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The carbonate tectonic units of northern Calabria (Italy): a record of Apulian palaeomargin evolution and Miocene convergence, continental crust subduction, and exhumation of HP–LT rocks

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Abstract: In northern Calabria (Italy), the metasedimentary succession of the Lungro–Verbicaro tectonic unit preserves mineral assemblages suggesting underthrusting to depths in excess of 40 km. Internal deformation of these rocks occurred continuously during the following decompression. Index mineral composition associated with progressively younger tectonic fabrics indicates that a substantial part of the structural evolution took place within the blueschist-facies P–T field. Despite their tectonic and metamorphic history, the rocks of the Lungro–Verbicaro Unit preserve significant sedimentary and palaeontological features allowing correlations with successions included in adjacent thrust sheets and the reconstruction of the Mesozoic continental margin architecture. The subduction–exhumation cycle recorded by the Lungro–Verbicaro Unit is entirely of Miocene age. This portion of the Apulia continental palaeomargin was involved in convergence-related deformation not earlier than the Aquitanian. The integration of our results with available constraints on the tectonic evolution of the Apennine–Calabrian Arc system suggests that subduction and most of the subsequent exhumation of the Lungro–Verbicaro Unit occurred, up to Langhian time, at maximum vertical rates in excess of 15 mm a^{-1}. The exhumation process was then completed, at much slower rates (<2 mm a^{-1}) in Late Miocene time, as indicated by both apatite fission-track data and stratigraphic information.

The process of continental collision is commonly envisaged to result from the convergence and final docking of two continental margins, generally following the subduction-related consumption of an intervening oceanic domain. However, it is now well established that the arrival of a continental margin at a subduction zone does not necessitate locking of the trench and collision-related mountain building. A variable amount of continental crust may be subducted (A-subduction), as confirmed by the focal mechanism solution of recent earthquakes in some orogenic regions and by the occurrence of rootless nappes and of high-pressure (HP) and ultrahigh-pressure (UHP) metamorphic minerals in continental palaeomargin successions (e.g. Chopin 1984; Compagnoni *et al*. 1995). Numerical and analogue modelling studies confirm the feasibility of this process (e.g. Chemenda *et al*. 1996; Ranalli *et al*. 2000). In particular, continental subduction is reported to have occurred everywhere in the peri-Mediterranean orogens. Important constraints on the modes of deformation in rocks involved in these processes may be obtained from well-preserved, dominantly carbonate, exhumed sedimentary successions exposed in Mediterranean regions such as Crete (e.g. Stöckhert 2002). In these instances, the fundamental role of palaeomargin architecture in the process of continental subduction may be investigated (e.g. Thomson *et al*. 1999).

This paper presents stratigraphic, structural, petrological and thermochronological data on northern Calabria, in southern Italy. Here, a comprehensive picture of continental margin evolution may be obtained, from Mesozoic rifting and passive margin development, to plate convergence and recent (Miocene-age) continental subduction and subsequent exhumation of HP–LT metasediments. This case history confirms that strongly heterogeneous deformation and strain localization may be associated with relatively good preservation of primary features in dominantly carbonate successions involved in underthrusting to great depths.

Geological setting

The Apennine Chain forms the backbone of peninsular Italy (Fig. 1) and constitutes part of the Alpine mountain system in the Mediterranean region. The latter evolved within the framework of the convergent motion between the African and European plates and related microplates since Late Cretaceous time (e.g. Mazzoli & Helman 1994, and references therein).

The Apennine fold and thrust belt–foredeep system records east-directed thrust transport directions and the development and deformation of progressively younger turbiditic deposits to the east. These features indicate tectonic accretion atop a westward-dipping subduction slab since at least Oligo-Miocene time. As a result, the back-arc nature of the Western Mediterranean basins

