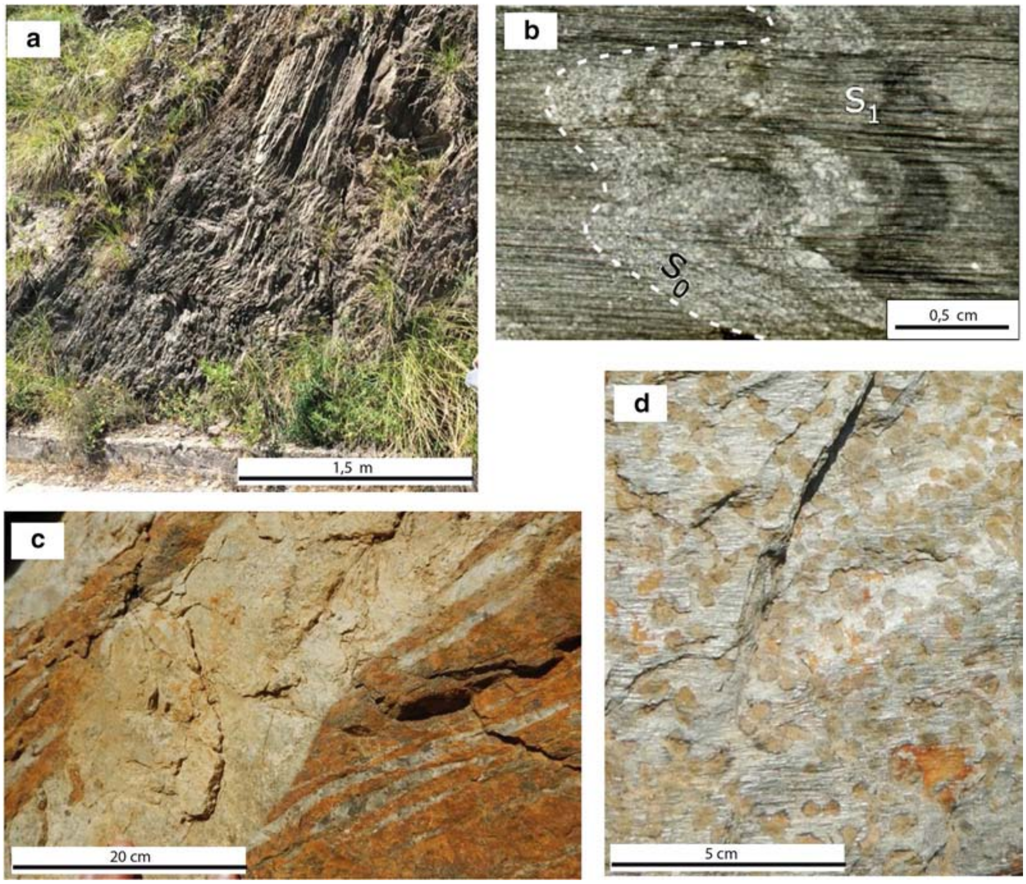


Fig. 9 **a** Calabria upper crust phyllites and metasandstones of the low grade unit of Stilo-Bivongi (Fiumara di Stilaro); **b** microscopic fabric of phyllite samples collected close to Bivongi village. S1 foliation in phyllite is parallel to axial plane of folds affecting S0 layering; **c** decimetre-thick granodioritic dyke within the low-grade host rock close to the margin of the Serre batholith near Stilo; **d** spotted schist in the aureola of Serre batholith near Stilo (see details Festa et al. 2013)

between the granitoids and the lower-crustal metapelites is characterized by a wide migmatitic border zone, locally affected by shearing. In contrast, the contact with the upper-crustal metapelites is sharp and marked by a wide metamorphic aureole. Granitoid emplacement took place incrementally from 306 to 295 Ma, during the decompression event recorded by the lower-crustal rocks (Langone et al. 2014). The older tonalitic magma was emplaced at deeper levels and underwent a slow cooling history, with development of an intense fabric anisotropy (see below). The younger granodioritic magma emplaced at shallower level and was subjected to a rapid cooling, preventing the development of significant bulk ductile deformation (Caggianelli et al. 2000).

Lower and intermediate crust are separated by a regional-scale shear zone (Fig. 8) decribed in the Sila massif (Liotta et al. 2008), but also recognizable in the Serre (e.g. Fornelli et al. 2011; Festa et al. 2012) where it essentially corresponds to the Quartz-Dioritic Gneiss unit of Graessner and Schenk (2001). In the Sila massif, the shear zone, called Mesoraca Shear Zone, may be traced for more than 60 km, with a thickness of more than 4 km (Liotta et al. 2004; Festa et al. 2006). Simultaneous deformation and magmatism which involved hybrid magmas with a dominant contribution from a mantle source (Liotta et al. 2008) is constrained by U/Pb dating of zircon and monazite, at 304–300 Ma, coeval with the regional metamorphic peak (Graessner et al. 2000). The deformation within the shear zone, which remained steady during magma crystallization and cooling in subsolidus conditions, was associated to a top-to-the-W sense of shear (Fig. 8b-d), in the present geographic coordinates (Liotta et al. 2004, 2008; Festa et al. 2006). Foliated granitoids and wall rocks were then intruded by poorly foliated Hbl-gabbro and, finally, by undeformed leucogranite, pegmatite and felsic porphyritic dykes (Fig. 8e). U/Pb zircon dating

(See figure on next page.)

Fig. 10 **a** Schematic geological map of Alpine and Crystalline (or "Autochtonous") Corsica. Developed after Cocherie et al. (2005) and Lin et al. (2018). **b** Simplified Mesozoic paleotectonic frame with indication of the domains discussed in the text and their relative positions. **c** Geological map of the Alpine Santa Lucia nappe, with (c1) a tectonic scheme of its Alpine orogenic framework (continuous black line trace of cross-section in (**d**); **d** Cross-section of Santa Lucia Nappe between E Corte and Santa Lucia di Mercurio, vertical = horizontal scale; **e** Schematic cross-section through the Santa Lucia pre-Alpine basement (i.e. the western part of Santa Lucia Nappe, cfr. (**d**). The orientation of foliation and lineation and the shear sense remain nearly constant during all deformation stages and states, solid-state (High- and Low-Temperature HT, LT, magmatic to submagmatic), as documented in the equal-area projection plots (see also Zibra et al. 2010; 2102 and unpublished data IZ and GM). **f** *P–T* conditions for the crystallization of the Granite Complex, for the Diorite-Granite Complex and for M1 and M2 assemblages in the Santa Lucia Shear Zone (modified after Zibra et al. 2012). Al_2SiO_5 phase relations are calculated using winTWQ. The α-/β-quartz transition is from Gross and Heege (1973). Equilibrium (1) is the geothermometer Alm + Phl = Prp + Ann; equilibrium (2) is the geobarometer 2 Alm + Grs + 3 βQtz = 6Fs + 3 An. J: solidus for the tonalite system (biotite-plagioclase-quartz) from Johannes (1984); other fluid-absent solidi for granitoid systems are taken from Clemens and Wall (1981): CW; Vielzeuf and Montel (1994): VM; (Vielzeuf 2001): VS.W = H_2O-saturated haplogranite solidus (Singh and Johannes 1996). The maximally relaxed geotherm (V∞) for reasonable heat supply after thrusting is after (Thompson and England 1984). V∞ coincides with the geotherm for the c. 280–260 Ma crust of the Alpine area (SS08) from Schuster and Stüwe (2008). These geotherms are used here to estimate the pressure for the retrograde evolution along the BCVZ (Bocca di Civenti Shear Zone)

of the last intrusions indicates an emplacement age of 281 ± 8 Ma, providing a minimum estimate for the end of the shear zone activity at mid-crustal level (Liotta et al. 2008).