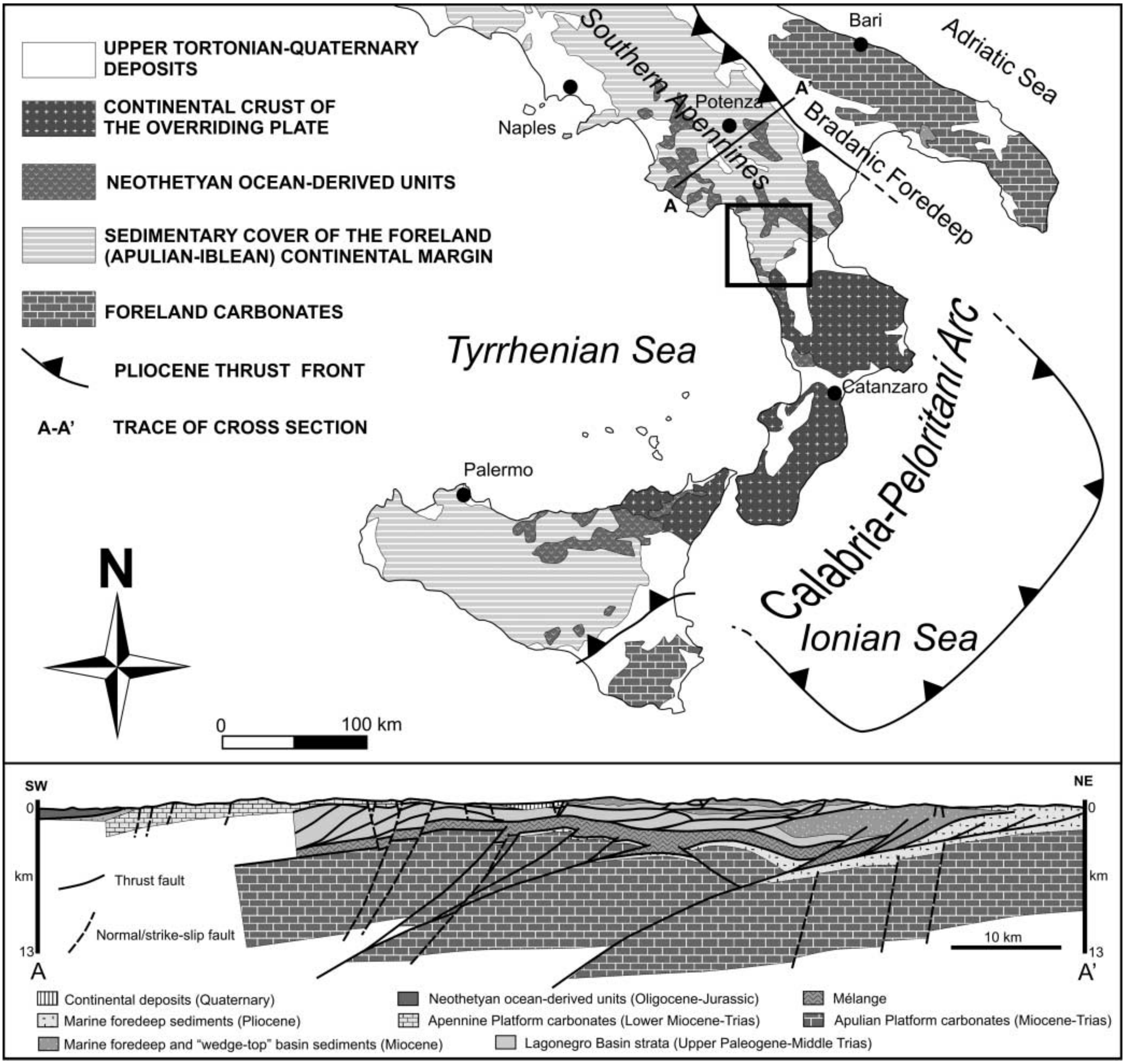

Fig. 1. Geodynamic and regional geological framework of southern Italy. The geological section is modified after Mazzoli *et al*. (2000).

linked to the eastward rollback of the Adriatic slab is now generally accepted (Royden 1993; Faccenna *et al*. 2001). The early configuration of the subduction system (Late Cretaceous–Oligocene interval) is still a matter of debate, particularly for the segment of chain between the southern Apennines and the Sicilian Maghrebids, commonly known as Calabria–Peloritani Arc, where crystalline basement and ophiolitic nappes crop out (Fig. 1). It is unclear whether the Calabria block formed part of the European continental palaeomargin (e.g. Boullin *et al*. 1986; Cello *et al*. 1996; Jolivet & Faccenna 2000; Rossetti *et al*. 2001) or resulted at least in part from earlier periods of northwestward accretion and/or microplate collision and amalgamation (e.g. Amodio Morelli *et al*. 1976; Grandjacquet & Mascle 1978; Bonardi *et al*. 1982, 2001). On the other hand, there is general consensus that since Early Miocene time convergence occurred between the Corsica–Sardinia–Calabria block, to the west (present-day coordinates), and the Apulia (or Adria) block, of African affinity, to the east.

The occurrence of a subducted slab beneath southern Calabria is inferred from a well-imaged positive anomaly of seismic-wave velocities identified by seismic tomography beneath the southern Tyrrhenian Sea. The high-velocity body, continuous from the surface and dipping towards the central Tyrrhenian Sea, flattens down onto the lower mantle discontinuity, with a total length of about 1200 km. It is accompanied by a well-developed seismicity to a depth of at least 400 km (Lucente *et al*. 1999). The distribution of this seismicity follows the upper surface of the tomographically modelled slab and indicates the presence of a

Wadati–Benioff surface (Anderson & Jackson 1987). Subduction beneath the Calabrian Arc is also accompanied by late Miocene–Quaternary back-arc extension and sea-floor spreading in the southern Tyrrhenian Sea (Sartori 2001, and references therein). In the northern part of this region, three main groups of nappes may be distinguished (Fig. 1): (1) crystalline basement nappes; (2) ophiolite-bearing nappes; (3) mostly calcareous nappes derived from the inner part of the Apulia continental margin. Group (1) includes Variscan metamorphic rocks and granitoids, and Mesozoic low-grade metasedimentary cover of the Bagni Units and sedimentary cover of the Sila Unit. These overlie ophiolite-bearing units (group (2) above) ranging from greenschist-facies (Malvito Unit), showing in some cases (Reventino–Gimigliano Unit) evidence of an earlier HP–LT metamorphic event, to blueschist-facies metamorphism (Diamante–Terranova Unit). Some of these units, the HP–VLT (very low temperature) Frido Unit and the unmetamorphosed North Calabrian Unit (Bonardi *et al*. 1988), crop out NE of the Pollino Massif, in the region described in the literature as the Calabria–Lucania border area. This metasedimentary or sedimentary cover includes the uppermost Oligocene and Aquitanian units, respectively (Bonardi *et al*. 1993; Di Staso & Giardino 2002). A mélange unit, including huge blocks of serpentinites and continental crust rocks, is locally interposed between them.

The ophiolite-bearing units tectonically overlie tectonic units of the southern Apennines fold and thrust belt, derived from Miocene to early Pleistocene deformation of the sedimentary cover of the Apulia passive margin. These tectonic units (group (3) above) are dominated by Mesozoic–Neogene sedimentary successions comprising kilometre-thick, shallow-water to slope carbonates and pelagic basin cherts, limestones and pelites, as well as a complex pattern of Neogene–Quaternary siliciclastic strata deposited in thrust-top and foredeep basin environments. The shallow part of the southern Apennines forms a displaced allochthon that has been carried onto a footwall of foreland strata essentially continuous with the foreland Apulian platform (e.g. Cello & Mazzoli 1999; Butler *et al*. 2004; Fig. 1). The Apulian carbonates (6–8 km thick) buried beneath the allochthon are deformed by relatively low-displacement, high-angle reverse faults involving the basement (Butler *et al*. 2004; Shiner *et al*. 2004). Therefore, a switch from thin-skinned to thick-skinned thrusting appears to have occurred in the southern Apennines as the Apulian carbonates, and the underlying thick continental lithosphere, were deformed (Butler *et al*. 2004). This process should have been preceded by a stage of continental margin subduction associated with the earlier, extensive detachment and accretion of continental sedimentary cover material. However, most of the Apulia margin derived units forming the Apennine belt have been deformed at P–T conditions never exceeding those of burial diagenesis (Corrado *et al*. 2005). Greenschist-facies metamorphism has been reported in Middle–Upper Triassic marbles and phyllites of northern Calabria since the end of the nineteenth century, as well as in Lower Miocene metapelites (Damiani 1970) exposed in the area of the present study. The Triassic phyllites were interpreted by the majority of researchers as a lower part of the sedimentary cover of the Apulia plate. Recently, Rossetti *et al*. (2004) reported HP–LT mineral assemblages from Triassic phyllites and proposed that they are dismembered blocks within a tectonic mélange and mark the remnants of an ophiolite-bearing accretionary wedge. However, Iannace *et al*. (2005*b*) proposed that the bulk architecture of a subducted and partly exhumed continental margin could be reconstructed in this area. A good preservation of coherent continental margin successions is supported by a large amount of detailed multidisciplinary data, as illustrated in the following sections.

Stratigraphy

The Meso-Cenozoic sedimentary–metasedimentary successions analysed in this study (Fig. 2), can be grouped into three tectonic units: the Lungro–Verbicaro Unit, the Pollino–Ciagola Unit and the Cetraro Unit. The Pollino–Ciagola Unit is tectonically overlain by ophiolitic units in the northeastern part of the study area (Fig. 2), whereas toward the NW it is overlain by the Lungro–Verbicaro Unit. The Cetraro Unit is tectonically overlain by the Lungro–Verbicaro Unit and both are tectonically overlain by ophiolitic and continental crust-derived units (Sila and Bagni Units). The tectonic contacts between the latter units are sealed by Tortonian–Messinian coarse clastic deposits and by Pliocene and Pleistocene marine and alluvial sediments.

The Lungro Verbicaro Unit

The stratigraphic succession of the Lungro–Verbicaro Unit affected by HP–LT metamorphism consists in its lowermost part of Middle Triassic phyllites and metarenites with carbonate intercalations (Fig. 3). The latter contain locally rich assemblages of strongly recrystallized dasycladacean algae of Anisian and Early Ladinian age (Bousquet *et al*. 1978), suggesting a shallow-water origin. These deposits are followed by Ladinian–Carnian metalimestones, marly metalimestones and dolomites, with local build-ups rich in dasycladales, crinoids and spongiomorphids and a well-developed reef complex in the area of Monte Caramolo (Iannace *et al*. 1995). One of the most characteristic facies is represented by dark marly metalimestones with abundant *Thalassinoides*-type burrows and thin bivalve- and gastropod-rich beds. Pervasive dynamic recrystallization and local isoclinal folding hinder a detailed facies analysis. However, analogy with similar successions occurring in Spain (Iannace *et al*. 2005*a*) suggests deposition in a wide epicontinental shelf with sedimentation controlled by the interplay between carbonate production and siliciclastic input, such a shelf being bordered by sponge and cement-rich buildups.

In the Carnian layers there is a significant increase of siliciclastic beds intercalated with metadolomites, metalimestones and evaporites. Siliciclastic and evaporite strata prevail to the SW, whereas to the NE the succession is dominated by metadolomites with still recognizable tempestites, breccia beds and neptunian dykes. This facies distribution is interpreted as a transition from a coastal, partly evaporitic environment toward an open shelf and a margin affected by synsedimentary tectonics (Iannace *et al*. 2005*a*). This succession is followed by several hundred metres of metadolomites referable to the Norian–Rhaetian, showing sudden lateral facies changes. Within a few kilometres, inner platform facies pass to marginal build-ups, consisting of an unusual assemblage dominated by microbes and serpulids and small calcareous sponges, and then to slope and restricted basin facies (Climaco *et al*. 1997; Zamparelli *et al*. 1999; Perri *et al*. 2003).

The Jurassic is mostly represented by a cherty metalimestone succession (Calcari con Selce Fm) of variable thickness, which can be found either in stratigraphic continuity with the metadolomites or locally above an angular unconformity. The metalimestones are generally coarse and crystalline; however, it is possible to recognize abundant beds of calcareous turbidites. The total thickness of this formation varies from 70 to 200 m, and it is generally thicker to the NE. On top of this lithological interval,

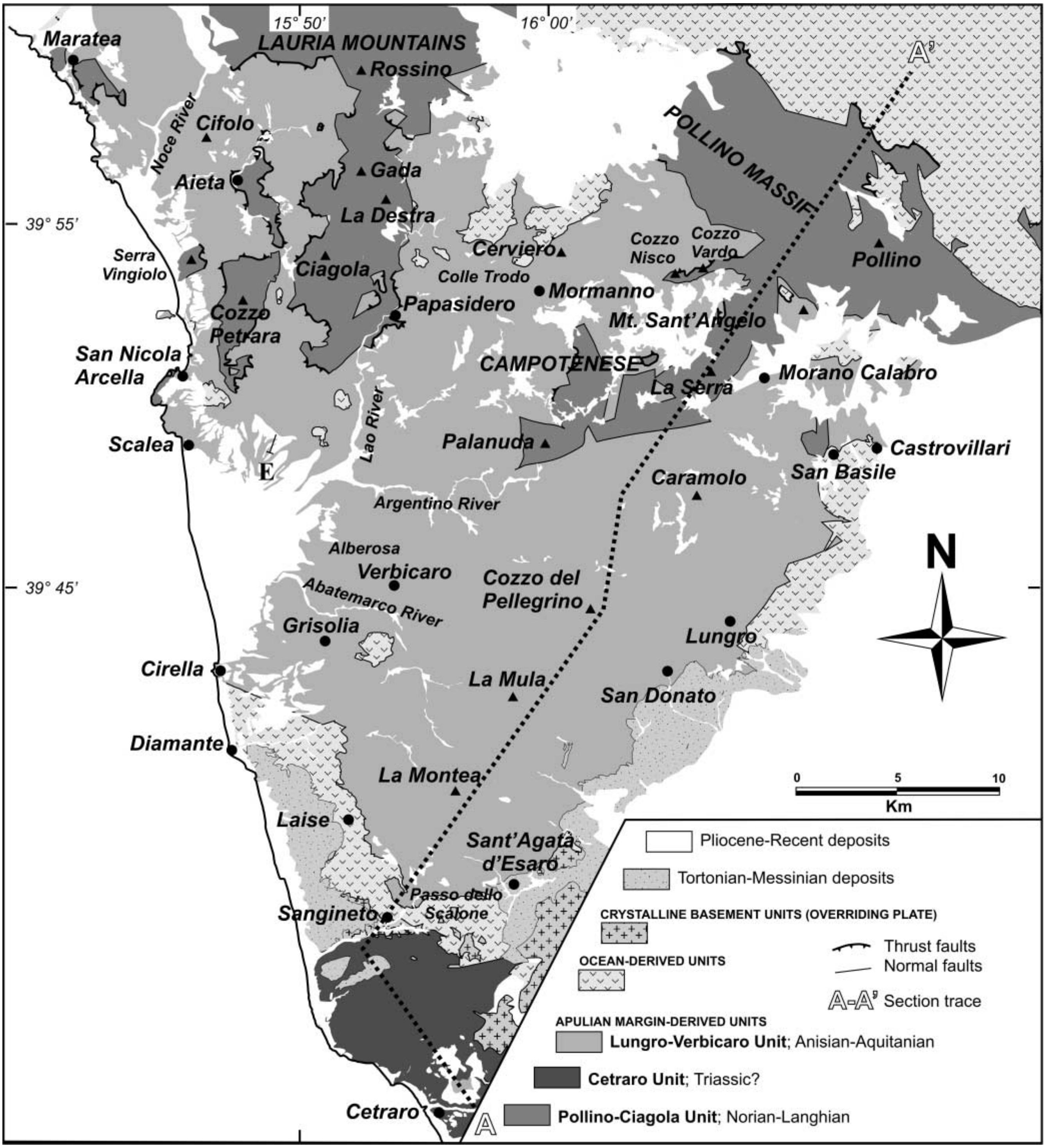

Fig. 2. Tectonic sketch map of the study area, showing location of cross-section of Figure 3.