4.3 The upper crustal units

The upper crust (Fig. 9), probably having an original thickness of c. 7 km, is essentially represented in the Serre massif by two Variscan tectonic units (Colonna et al. 1973): a first one, the Stilo-Pazzano phyllite is a former Cambro-Ordovician to Carboniferous succession, made up of pelite with intercalations of volcanics rocks and impure limestone layers, subjected to a low-grade Variscan metamorphism (Bouillin et al. 1987; Spalletta and Vai 1989); the second one, the Mammola paragneiss unit, is mainly represented by micaschists and meta-greywackes with minor amphibolites intercalations that recorded a dual peak metamorphic evolution from medium- to low-pressure series (Festa et al. 2004; Angì et al. 2010; Tursi et al. 2020). Both tectonic units were overprinted by contact metamorphism, with peak T condition of 590 °C at a pressure of 0.175–0.200 GPa, in response to the thermal perturbation produced by the emplacement of the granodioritic magma that was also responsible for the genesis of an ample rim fold in the wall rocks bordering the pluton (Festa et al. 2013).

Relevant for the upper Palaeozoic tectonic history is to recall that below the Calabria-Peloritani Terrane and oceanic-derived units of the Apenninic-age nappe architecture (Fig. 6b, c) there are some units correlated with those of the Tuscan domains (i.e. Verbicaro and S. Donato, Elter and Scandone 1980; Iannace et al. 2007) and, in a lowermost position, some others (Lagonegro units) in north Calabria and the Imerese-Sicani in Sicily (Scandone 1975; Catalano et al. 1995), in which lower to upper Permian carbonate platforms and deep-water basins are documented (Imerese-Sicani) or inferred (Lagonegro) (Catalano et al. 1991, 1995; Vai 2001).

5 The Palaeozoic basement in Alpine Corsica

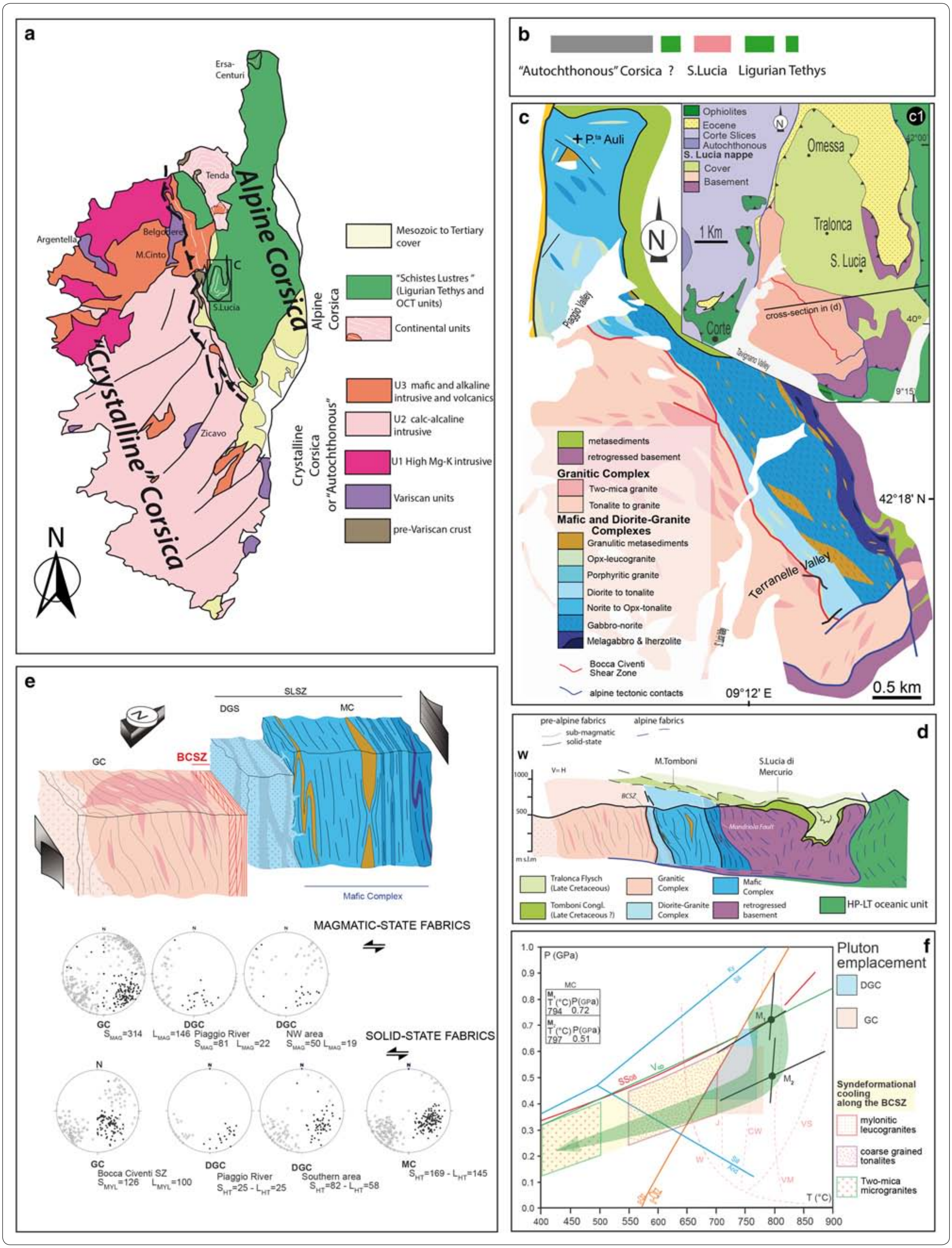
Corsica is subdivided into two different geological domains (Durand Delga 1984; Molli and Malavieille 2011): the northeast of the island, forming the so called "Alpine Corsica", and the western part, identified as "Autochthonous", "Variscan" or "Crystalline" Corsica (Fig. 10a).

The western domain mainly consists of Variscan granitoids and Permian magmatic rocks, intruded within the Variscan and pre-Variscan basement (Durand Delga 1984; Rossi et al. 2009). The main features and tectonic context of the Variscan basement and of the syn- to post orogenic magmatism (i.e. the U1, U2 and U3 suites) of the Corsica-Sardinia Batholith have been already introduced.

It is noteworthy that the U2 and U3 intrusions in the "Authochthonous" Corsica were emplaced within their own volcanic apparatus at shallow crustal depth (Rossi and Cocherie 1991; Rossi et al. 1993) whereas the relationships between shear zones in the lower crust and pluton emplacement can be only directly observed and reconstructed in Central Alpine Corsica, in the Santa Lucia Nappe (Libourel 1988a, b; Caby and Jacob 2000; Zibra et al. 2010; Rossi et al. 2015).

5.1 The Santa Lucia Nappe

The Santa Lucia Nappe is a continental-derived unit exposed a few kilometers north east of Corte (Fig. 10) in Central Alpine Corsica (Durand Delga 1984). Toward the north the unit overthrusts the Caporalino Eocene Flysch (Puccinelli et al. 2012), to the east it is overlain by an HP/LT oceanic unit (Monte Piano Maggiore, Vitale Brovarone et al. 2013) whereas, it is separated to the west,


a
Ersa-Centuri
Tenda
Argentella
Belgodere
M.Cinto
S.Lucia
Zicavo
Alpine Corsica
"Crystalline" Corsica
Mesozoic to Tertiary cover
"Schistes Lustres" (Ligurian Tethys and OCT units)
Continental units
Alpine Corsica
U3 mafic and alkaline intrusive and volcanics
U2 calc-alcaline intrusive
U1 High Mg-K intrusive
Variscan units
pre-Variscan crust
Crystalline Corsica or "Autochthonous"
b
"Autochthonous" Corsica ? S.Lucia Ligurian Tethys
c
P.ta Auli
Piaggio Valley
Ophiolites
Eocene
Corte Slices
Autochthonous
S. Lucia nappe
Cover
Basement
c1
Omessa
Tralonca
S. Lucia
Corte
1 Km
cross-section in (d)
Tavignano Valley
metasediments
retrogressed basement
Granitic Complex
Two-mica granite
Tonalite to granite
Mafic and Diorite-Granite Complexes
Granulitic metasediments
Opx-leucogranite
Porphyritic granite
Diorite to tonalite
Norite to Opx-tonalite
Gabbro-norite
Melagabbro & lherzolite
Bocca Civenti Shear Zone
alpine tectonic contacts
Terranelle Valley
42°18' N
09°12' E
0.5 km
d
pre-alpine fabrics
sub-magmatic
solid-state
alpine fabrics
W
M.Tomboni
S.Lucia di Mercurio
BCSZ
Mandriola Fault
Tralonca Flysch (Late Cretaceous)
Tomboni Congl. (Late Cretaceous ?)
Granitic Complex
Diorite-Granite Complex
Mafic Complex
retrogressed basement
HP-LT oceanic unit
e
SLSZ
DGS
MC
GC
BCSZ
Mafic Complex
MAGMATIC-STATE FABRICS
GC S_{MAG}=314
DGC L_{MAG}=146 Piaggio River S_{MAG}=81 L_{MAG}=22
DGC NW area S_{MAG}=50 L_{MAG}=19
SOLID-STATE FABRICS
GC Bocca Civenti SZ S_{MYL}=126 L_{MYL}=100
DGC Piaggio River S_{HT}=25 - L_{HT}=25
DGC Southern area S_{HT}=82 - L_{HT}=58
MC S_{HT}=169 - L_{HT}=145
f
P (GPa)
T (°C)
MC
M_1 T (°C) P (GPa) 794 0.72
M_2 T (°C) P (GPa) 797 0.51
Pluton emplacement
DGC
GC
Syndeformational cooling along the BCSZ
mylonitic leucogranites
coarse grained tonalites
Two-mica microgranites