dated by Damiani (1970) to the Early Lias–Late Dogger, are locally preserved (Monte Cerviero, Verbicaro, Cirella) red siliceous slates and radiolarite beds, of probable Dogger age (O' Dogherthy, pers. comm.). The overlying Colle Trodo Fm (Selli 1957), rests disconformably on the older formations. Its base consists of coarse carbonate conglomerates (Fig. 4a; the Brèche à silex of Grandjacquet & Grandjacquet (1962) or Brecce Poligeniche of Damiani (1970)) of Maastrichtian–Palaeocene age. The breccia clasts consist mainly of platform limestone but locally fragments of chert are also present (Fig. 4b). The size of the carbonate clasts decreases from the NE (diameters of 15–80 cm, 30 cm on average) to the SW (2 cm on average). These rudites are followed abruptly by red and green metapelites with frequent beds of calcareous turbidites grading upward to

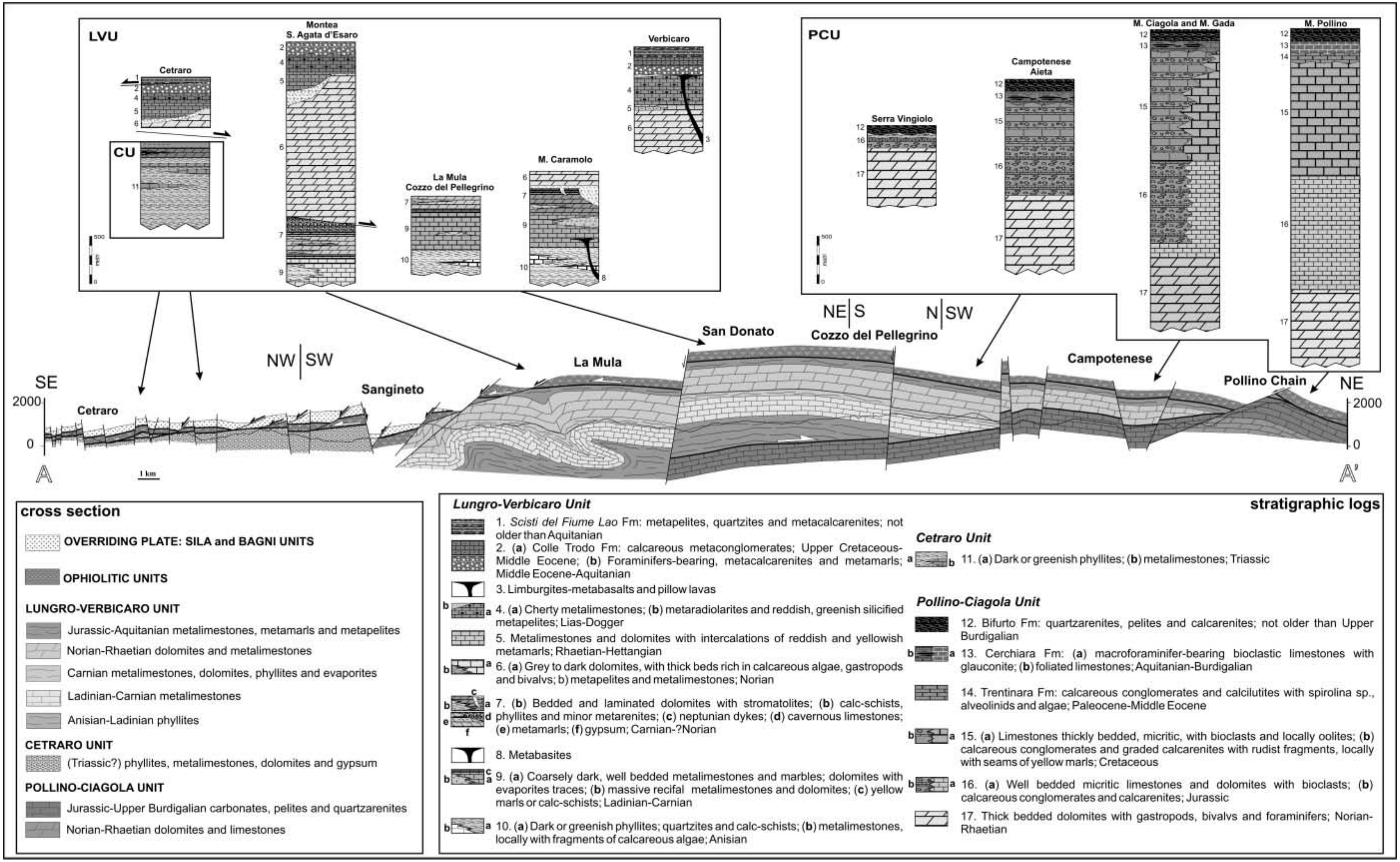

Fig. 3. Representative geological section (location shown in Fig. 2) and stratigraphic columns.

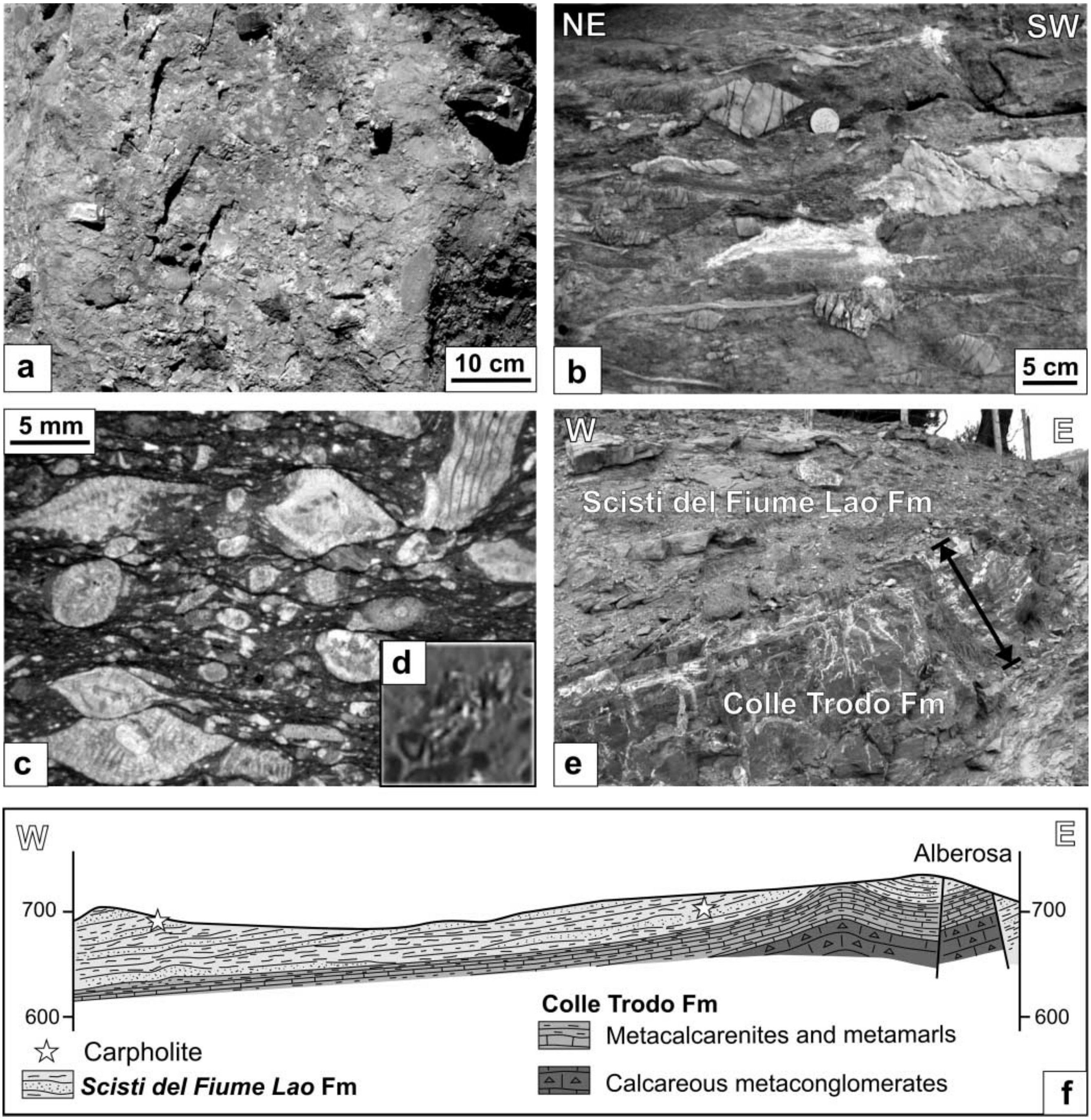

Fig. 4. (**a**) Calcareous breccia (including chert clasts) at the base of the Colle Trodo Fm (Alberosa area). (**b**) Highly strained calcareous breccia of the Colle Trodo Fm (Cirella area). (**c**) Foraminiferal bioclastic limestone of the upper part of Colle Trodo Fm (Maierà area). (**d**) *Helicosphaera carteri* (nannoplankton) from Miocene metapelites (Scisti del Fiume Lao Fm, Alberosa area). (**e**) Stratigraphic contact between Colle Trodo and Scisti del Fiume Lao Fms (Alberosa area). (**f**) Geological section across the Alberosa area. In this area, the stratigraphic relationships between the youngest formations of the Lungro–Verbicaro Unit are well preserved. Stars indicate localities of carpholite findings.

yellowish metamarls. The abundant macroforaminifera characterizing the latter interval (Fig. 4c) indicate an Eocene to early Miocene age (Grandjacquet & Grandjacquet 1962). A refined Middle Eocene to Aquitanian age was obtained by nannoplankton biostratigraphy (D'Errico 2004).

The Colle Trodo Fm grades upward to metapelites and metarenites containing some calcarenite and calcirudite beds with microforaminifera of early Miocene age (the Flysch del Lao of Damiani (1970) or Scisti del Fiume Lao of Burton (1971)). The stratigraphic contact with the underlying Colle Trodo Fm is preserved in only a few localities. The best exposure is north of the Verbicaro village (Fig. 4e and f) where the marker species *Helicosphaera carteri*, indicating an age not older than Aquitanian (Fig. 4d), has been found.

Basic rocks, represented by metabasaltic lavas with pillow structures and dykes ('limburgites') are also present, partially bearing a signature of HP–LT metamorphism (Pierattini *et al*. 1975). These rocks generally cut across the Triassic and Jurassic formations of the unit and are locally capped by the siliceous slates and radiolarites. Moreover, clasts of these magmatic rocks occur within the breccias of the Colle Trodo Fm. Therefore, although these lavas were considered as Tertiary in age by Pierattini *et al*. (1975), a Jurassic age appears to be most probable for this magmatism.

The Pollino–Ciagola Unit

The main outcrops of the Pollino–Ciagola Unit that have been studied are located on the Ciagola–Gada ridge, and around the villages of Papasidero, Aieta, Maratea and Campotenese, and comprise carbonate slope facies deposits. Compagnoni & Damiani (1971) pointed out the large and variable stratigraphic gaps occurring in these slope facies successions, which were considered as the lateral equivalent of the platform succession cropping out in the Pollino Massif. As a result of localized intense deformation (Iannace & Vitale 2004), some of these outcrops had been incorrectly ascribed to the metamorphic San Donato Unit (Amodio Morelli *et al*. 1976; Bigi *et al*. 1991).

The Late Triassic is almost everywhere represented by thick-bedded, white to light grey dolomites. Laminated, fenestral facies generally alternate with bivalve- and gastropod-rich beds, indicating sedimentation in peritidal, low-energy environments. Locally dasycladacean- and sponge-rich facies occur. The fossil content (gastropods, foraminifers and algae) is typical of the Norian–Rhaetian. In some localities (Serra Vingiolo, Monte La Serra) these dolomites grade upward to limestones with megalodontid shells, tens of centimetres in size, of latest Norian–Early Rhaetian age.