from the continental basement and cover units of the Corte slices and the overlying HP/LT oceanic unit by the Central Corsica Fault (Molli and Malavieille 2011).

The Santa Lucia Nappe (Fig. 10) includes a pre-Mesozoic basement overlain by a Cretaceous metasedimentary cover (Amaudric Du Chaffaut and Saliot 1979; Rieuf 1980; Durand Delga 1984; Libourel 1988a, b; Lin et al. 2018).

The nappe experienced a low-grade Alpine metamorphism (RSCM temperature lower than 330 °C, Vitale Brovarone et al. 2013) and it is structurally subdivided into two portions, separated by the NW–SE trending intra-nappe Mandriola Fault of Alpine age (Fig. 10c, d). The eastern portion of the nappe experienced a higher finite strain during Alpine tectonics (Amudriac du Chaffault and Saliot 1979; Egal 1992; Caby and Jacob 2000; Zibra 2006), with pre-Mesozoic basement strongly retrogressed and widely affected by Alpine low grade fabrics (Fig. 10d). On the contrary, the western portion of the nappe is made up of two main tectonomagmatic suites that largely preserve the preorogenic tectonic grain and fabrics (Libourel 1988a, b; Caby and Jacob 2000; Zibra 2006).

A gabbroic layered intrusion, called Mafic Complex, includes melagabbros with subordinate hornblendites and pyroxenites, to the east, and progressively more evolved components to the west, where the main gabbro-norite central unit grades into quartz-norite to Opx-tonalite, in turn intruded by amphibole-rich diorite–tonalite and by porphyritic granite (Fig. 11). This lithologically heterogeneous roof zone of the Mafic Complex is known as Diorite–Granite Complex (Zibra et al. 2010, 2012). Slivers of mantle lherzolite occur near the base of the Mafic Complex (Libourel 1988a, b; Caby and Jacob 2000; Montanini et al. 2014), and lenses of granulite-facies metapelitic country rocks occur throughout the Mafic Complex, being interlayered with gabbros and dioritic to granitic rocks (Fig. 10c, d). Mafic Complex and Diorite-Granite Complex are juxtaposed to a westernmost Granite Complex (Figs. 10c, d and 11), which mainly consists of Bt-bearing leucotonalites intruded by two-mica microgranitoids. The contact between Diorite-Granite Complex and Granite Complex (Fig. 10c, d) is represented by an upper-greenschist facies mylonitic zone called Bocca di Civenti Shear Zone (Zibra et al. 2010; Beltrando et al. 2013).

Zircon U–Pb geochronology on the Mafic Complex (Paquette et al. 2003; Zibra 2006; Seymour et al. 2016) documents the gabbronorite emplacement at 280–286 Ma (time range of U3 suite of the Corsica-Sardinia Batholith). The granulitic metasediments within the Permian gabbronorites were deformed during the thermal re-equilibration of the lower crust as country rock septa, reaching P–T conditions of 0.7 ± 0.1 GPa and 800 ± 50 °C (Libourel 1988a, b; Rossi et al. 2006; Zibra et al. 2010). A zircon U–Pb age population at 286 ± 1 Ma in metapelites is interpreted to date granulite-facies metamorphism (Paquette et al. 2003), whereas zircon U/Pb data (Seymour et al. 2016, Fig. 10f) record emplacement and crystallization of the Granite Complex in the 310 to 270 Ma time span, at ~ 730 °C, in broad agreement with the previously-published ~ 278 Ma muscovite $^{40}Ar/^{39}Ar$ cooling age (Zibra 2006; Seymour et al. 2016).

Structural studies of the area (Libourel 1988a, b; Caby and Jacob 2000; Zibra 2006; Zibra et al. 2010; 2012) document the same lineation trend and kinematics (sinistral shear) (Fig. 10e) between the two tectono-magmatic crustal domains (Mafic Complex + Diorite-Granite Complex and Granite Complex) which were juxtposed along an oblique-slip shear zone (in present orientation) known as Santa Lucia Shear Zone, with magmatism, metamorphism and shearing that interacted over a temperature range from 800 to 400 °C, when deformation was localized along the Bocca di Civenti Shear Zone (Zibra et al. 2010, 2012).

More recent contributions of Beltrando et al. (2013) and Seymour et al. (2016), however, proposed to divide the Mafic Complex from the Diorite-Granite Complex and Granite Complex by an amphibolite to greenschist facies shear zone, the Belli Piani Shear Zone. These authors based the presence of this structure (with Triassic-Jurassic activity) on three $^{40}Ar/^{39}Ar$ step-heating analyses on amphiboles (Beltrando et al. 2013) and using zircon, rutile, and apatite $^{206}Pb/^{238}U$ depth profiling coupled with garnet trace-element diffusion modeling (Seymour et al. 2016).

The two contributions, however, strongly differ in the envisaged role of the Belli Piani Shear Zone, which has been considered to produce no significant exhumation during its activity, thereby residing at a broadly constant depth (Beltrando et al. 2013) or, alternatively, to produce a synkinematic juxtaposition of the Diorite-Granite and Granite Complexes against the hot footwall of the Mafic Complex and whole-sale conductive steepening of geothermal gradients (Seymour et al. 2016).

However, while the dating of a deformation event relies on the isotopic analysis of synkinematic minerals that belong to the metamorphic assemblage associated with deformation (e.g. Di Vincenzo et al. 2004; Cenki-Tok et al. 2014; Erickson et al. 2015; Papapavlou et al. 2017), to date no direct dating of minerals synkinematic with any of the shear-related microfabrics exposed in the Santa Lucia basement is available, therefore other tectonic scenarios may be proposed to better fit the data of Beltrando et al. (2013) and Seymour et al. (2016). In particular, similarly