In the Jurassic, a marked facies differentiation becomes apparent between the eastern and western outcrop areas. In the former, calcareous wackestones and packstones with Foraminifera, algae, ooids and oncoids are dominant, indicating a persistent shallow-water, lagoonal environment. As such, these successions are comparable with those of the Monte Pollino area. In contrast, resedimented, calcareous facies are dominant at Monte La Serra, Monte Ciagola and Aieta, indicating a lateral transition to a slope domain. The most complete succession is that of Monte La Serra, where the upper Triassic megalodontid limestones are replaced upward by mud-supported calcareous conglomerates rich in coral, gastropods, sponges and echinid fragments. The Jurassic age is confirmed in the lower part of the interval by the presence of *Stylothalamia* sp. At Monte Ciagola the conglomerates contain a marly, yellowish matrix and are generally strained.

A comparable facies distribution is recorded also in Cretaceous rocks, with the eastern area dominated by shallow-water facies and the western area mainly occupied by resedimented slope deposits. The former are represented by dark, well-bedded, generally cyclic limestones with characteristic shelf macrofossils of this age (requienids, radiolitids, hippuritids) and a rich assemblage of microforaminifera and algae. These resedimented deposits comprise graded coarse calcarenites, floatstones and rudstones with abundant bioclasts of marginal environments (rudists, gastropods, corals, echinoderms). The resedimented facies are more abundant in the upper part of the succession. Yellow–greenish marly beds are intercalated with the conglomerates at Aieta and Monte Gada.

In the Pollino Massif (see Fig. 2) the Cretaceous shallow-water rudistid limestones are disconformably covered by the brackish lagoonal to shallow marine Trentinara Fm, followed by middle Aquitanian–lower Burdigalian open shelf calcarenites (Cerchiara Fm, Selli 1957) and by siliciclastic turbiditic deposits that are not older than Langhian, which include 'Numidian' quartzarenites (Bifurto Fm, Selli 1957; Patacca *et al.*, 1992). On the SE side of the same massif (Cozzo Vardo) the Cerchiara and Bifurto Fms directly overlie the Cretaceous limestones. In all other areas, instead, the Cretaceous shallow-water carbonates are followed by coarse calcareous breccias, which are generally highly strained (Iannace & Vitale 2004; Fig. 5a), resulting in a characteristic foliated lithofacies (calcaires plaquettés of the French workers). The occurrence of *Nummulites* sp., in both the clasts and the matrix (Fig. 5b and c), indicates that these beds are not older than Palaeogene. These breccias rapidly evolve to thin-bedded calcarenites that are also strongly deformed and contain, in the upper part of the succession, *Myogipsina* sp. and other Miocene Foraminifera. The succession is topped by marls, pelites and rare quartzarenite intercalations, which can be correlated with the Bifurto Fm.

The Cetraro Unit

In the southernmost part of the study area there are wide outcrops of phyllites, variably rich in quartz and/or calcite, locally with lenses of quarzites and quartz metaconglomerates. Metalimestones, metadolomites and gypsum also occur, mainly in the upper part of the succession. Metalimestones consist of fine-grained calcite with frequent dolomite porphyroclasts. At the meso-scale the metadolomite forms boudins, embedded in the metalimestones, varying in size from a few centimetres to tens of metres. Gypsum is present as fine-grained layers, frequently associated with vacuolar calcareous breccia. According to Amodio Morelli *et al*. (1976), the age of this succession is Anisian–Carnian by lithological correlation with the phyllites of the Lungro area, whereas according to Iannace *et al*. (2005*a*) the whole succession is a lateral equivalent of the Carnian part of the Lungro–Verbicaro Unit cropping out north of the Sangineto River. It should be emphasized that no direct evidence for the age of this unit is available.

Structure

The structure of the carbonate units exposed in the footwall to the ophiolitic sheets (see Fig. 2) is dominated by the tectonic superimposition of the Lungro–Verbicaro Unit on the Pollino–Ciagola and Cetraro Units. The low-angle tectonic contact between the Lungro–Verbicaro Unit and the Pollino–Ciagola Unit invariably juxtaposes older on younger rocks. The contact shows a general footwall flat thrust geometry, the rocks immediately in the footwall being the Palaeogene–Miocene stratigraphic top of the Pollino–Ciagola Unit succession. The thrust geometry in the hanging wall (Lungro–Verbicaro Unit) includes a frontal ramp, broadly trending WNW–ESE from San Nicola Arcella to San Basile (see Fig. 2). An upper detachment horizon occurs within the Lungro–Verbicaro Unit along evaporites and siliciclastic deposits characterizing the Carnian–Norian boundary. This is manifested by an extensive hanging-wall flat, which characterizes the area NNE of the San Nicola Arcella–San Basile alignment, where Upper Triassic dolomites of the Lungro–Verbicaro Unit overlie Miocene strata of the Pollino–Ciagola Unit (Fig. 5d).

At several sites (e.g. Monte Palanuda, Sassonia near San Basile), younger extensional faults offset the original contact