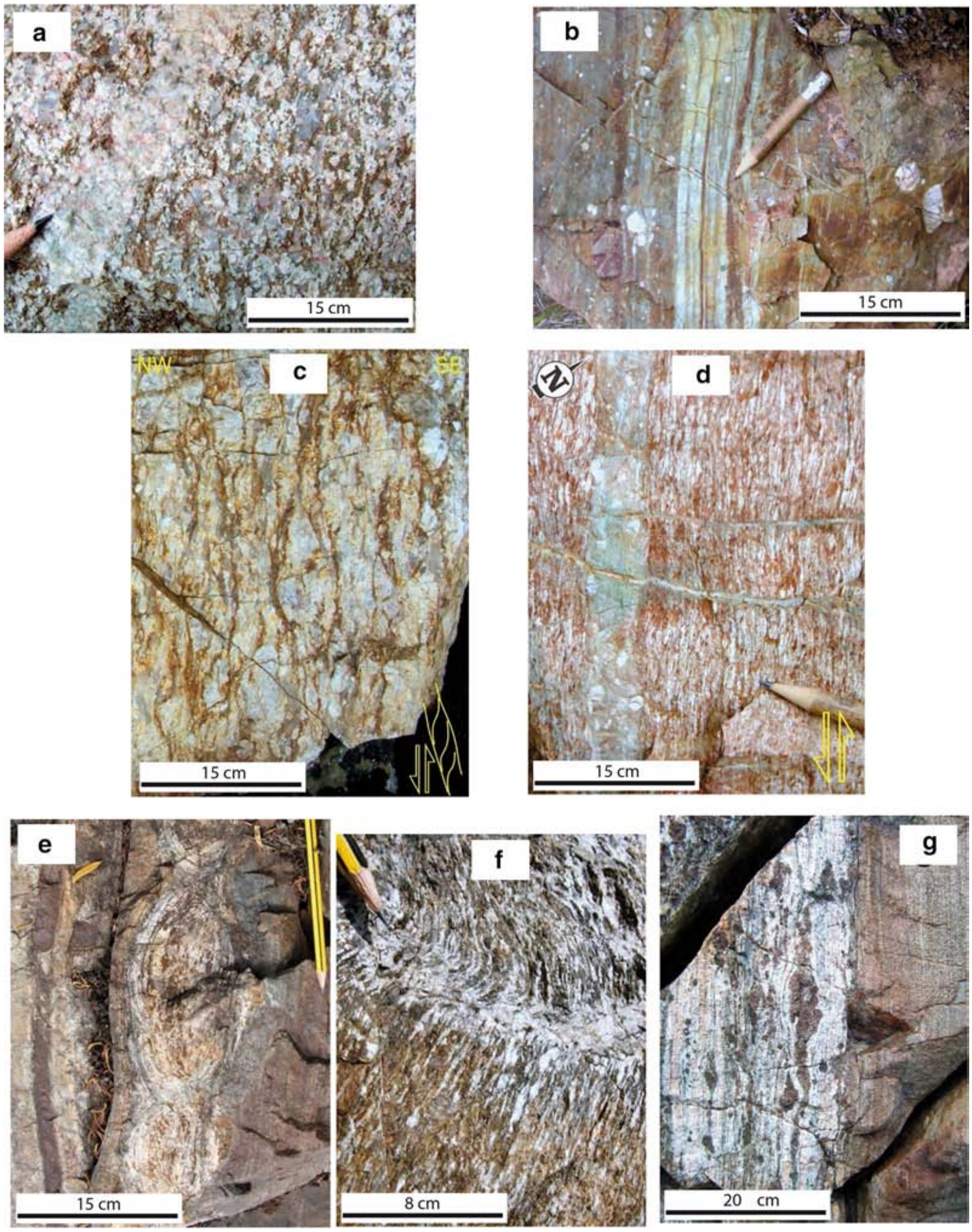


Fig.11 **a** Coarse-grained tonalite of the Granite Complex, showing magmatic to submagmatic fabric which is highlighted by aligned euhedral to subhedral plagioclase phenocrysts, aggregates of biotite flakes and quartz ribbons; **b** Sheared pegmatite, intruded into gabbro-norite, from western margin of SLSZ (see Fig. 10c, d). This pegmatite was intruded and sheared along the SLSZ during isothermal decompression from ~ 0.7 to 0.5 GPa at 800 °C (Zibra et al. 2012 and Fig. 10e,f); **c** Coarse-grained granitic gneiss along the contact between Granite Complex and Diorite-Granite complex (Bocca Civenti Shear Zone). Natural section nearly perpendicular to mylonitic foliation (subvertical) and subparallel to the stretching lineation. S/C and C′ subfabrics indicate sinistral shear; **d** Mylonitized intrusive contact between a granite pegmatite and host gabbro–norite, near the top of the SLSZ. Sinistral shear sense indicated by σ-type mantled K-feldspar (in pegmatite) and plagioclase porphyroclasts (in gabbro–norite); **e**, **f** Outcrop-scale evidence of melt-present deformation (**e**) boudinaged pegmatoid Opx-bearing tonalite, previously injected into the host melagabbro. Boudin neck is locally filled by undeformed leucotonalite (f), interpreted to have been a melt. Stretched Opx porphyroclasts may be observed adjacent to the vein boundaries; (**g**) Mylonitic fabric affecting felsic granulite and meta-gabbro-norite. Shear bands and asymmetric porphyroclast systems show sinistral shear sense (more details in Zibra et al. 2010)

to what has been proposed for analogous lower- to mid-crustal settings and the same age-related issues of the Ivrea Zone (e.g. Siegesmund et al. 2008; Ewing et al. 2015) and Calabria-Peloritani Terrane (Festa et al. 2004; Liberi et al. 2011), it is possible that heat advection by hot fluids and melts, as well as conductive heating from the rising asthenosphere during Mesozoic rifting events may have produced thermal pulses (with partial resetting of Permian ages) in the Santa Lucia basement. This is in line with what has already been proposed (Rossi et al. 2006) and supported by the presence of undeformed dolerite dykes with MORB affinity, and thus of Jurassic age (Zibra 2006), cross-cutting the main foliation of the Mafic Complex (Caby and Jacob 2000; Paquette et al. 2003; Zibra 2006). These dolerites characterized by chilled margins represent, moreover, further evidence that during Jurassic the Santa Lucia basement was already cooled below 350 °C.

It is also worth noticing that both studies (Beltrando et al. 2013; Seymour et al 2016) do not provide any structural data for the Belli Piani Shear Zone (e.g. micro to mesoscopic elements and their kinematic characterization) and both papers used maps and structural data reported in Zibra (2006) and Zibra et al. (2010, 2012) which, similarly to what was documented by Libourel (1988a, b a, b) and Caby and Jacob (2000), demonstrated the complete kinematic coupling between the two tectono-magmatic crustal domains.

We therefore support the interpretation that the western portion of the Santa Lucia Nappe shows a well preserved fragment (up to 3.5 km thick) of a Permian lower-middle crust with record of deformation of a crustal scale shear zone (Libourel 1988a, b; Caby and Jacob 2000; Zibra et al. 2010; Beltrando et al. 2013).

Moreover, the original stratigraphic relationships between the metasedimentary cover (Tomboni Conglomerate and Tralonca Flysch) and basement in the western portion of the nappe (Rieuf 1980; Caby and Jacob 2000) and in the highly strained eastern domain (Caby and Jacob 2000; Zibra 2006; cfr. Murato sub-unit of Beltrando et al. 2013), allowed some inferences about original geometries. The sub-vertical present-day attitude of foliations and tectonic grain of the basement in the western portion of the Santa Lucia basement could be considered as quite close to its original orientation at the time of cover deposition (Fig. 10d). Conversely, the basement-cover is transposed and parallelized to the fold limbs in the east (subparallel relationships between bedding of metasediments and basement-cover contact quoted in Beltrando et al. 2013 within the Murato sub-unit i.e. our eastern nappe structural domain). The presence of exhumed brittle fault-rocks at the top of the Santa Lucia basement, however, clearly indicates a shallow crust deformation occurred during Mesozoic (Caby and Jacob 2000; Zibra 2006; Beltrando et al. 2013). To date, no structural data are available to directly constrain the post-Permian to Cretaceous tectonic and kinematic history evolution in the upper crust for the studied domain and therefore an accurate restoration of the original Permian geometries is at the moment not possible. However, *P–T* data from the Santa Lucia basement (Libourel 1988a, b; Zibra et al. 2010 and this paper) document a minimum of ~ 0,3 GPa of isothermal decompression during Permian tectonics (Libourel 1988a, b; Caby and Jacob 2000; Zibra et al. 2010, 2012; Beltrando et al. 2013), supporting an oblique-slip kinematics and an overall transtensional setting (Libourel 1988a, b; Caby and Jacob 2000; Zibra et al. 2010).

Therefore, we consider that the Santa Lucia Shear Zone, as a whole, has accommodated high finite strain on a 1 km-wide crustal domain representing at the time of shearing a lower-middle crustal segment of a Permian crustal-scale regional fault.

It is noteworthy to notice that rock-types similar to those found in the Santa Lucia Nappe may be also recognized in the Ersa-Centuri continental slice (Malavieille 1983; Harris 1985; Lahondere 1996) within the blueschist Schistes Lustres (Vitale Brovarone et al. 2013), in the North-West of Cap Corse. This Alpine unit includes intercalations of mafic rocks (Gneiss of Centuri), and kinzigites (Gneiss of Ersa) essentially composed of aluminous paragneisses. The latter displays Ti-biotite + plagioclase-sillimanite + graphite + quartz-garnet-Kfeldspar andalusite(?)-cordierite(?) pre-Alpine mineral association, which was largely obliterated during the Alpine deformation and metamorphism. These occurrences possibly document that the width of crust originally affected by the Santa Lucia Shear zone was larger than what is now preserved in the western part of the Santa Lucia Nappe.

6 Late Palaeozoic tectonic framework in central Mediterranean: a discussion

6.1 Pre-Mesozoic setting in the Northern Apennines: some key remarks

The major limits for reconstructing the pre-Mesozoic history of the Northern Apennines are the uneven and geographically limited occurrences of Palaeozoic rocks and their strong involvement in the Tertiary orogenic building (Figs. 2, 5c). Nevertheless, on the basis of previously presented data, some relevant tectonic features may be highlighted about the continental area, which subsequently evolved into the Mesozoic Tuscan Domain.

The first one is related to the presence of two late Palaeozoic sedimentary cycles, developed within fairly narrow continental or epicontinental domains where the activity of the steep bounding faults produced local vertical movements with erosional episodes and a migration of depocenters and source areas (Rau and Tongiorgi 1974). These depositional features have been related to pull-apart basins, within an overall transcurrent to transtensive tectonic setting (Rau 1990, 1991, 1994; Spina et al. 2019). Importantly, a connection between Upper-Carboniferous Permian basins of Tuscany and the open marine domains of Sicily (Imerese, Sicani) has been suggested (Rau 1994; Vai 2001). Moreover, the middle to upper Permian successions (second sedimentary cycle) recorded the occurences of Permian magmatism, witnessed by reworked rhyolites in the porphyritic Schists of Iano (Barberi 1966), and within the Brecce di Asciano of the Monte Pisano (Rau and Tongiorgi 1974). Further evidence for this magmatism has been recently found, as first in-situ occurences, within the Variscan basement of the Alpi Apuane (Pieruccioni et al. 2018; Vezzoni et al. 2018).

A second key point, critical, for constraining the late Palaeozoic history within the Apennines, is found in the mid-crustal tectono-metamorphic record of samples from the bottom of the Pontremoli (Anelli et al. 1994; Pandeli et al. 1994; Lo Po' et al. 2016; Molli et al. 2018) and Larderello-Travale deep boreholes (Franceschelli et al. 2004; Pandeli et al. 2005). The tectono-metamorphic history combines with the most recent geochronological studies in Pontremoli, which document a metamorphic event at 293 Ma by monazite geochronology (Lo Po' et al. 2016) close to the 285 ± 11 Ma Rb/Sr age for a muscovite associated with andalusite obtained in the Larderello micaschist (as reported in Del Moro et al. 1982).

As illustrated in Fig. 5c, the retrodeformation at a mimimum displacement of the Tertiary overthrust of the Tuscan metamorphic units along the Apenninic NE-transport direction (e.g. Molli 2008; Le Breton et al. 2017), defines Pontremoli and Larderello-Travale as two subsurface occurrences of a crustal domain originally located eastward (in present coordinates) with respect to those from which the exposed Tuscan units derived (Figs. 2c, 5c). This crustal sector (pre-Mesozoic "external" domain) includes the remnants of a pre-Apenninic regional-scale north–south (in present coordinates) crustal structure, here called the "East Tuscan Fault". This "hidden" regional structure may be constrained by its overall trend (nearly north south in present coordinates) and by the fact that the deepest parts of the Pontremoli and Larderello boreholes shared the common structural architecture, displaying a subvertical attitude of the main pre-Alpine foliations (Elter and Pandeli 1990; Conti et al. 1993; Pandeli et al. 2005; Molli et al. 2018). This structural feature, the reconstructed *P–T* history and age of Larderello-Travale (Bertini et al. 1994; Franceschelli et al. 2004) and the peculiar Pontremoli anticlockwise PT path (T_{max} before P_{max} followed by a nearly-isothermal decompression), are in line with what proposed in Lo Pò et al. (2016): i.e. the interpretation adressed the crustal record of transpressive and then transtensive deformation within a late Palaeozoic regional-scale shear zone.

The coeval tectonic evolution recorded at mid crustal depths in Pontremoli and Larderello-Travale, coupled with the tectono-sedimentary and magmatic history of the upper Carboniferous-early Permian depositional cycles of the Tuscan Metamorphic units, may be combined to suggest a large-scale paleotectonic framework in Tuscany. We are therefore suggesting the scenario of an interlinked transtensional setting (Figs. 12a, b), associated with pull-apart basins (Rau 1990, 1994), possibly kinematically connected (Fig. 12a, b) to a major regional fault here defined as the "East Tuscan Fault".

6.2 The puzzle of investigated crustal domains and the framework of the South Variscan Belt

Although the large-scale structure of the Variscan belt is still, in some areas, open to discussion (Martinez-Catalan 2011; Kroner, and Romer 2013; Franke et al. 2017; Ballevre et al. 2018; Tomek et al. 2019), most authors agree that the Corsica-Sardinia block shows in an almost complete framework all the tectonic zones of the South Variscan belt with a well-preserved orogen-scale architecture from the SW foreland in Sardinia to the hinterland and retrobelt zone in the NE of Corsica (Rossi et al. 2009; Oggiano et al. 2013). Each of the different tectonic zones are characterized by peculiar and distinctive rock-unit associations or stratigraphy with their own tectono-metamorphic and magmatic history (Lardeaux et al. 1994; Carmignani et al. 1994; Rossi et al. 2009). Below, we will insert our data and discuss their tectonic implications in a regional-scale tectonic scheme (Figs. 12,13 and 14), based on those in Burg et al. (1994), Matte (2001), Ziegler and Stampfli (2001); Martinez-Catalan (2011); Franke et al. (2017); Ballevre et al. (2018), having the Corsica-Sardinia block as reference frame. In the scheme (Fig. 14), following Matte (2001) and previous proposals (e.g. Vai and Cocozza 1986; Vai 2001), the external southernmost part of the South Variscan belt is continued eastward into the Carnic foreland fold and thrust belt (Mariotto Pasquaré and Venturini 2018). This external domain of the Variscan belt would have its hinterland, in present coordinates, within the South-Alpine crust toward the Ivrea Zone (e.g. Milano et al. 1988; Pfiffner 1993; Schmid 1993; Vai 2001; Schaltegger and Brack 2007; Spiess et al. 2010). Using this as a major regional

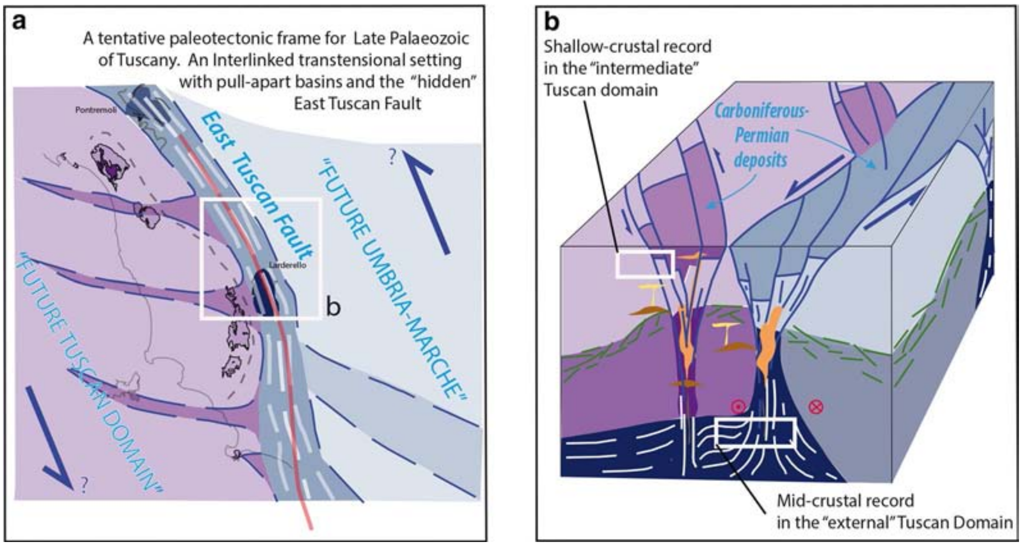


Fig. 12 **a** A tentative paleotectonic frame for late Palaeozoic in Tuscany showing the linked transtensional setting with pull-apart basins in the "intermediate" Tuscan domain and the "hidden" East Tuscan Fault whose remnants in terms of basement and cover may be found in subsurface in the northern (Pontremoli) and southern Tuscany (Larderello, Amiata). Also represented the possible trends of Permian basins in Tuscan and Umbria-Marche domains (Rau and Tongiorgi 1974; Scisciani et al. 2014); **b** geological sketch (not to scale) illustrating the inferred tectonic context of the late Palaeozoic tectonics in Tuscany. Pull-apart basins developed in splay structures of a major regional fault whose mid-crustal relict remnants may be found in Pontremoli and Larderello deep well with their Carboniferous-Permian cover in the M.Amiata subsurface occurences

constraint, Tuscany and Calabria are inserted according to their chief Variscan signature.