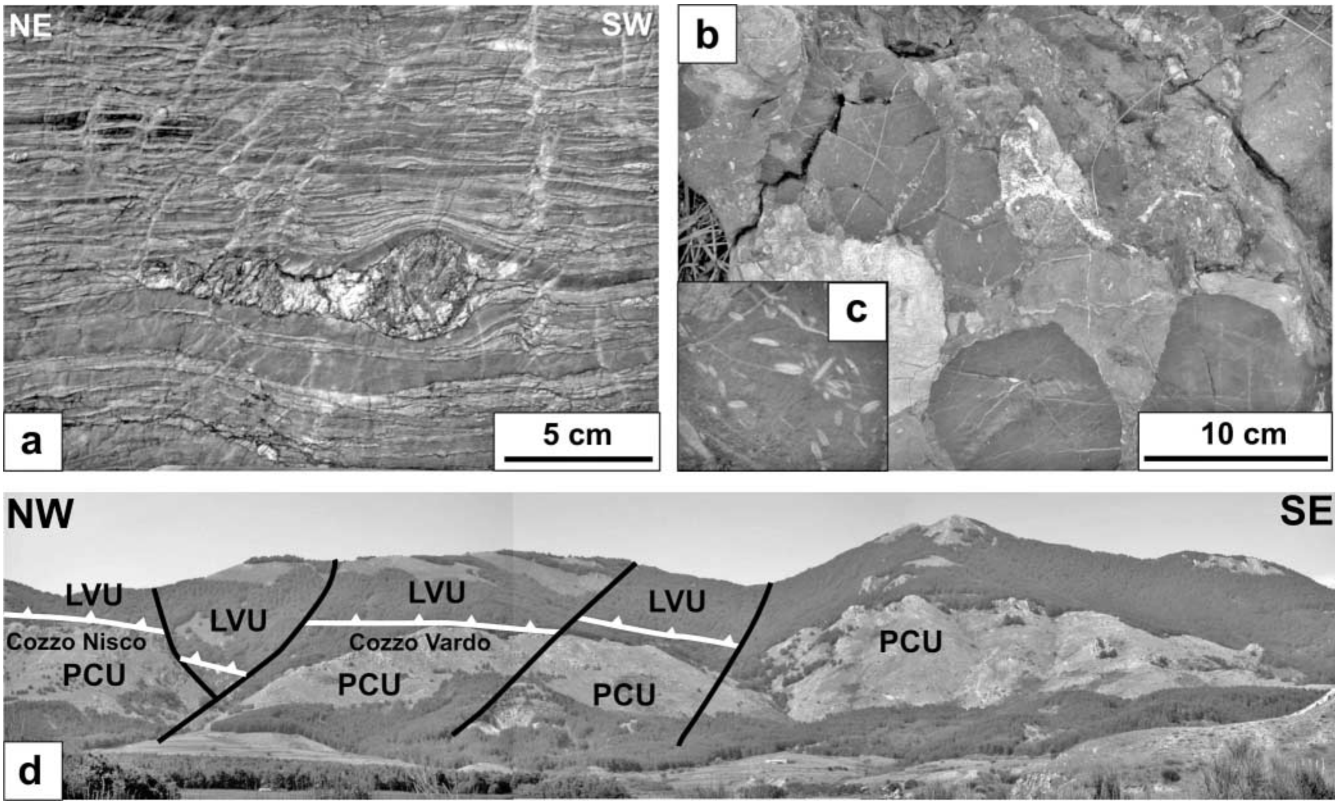

Fig. 5. (**a**) Intensely deformed calcareous conglomerate of the Pollino–Ciagola Unit in the Campotenese area. (**b**) Undeformed calcareous conglomerate of the Pollino–Ciagola Unit in the Campotenese area. (**c**) Well-preserved macroforaminifers (nummulites) in calcareous conglomerate of the Pollino–Ciagola Unit (Campotenese area). (**d**) Tectonic superimposition of the Lungro–Verbicaro Unit onto the Pollino–Ciagola Unit north of Campotenese. The original low-angle contact (white) is offset by late faults (black lines).

between the Lungro–Verbicaro Unit and the Pollino–Ciagola Unit, producing various tectonic juxtapositions of different parts of the stratigraphy of the two units. In particular, starting from the southeastern edge facing the Sibari Plain (San Basile to Sant'Agata d'Esaro villages) and from the Sangineto area, southward, the tectonic pile is affected by low-angle extensional faults, mainly south dipping. In the Passo dello Scalone and Sant'Agata d'Esaro areas, Rhaetian limestones directly overlie the Carnian part of the Lungro–Verbicaro Unit succession, with the interposition of a cataclastic breccia (Iannace *et al*. 2005*a*). In the San Basile and Timpone Sant'Angelo areas, the Norian dolomites of the Lungro–Verbicaro Unit overlie the Norian succession of the Pollino–Ciagola Unit by means of a low-angle tectonic contact, with the interposition of slices of Miocene pelites of the Pollino–Ciagola Unit. In the Cetraro area, the Norian–Miocene succession of the Lungro–Verbicaro Unit crops out discontinuously, and frequently the phyllites and metalime stones of the Cetraro Unit are directly overlain by the ophiolitic units. Upper Tortonian and Messinian deposits, unconformably overlying the Lungro–Verbicaro Unit, the Cetraro Unit, and the crystalline and ophiolitic nappes, are locally deformed along WNW–ESE-trending strike-slip faults and offset by younger, high-angle normal faults.

The Lungro–Verbicaro Unit, Cetraro Unit and Pollino–Ciagola Unit are all characterized by heterogeneous ductile deformation. In the Lungro–Verbicaro Unit, internal ductile strain is mainly localized at certain levels, particularly in Triassic phyllites and metalimestones, the Cretaceous–Middle Eocene calcareous conglomerates (Fig. 4a and b), and Miocene metapelites. Well-developed deformation features from these levels have been analysed in detail (see section below). However, the most spectacular structures in the Lungro–Verbicaro Unit consist of large-scale, SW-vergent backfolds involving most of the stratigraphic succession (Fig. 6). In the Cetraro Unit the ductile deformation affected phyllites and metalimestones, whereas brittle deformation affected dolomite levels, producing a pervasive boudinage at the micro- to macro-scale. In the Pollino–Ciagola Unit, deformation is localized mainly in Palaeogene–Miocene pelites, calcschists and conglomerates (see Fig. 5a and b), and subordinately in some levels within the Jurassic–Cretaceous succession. Examples of meso- and micro-structural features are shown in Figures 7–9, and orientation data are shown in Figure 10.

Lungro–Verbicaro Unit

D_1 structures. In Triassic phyllites, the oldest structures consist of a relict foliation $(S_1)_{LVU}$, locally preserved as a relict planar fabric within the microlithons of a later $(S_2)_{LVU}$ main foliation. Weakly zoned, millimetre-sized chloritoid prisms or rosettes (Fig. 7a) occur along $(S_1)_{LVU}$, tightly associated with static chlorite plates. An early foliation $(S_1)_{LVU}$ is preserved also in younger quartzites and metarenites (Carnian succession), where it is defined by abundant quartz and subordinate white mica (Wm I, mean Si content 3.27 p.f.u.). Small, static ankerite is common; ilmenite, magnetite and hematite are very subordinate