For Tuscany, following previous stratigraphic and structural correlations (Bagnoli et al. 1979; Vai and Cocozza 1986; Gattiglio et al. 1989; Conti et al. 1991; Pandeli et al. 1994, 2005), what is here defined as intermediate pre-Mesozoic Tuscan domain (e.g. the Palaeozoic basement associated with the Tuscan metamorphic units) has been considered as matching and correlating with the nappe zone of the Variscan belt (Gattiglio et al. 1989; Conti et al. 1991, 1993), i.e. with Central Sardinia where low-grade Cambrian to Devonian sequences may be observed (Figs. 1, 14). On the other hand, the available petrological data and age constraints support, for the inner or internal domain, its matching with the medium- to high-grade units of the axial zone (Molli et al. 2002; Lo Pò et al. 2017) and a correlation with NE Sardinia (Carmignani et al. 1994), SW Corsica (Rossi et al. 1995; Oggiano et al. 2013), the basement of Ligurian Briançonnais (Messiga et al. 1992; Cortesogno et al. 1997; Giacomini et al. 2007) and some of the basement units of the western Alps (Von Raumer 1984; Fernandez et al. 2002; Simonetti et al. 2018). Lower and upper crust rocks associated with the External Ligurian domain may consequently be referred to an even further western (in present coordinates) crustal sector adjacent to the Alpine Sesia-Austroalpine basement (Dal Piaz 1993; Pfiffner 1993; Pennacchioni 1996; Venturini et al. 1994; Hermann et al. 1997; Manzotti et al. 2017; Petri et al. 2017; Schmid et al. 2017).

In the external, pre-Mesozoic Tuscan domain, instead, the presence of a crustal scale shear zone that accomodated transpressive and then transtensive deformation during lower Permian (Lo Pò et al. 2016) may be inferred. This shear zone may be considered the mid-crustal expression of what is defined here as the East Tuscan Fault (Fig. 5b, c).

Regarding the structural grain of the Calabria-Peloritani Terrane, former attempts to match it with the Variscan belt in Sardinia were problematic [see the different solutions proposed in Vai (2001); Alvarez and Shimabukuro (2009)] due to the correlation between the medium- to high-grade Calabria rocks with those of the axial zone of NE Sardinia. However, focusing on the recent literature in terms of the age of deformation, and the magmatic and metamorphic characters, major differences between Calabria medium- to high-grade rocks and those of the Corsica-Sardinia axial zone are highlighted (see chapt. 4). Moreover, we consider the upper-crustal section in the Calabria-Peloritani Terrane, made of a low-grade fold and thrust belt affecting a Palaeozoic (meta)sedimentary succession ranging in age from Cambrian to Lower Carboniferous and including Ordovician metavolcanics (Spalletta and Vai 1989; Gattiglio

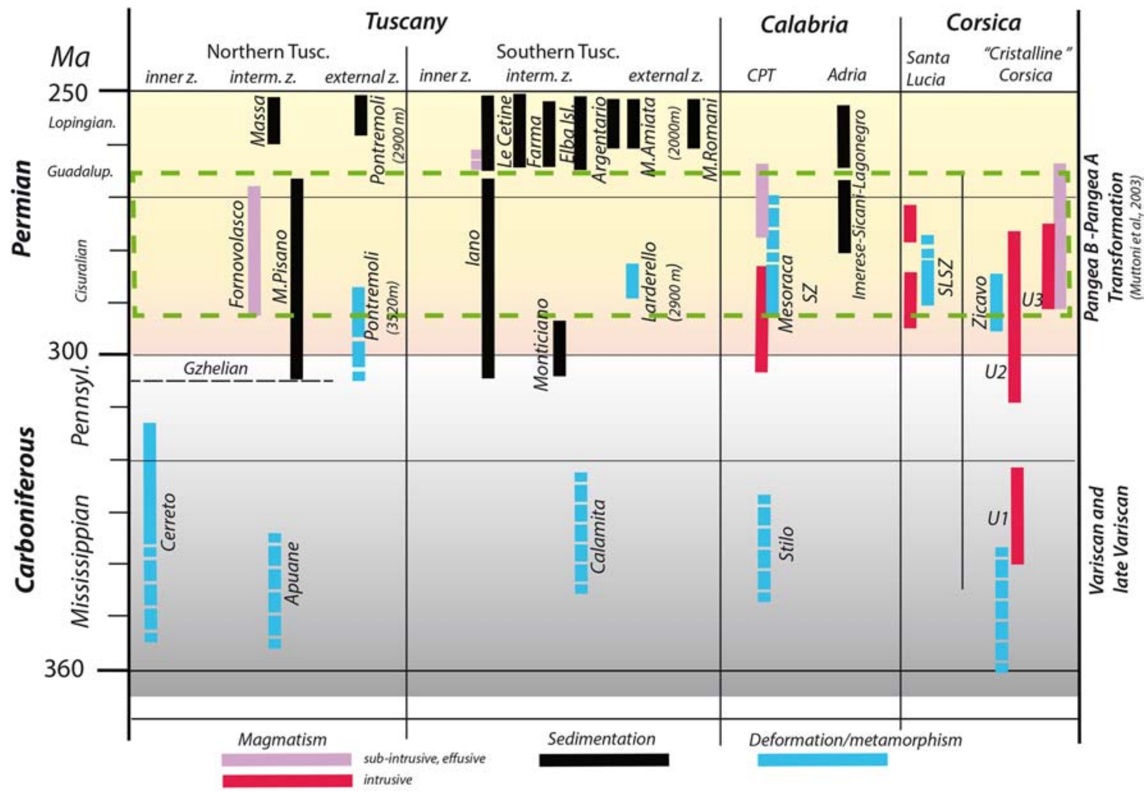


Fig. 13 A summary of the age data of sedimentary, magmatic and tectonic events in late Palaeozoic record in Tuscany, Calabria and Corsica (Santa Lucia and Crystalline Corsica). Red bars refer to ages in intrusive mainly granitic rocks; whereas in pink subintrusive to effusive magmatic rocks, bluish bars refer to deformation and metamorphic ages, and black lines denotes the range of sedimentation as deduced from paleontological data. Sources of data are reported in text and references

et al. 1989; Vai 2001), as a key character for its location. Locally, the low-grade upper crustal units are observed as tectonically overlying medium-grade paragneiss (i.e. Mammola paragneiss) to be considered "allochthonous tectonic window", with an overall tectonic grain, therefore supporting a correlation with the nappe zone of the Sardinian Variscan orogen (Gattiglio et al. 1989; Vai 2001).

Finally, if "Autochtonous" Corsica was an integral part of the Corsica-Sardinia Variscan orogen, the basement of Santa Lucia Nappe can be considered (similarly to the remnants of the Ersa-Centuri unit within the Corsican Schistes Lustres) as part of a further eastern (in present coordinates) crustal domain. Whether, after the Mesozoic rifting, the Santa Lucia domain was still attached to the "Autochothonous" (Durand Delga 1984; Rossi et al. 2006; Li et al. 2015) or part of an OCT or AlKaPeCa-type microblock (Lahondere 1996; Michard et al. 1992; Molli and Malavieille 2011; Lin et al. 2018), we suggest that a regional-scale transtensional fault, here called Santa Lucia Fault, existed during Permian, east (in present coordinates) of Corsica-Sardinia.

6.3 The late Palaeozoic tectonics in Corsica-Sardinia, Calabria and Tuscany: a strain partitioned regional frame and two previously overlooked regional faults

The crustal domains investigated here, show evidence of late Palaeozoic tectonics and magmatism that have been recorded differently in their lower to upper crustal levels (Fig. 13). Key features to be highlighted are the remnants of two regional-scale structures bounding, to the west and the east, the pre-Mesozoic Calabria and Tuscan crustal domains. The south-eastward structure (East Tuscan Fault), restoring the Apenninic displacements (Figs. 5c, 12a), has an original paleotectonic position between the external Tuscan and the Umbria-Marche domains as defined in the Mesozoic paleotectonics (Figs. 2c, 6c and 12a) within the Adria Plate to come. This fault, whose exhumed mid-crustal remnants can be found in subsurface from Pontremoli to Larderello-Travale (Figs. 2, 12a, b), may be further prolonged south of Tuscany (Figs. 6c, d) into the early to late Permian rifted crustal domain hosting basins and carbonatic platforms of the Lagonegro-Imerese-Sicani in the Southern Apennines and Sicily (Catalano et al. 1995; Mazzoli et al. 2001; Patacca

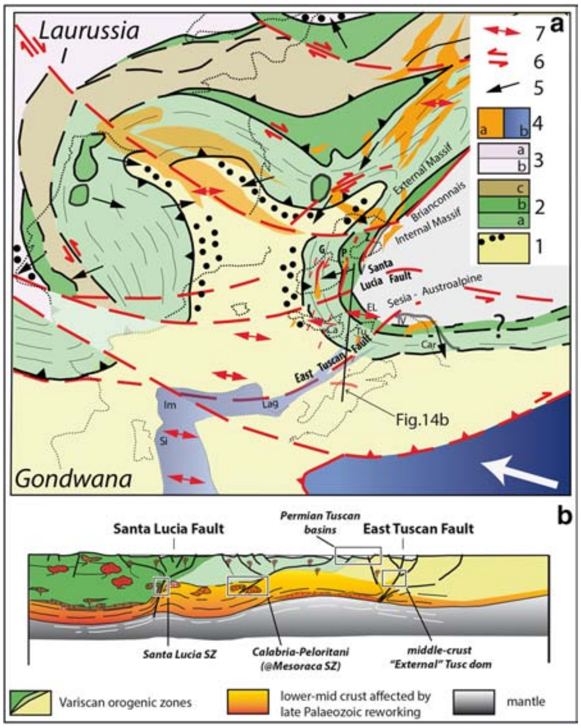


Fig. 14 Interpretative configuration of the post-Variscan setting (at c. 270 Ma) based on a modified version of Fig. 2 in Matte (2001). The scheme show a possible frame of the Variscan belt of western Europe and north Africa, with the main Upper Carboniferous-Permian regional fault systems and related continental basins (after Burg et al. 1994). The original scheme of Matte (2001) is modified taking into account the data and interpretations proposed in this work. The Santa Lucia Fault (whose remnants are documented in the lower to mid-crust shear zone in the Alpine Santa Lucia Nappe), and the East Tuscan Fault (remnants in subsurface of Pontremoli and Larderello deep well) in the Apennines are represented. The East Tuscan Fault is prolonged southwards into the Lagonegro-Imerese-Sicani marine rifted domains (Catalano et al. 1995) westernmost extension of the Permian rifted basin related with Neo-Tethys (Ziegler and Stampfli 2001; Stampfli et al. 2002; Garfunkel 2004; Xyapolis et al. 2006; Schettino and Turco 2011). (1) Gondwana and Gondwana-derived crust blocks (Apulia and Adria figured); (2) Southern Europe Variscan belt: (a) low-grade external and nappe zone dots foreland domains, (b) axial zone medium to high grade units, suture/s and Ordovician arcs, (c) hinterland and Variscan retrowedge (Armorica, Hun or Brunia terranes); (3) (a) Laurussia-derived blocks of the northern continent; (b) inner domains of the South Europe Variscan orogen; (4) Permian sedimentary basins: (a) continental, (b) marine; (5); main vergence of the nappes; (6) kinematics of late Palaeozoic regional faults; (7) extension direction during Upper Carboniferous-Permian. The possible positions of the Briançonnais, South-Alpine and Austro-Alpine domains are also reported (see also Festa et al. 2018; Ballevre et al. 2018 and references); thin dashed red lines represent dyke trend in Calabria (after Festa et al. 2010). Si: Sicani; Im: Imerese; Lag: Lagonegro; Ca: Calabria; Tu: Tuscany; Car: Carnia; IV Ivrea Zone; EL: External Ligurian; Z: Zicavo Shear Zone; P: Posada-Asinara; G: Grighini Shear Zone. (b) Conceptual schematic cross-section to envisiged the proposed tectonic scenario for late Palaeozoic in Central Mediterranean. The regional cross-section is traced not-perpendicular to the main regional fault systems to show and insert all the late Palaeozoic crustal domains and the different zones of South Europe Variscan orogen discussed in the text

and Scandone 2007). For the same regional structure, its northward prolongation may be found in the shallow-crustal splays defining the extensional/transtensional pull-aparts of Forni, Pramollo and Tarvisio in the external Carnian Alps (Massari 1986; Venturini 1990; Cassinis et al. 2012). Following this reconstruction (Fig. 14), a minimum estimated lenght for the East Tuscan Fault system may be regarded as c. 1000 km.

It is to be noticed that our proposal follows some precursory suggestions by Rau and Tongiorgi (1981), Rau (1990, 1991, 1994) who postulated the presence of a major crustal discontinuity to separate the "pre-Mesozoic" Tuscan Domain from that of Umbria-Marche, with a southernmost prolongation into the Lagonegro-Sicani-Imerese (as also suggested by Scandone 1975). Moreover, Deroin and Bonin (2003), Ziegler and Stampfli (2001), von Raumer et al. (2013) recently located a major late Permian structure in the same position, i.e. within the becoming Adria plate.

A second regional scale fault, the Santa Lucia Fault, characterized by a similar length and a field-constrained sinistral kinematics, could instead be located east of Corsica-Sardinia in a significant position to become a weak domain with a prolonged structural heritage during Mesozoic and Tertiary evolution. Also in this case, our proposal follows some previous suggestions, for instance that of Bard (1997), who however hypothesizes a dextral kinematics, or of Deroin and Bonin (2003) and Stampfli et al. (2002).

Therefore, the regional scale tectonic scenario that we propose (Fig. 14a, b) includes two major transcurrent/transtensive fault systems, the Santa Lucia Fault and the East Tuscan Fault, bounding the former crustal domains of Calabria and Tuscany, where extensional/transtensional structures were formed in the deep, intermediate and shallower portions of their crust (Fig. 14b).

Figure 13 shows the intimate relationships that link the period of magmatism (intrusive and effusive), the age of high-temperature and low-pressure metamorphism in the lower crust (Calabria and Santa Lucia in Corsica), the ages and kinematics of lower to mid crustal shear zones (Mesoraca in Calabria, Santa Lucia in Corsica and the East Tuscan Fault), the activity of transtensional shallow crustal splays where late Palaeozoic Tuscan pull-apart basins developed (Rau 1994; Spina et al. 2019), inferring their marine communication with the southernmost rifted domains of the Imerese-Sicani-Lagonegro in which deep basins and platforms developed, in turn connected with Neo-Tethys paleodomains (Ziegler and Stampfli 2001; Vai 2001). All these data point out the interconnection between the Tuscany-Calabria and Corsica crustal domains in a regional-scale strain–partitioned geodynamic setting controlled by a first-order transcurrent/

transtensive fault network which provided pathways for the intrusion and melt migration within it and/or outward in the intervening crustal domains where extensional to transtensional deformation developed (see also Rossi et al. 2015).

In this respect, it is worth attempting a comparative analysis of the late Carboniferous—early Permian tectono-magmatic evolution of Corsica and Calabria lower-crustal segments. Geochronological data indicate a substantial synchronism between magmatism and metamorphism both in Corsica (Rossi et al. 2009, 2015; Zibra et al. 2010) and in Calabria (Grässner et al. 2000; Caggianelli et al. 2013) in the time span of c. 310–280 Ma. The *P*–*T* path outlined for Corsica in Fig. 10f, presents analogies and some differences with respect to the *P*–*T* paths to that of the lower crust exposed in Calabria (Schenk 1981, 1989; Grässner et al. 2000; Caggianelli et al. 2013). In both regions, peak pressure occurred at 0.6–0.8 GPa and was followed by a decompression event with temperatures remaining close to 800 °C. The extent of decompression is slightly lower in Calabria (up to 0.2 GPa) than in Corsica (in the order of 0.25–0.3 GPa). Afterwards, the Permian cooling event was accompanied by moderate decompression in Corsica and was near-isobaric in Calabria. Thus, the resulting cooling path crossed the andalusite P–T field in Corsica and, limited to the top levels of the lower crust, in Calabria, in agreement to the major component of transtension recorded in Santa Lucia basement of Corsica compared to those in Calabria lower crust as suggested in our proposed tectonic model.

6.4 Late Palaeozoic tectonics in Central Mediterranean: between Variscan orogen and Tethyan rifting

The meaning and the geodynamic processes related to the late- and post-Variscan tectonics in Central Mediterranean, i.e. the Corsica-Sardinia, Tuscany, Calabria and Alpine domains, have been widely debated similarly to the other segments of the Variscan belt in Europe (Echtler and Malavieille 1990; Malavieille et al. 1990; Van den Driessche, Brun 1992; Burg et al. 1994; Franke et al. 2011; McCann et al. 2006; Roger et al. 2015). Some models, especially those focused on Corsica-Sardinia, try to tightly connect structures and upper Carboniferous and Permian tectono-magmatic processes to the latest stages of the Variscan orogen considering them as its waning shortening events, with or without an oblique convergence-setting (Elter et al. 1990; Conti et al. 2001; Carosi and Oggiano 2002; Corsini and Rolland 2009; Padovano et al. 2012); or to a post-collisional transpression of an Himalayan-type collisional belt in its easternmost marginal side (Carosi and Palmieri 2002; Iacopini et al. 2008; Frassi et al. 2009; Rolland et al. 2009; Padovano et al. 2012). Alternatively, authors have invoked gravitational collapse (Carmignani et al. 1994; Cappelli et al. 1992) or late- to post-orogenic extension (Thevoux-Chabuel et al. 1995; Renna et al. 2006; Giacomini et al. 2008; Rossi et al. 2015).

Moreover, for the pre-Mesozoic Alpine domains some authors have recently suggested a long-lasting period of active extension which started with the unroofing of the inner Variscan belts, followed by the Permo-Triassic thermal perturbation, to end with crustal break-up and the formation of the Alpine Tethys Ocean (Spalla et al. 2014). According to this view, a kinematic and thermo-mechanical link between the latest stages of Variscan orogenic construction and the beginning of the Alpine Tethys rifting has been suggested (e.g. Marotta and Spalla 2007; Roda et al. 2018; Festa et al. 2018).

All above mentioned, interpretation-types have been similarly proposed for the tectono-magmatic shaping and evolution of the Calabria-Peloritani Terrane (Angì et al. 2010; Fornelli et al. 2011; Liberi et al. 2011; Laurita et al. 2014).

Within our regional conceptual scheme, some of these proposals may be checked, discussed, and eventually ruled out. For instance, Fig. 14 clearly shows that the regional-scale faults developed across the former South Variscan belt, not only reactivating and reworking internal or axial domains (e.g. NE Sardina, SW Corsica; inner South Alpine) but also developing in the Variscan foreland, in the external domains as well as in the "stable" Gondwana. This evidently implies that late Palaeozoic tectonics and related magmatic processes cannot be considered as "Variscan orogen-related", i.e. related to the waning stages of syn- or late- convergence shortening, post-collisional extension or gravitational collapse. On the other hand, Fig. 14 quite evidently shows that late Palaeozoic structures were part of a post-Variscan setting characterized by an independent tectonics, kinematically and dynamically unrelated (Burg et al. 1994) with that in which building and collapse of the Variscan orogen occurred. This has been already remarked, at least locally for instance in the South Alpine basement by Handy and Zingg (1991); Schmid (1993); Handy et al. (1999); Handy et al. (2001); Pohl et al. (2018) as well as in the external domains of the central Variscan belt, e.g. in the Montagne Noire (Etchler and Malavieille 1990; Franke et al. 2011; Roger et al. 2015) and in other segments far from to the axial zone of the Variscan orogen (Burg et al. 1994; Ziegler and Stampfli 2001).

The post-Varsican regime, following the early suggestions by Arthaud and Matte (1975), Bard (1997), Burg et al. (1994), Matte (2001), Muttoni et al. (2003), Martinez-Catalan (2011), Ballevre et al. (2018) and references therein, may be connected with a network of crustal scale strike-slip regional faults which accommodated mainly transtensional (and locally transpressional) deformation, associated and coeval with the overall clockwise rotation of Gondwana with respect to Laurussia, with Gondwana-derived blocks and elements of the southern segment of the Variscan belt escaping toward east in the wake of subduction of the eastern ocean (Fig. 14). Our scheme in Fig. 14a shows quite clearly that for their locations some of the late Palaeozoic structures acted as zones of weakness and were only locally reactivated or reused during Mesozoic rifting (mid-Triassic and later Triassic-Liassic), during the Cretaceous to Tertiary convergence and/or during the Tertiary opening of Central Mediterranean extensional basins.

7 Conclusion

As a result of the proposed review of data derived from the research of our group over the past decades integrated with those of literature, the frame of a classical paleotectonic configuration of the Variscan belt of western Europe and northern Africa in the late Palaeozoic may be completed to include the crustal domains of Calabria and Tuscany, and by matching them with the structural zonation of the Variscan orogenic frame of Sardinia-Corsica. By doing that, some major large-scale tectonic features may be outlined (Fig. 14):

1. The presence of domains affected by two regional-scale faults whose relict remnants may be found in the crustal record of the Permian shear zones, i.e. the sinistral transtensive Santa Lucia, on the one side, and the East Tuscan Fault, possibly (but not directly constrained) with sinistral kinematics too, on the other;
2. Between these inferred crustal scale structures, distributed deformation, late Carboniferous-Permian in age, may be related to the record of the middle and shallow crust in Calabria and Tuscany, where extensional to transtensional deformation has been connected to the tectono-magmatic and sedimentary history of these regions;
3. The absence of unique relationships between late Carboniferous to Permian tectono-magmatic and sedimentary structures of the studied areas and the axial or inner structural domains of the former South-Variscan belt undermines, at least for Calabria, the concept of late-orogenic collapse of the overthickened crust as instead documented in other Variscan segments of Europe;
4. The late Palaeozoic history of the Central Mediterranean region with its structural inheritance might play a major role in controlling the Mesozoic and Cretaceous to Tertiary paleogeographic and paleotectonic evolution of the region, with some local to regional relationships, in terms of localization in space and time.

Although our proposal is based on some well-constrained local field and subsurface data, and on some interpretations rooted in classical literature of the Italian peninsula, as well as on some largely accepted regional correlations, most of the continental crust with its record of late Palaeozoic history that is the focus of this paper is no longer accessible since it subducted during the Tertiary orogenic history of central Mediterranean or it is unaccessible under younger rocks on land or under the sea. Therefore, our interpretation features as a working hypothesis to be further constrained by new data and checked against independently derived kinematic models (e.g. Agostini et al. 2020; Angrand et al. 2020; Le Breton et al. 2020; van Hinsbergen et al. 2020).

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Author details

[1] Università di Pisa, Via S.Maria 53, 56126 Pisa, Italy. [2] Dipartimento di Scienze della Terra e Geoambientali, Università di Bari, Via Orabona 4, Bari, Italy. [3] Consiglio Nazionale delle Ricerche - IGG, Via Moruzzi 1, Pisa, Italy. [4] Dipartimento di Scienze della Terra e Geoambientali, Università di Firenze, Via La Pira 4, Firenze, Italy. [5] Dipartimento di Fisica e Geologia, Università di Perugia, Via Pascoli, Perugia, Italy. [6] Department of Mines, Industry, Regulation and Safety, Geological Survey of Western Australia, 100 Plain Street East, Perth, WA 6004, Australia.

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