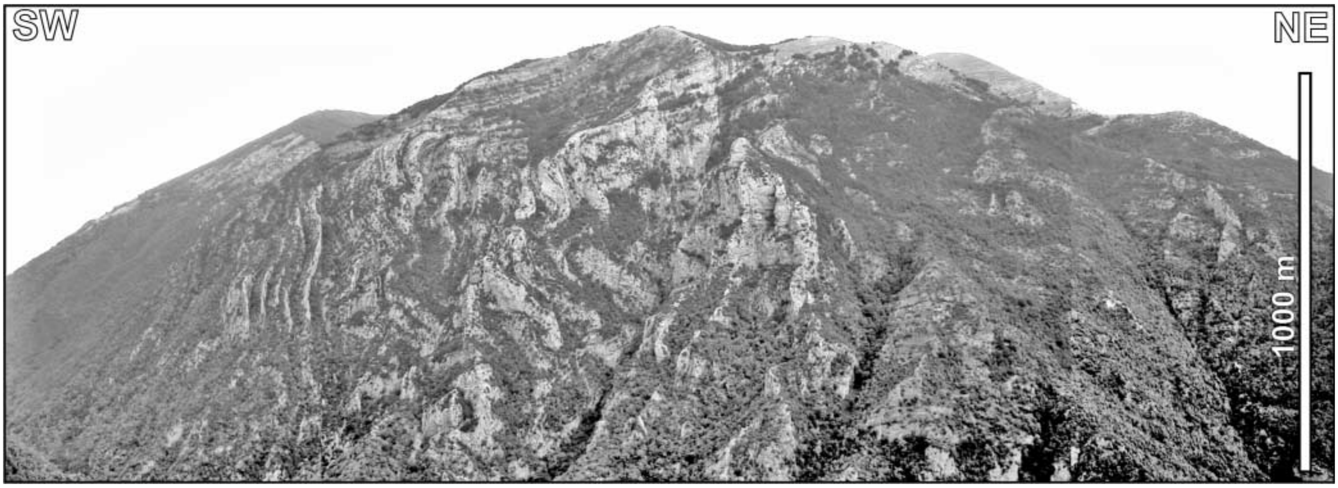

Fig. 6. Major recumbent $(F_3)_{LVU}$ fold in Triassic (Ladinian–Carnian) metacarbonates and phyllites of the Lungro–Verbicaro Unit (La Mula Mt.).

with respect to their occurrence in other lithotypes, and graphite, apatite, rutile, tourmaline and zircon are rare. The attitude of the early foliation $(S_1)_{LVU}$, measured from all lithotypes exposed in the study area, shows a wide distribution, with low to moderate angles of dip being dominant (Fig. 10a).

In Triassic metalimestones, the early foliation $(S_1)_{LVU}$ is preserved in intrafolial $(FA_1)_{LVU}$ folds and it is defined by abundant hematite, calcite, subordinate quartz and rare white mica and chlorite. Associated with this foliation is a subhorizontal stretching lineation (SL_1). The latter (Fig. 10b) is subparallel to the hinge lines of $(FA_1)_{LVU}$ minor folds (Fig. 10c), whose axial planes are also subhorizontal (Fig. 10d). Rare intrafolial folds occur also in the Ladinian marly layers (Fig. 7b).

In Miocene metapelites, the trace of bedding $(S_0)_{LVU}$ is locally preserved in calcareous–quartzitic metapelites as a relict fabric within microlithons. $(S_1)_{LVU}$ is the main foliation in calcareous–quartzitic metapelites, where it is defined by quartz, chlorite (Chl I), white mica (Wm I), calcite, dolomite and graphite. Phyllosilicates occur also as small static crystals, in addition to hematite and ankerite. Further accessories are apatite, rutile, ilmenite, magnetite, tourmaline and zircon. In the metapelites, $(S_1)_{LVU}$ is preserved as a planar fabric deformed by later intrafolial folds $(F_2)_{LVU}$ occurring at both micro- (Fig. 7c) and meso-scale. Fe–Mg-carpholite also occurs in at least two quartz-vein sets (Fig. 7d and e).

D_2 structures. In Triassic phyllites, pre-existing chloritoid is rotated into the main foliation $(S_2)_{LVU}$, as are several accessory and opaque minerals (zircon, apatite, magnetite and pyrite). In quartzites and metarenites, the early foliation $(S_1)_{LVU}$ is preserved within the microlithons defined by the main foliation $(S_2)_{LVU}$. The latter is defined by thin layers of white mica (Wm II, mean Si content 3.10 p.f.u.) + quartz + chlorite (Chl II) + calcite. NE-and SW-dipping shear bands and late, conjugate, ductile and brittle–ductile to brittle shear zones affect the Anisian–Carnian succession.

Triassic metalimestones show a strong grain shape preferred orientation along the main foliation $(S_2)_{LVU}$, which is defined by calcite, white mica, chlorite, hematite and graphite. In thin section, most of the metalimestones appear coarsely recrystallized. They show both symmetric and asymmetric recrystallized pressure fringes around calcite porphyroclasts, suggesting small-scale partitioning between coaxial and non-coaxial strain. Locally, high-strain zones comprising fine-grained recrystallized calcite are present within the Anisian–Carnian part of the succession. These are characterized by a stretching lineation $(SL_2)_{LVU}$, outlined by the elongation of calcite. Syntectonic rotated porphyroclasts and σ-mantled porphyroclasts in these calc-mylonites show a consistent sense of rotation, indicating a general top-to-the-NE sense of shear. The stretching lineation $(SL_2)_{LVU}$ varies from oblique to subparallel to $(F_2)_{LVU}$ fold hinges. Rare sheath folds also occur (Fig. 7f). A well-developed boudinage affects the dolomite levels interposed within the calc-mylonites, as well as quartz veins in intercalated calcareous–quartzitic phyllites. In the latter, the stretching lineation $(SL_2)_{LVU}$ is marked by quartz and calcite grains. The rest of the carbonate succession shows mainly evidence for static recrystallization; however, in the Cirella and Laise areas the Maastrichtian–Aquitanian metaconglomerates and marly metalimestones (Colle Trodo Fm) are affected by a ductile deformation resulting in stretched cherty clasts (Fig. 4b).

In Miocene calcareous–quartzitic metapelites the second foliation $(S_2)_{LVU}$ occurs as a crenulation cleavage related to meso-scale tight to isoclinal $(F_2)_{LVU}$ folds. In metapelites $(S_2)_{LVU}$ becomes the main foliation and is defined by irregular layers of recrystallized white mica (Wm II) + chlorite (Chl II) and of quartz + calcite. Carpholite-bearing quartz veins included in Miocene metapelites are also folded around $(F_2)_{LVU}$ isoclinal structures (Fig. 7e).

The poles to $(S_2)_{LVU}$ tend to be distributed around a NNE-striking great circle (Fig. 10e), whereas the stretching lineation $(SL_2)_{LVU}$ mainly plunges toward the SW (Fig. 10f). Second phase folds $(F_2)_{LVU}$ display mainly subhorizontal hinges (Fig. 10g) and dominantly gently inclined axial surfaces (Fig. 10h). Fold hinges and associated crenulation (microfold) lineations show two main clusters: the (subordinate) NW–SE trend is related to folds in the moderately deformed Miocene metapelites, whereas the main NE–SW trend is related to folds in highly deformed Triassic phyllites and metalimestones, where fold hinges tend to lie at a low angle to the stretching lineation $(SL_2)_{LVU}$ (Fig. 10f).

D_3 structures. The distribution of $(S_2)_{LVU}$ poles around a great circle (Fig. 10e) reflects third phase $(F_3)_{LVU}$ folding characterized

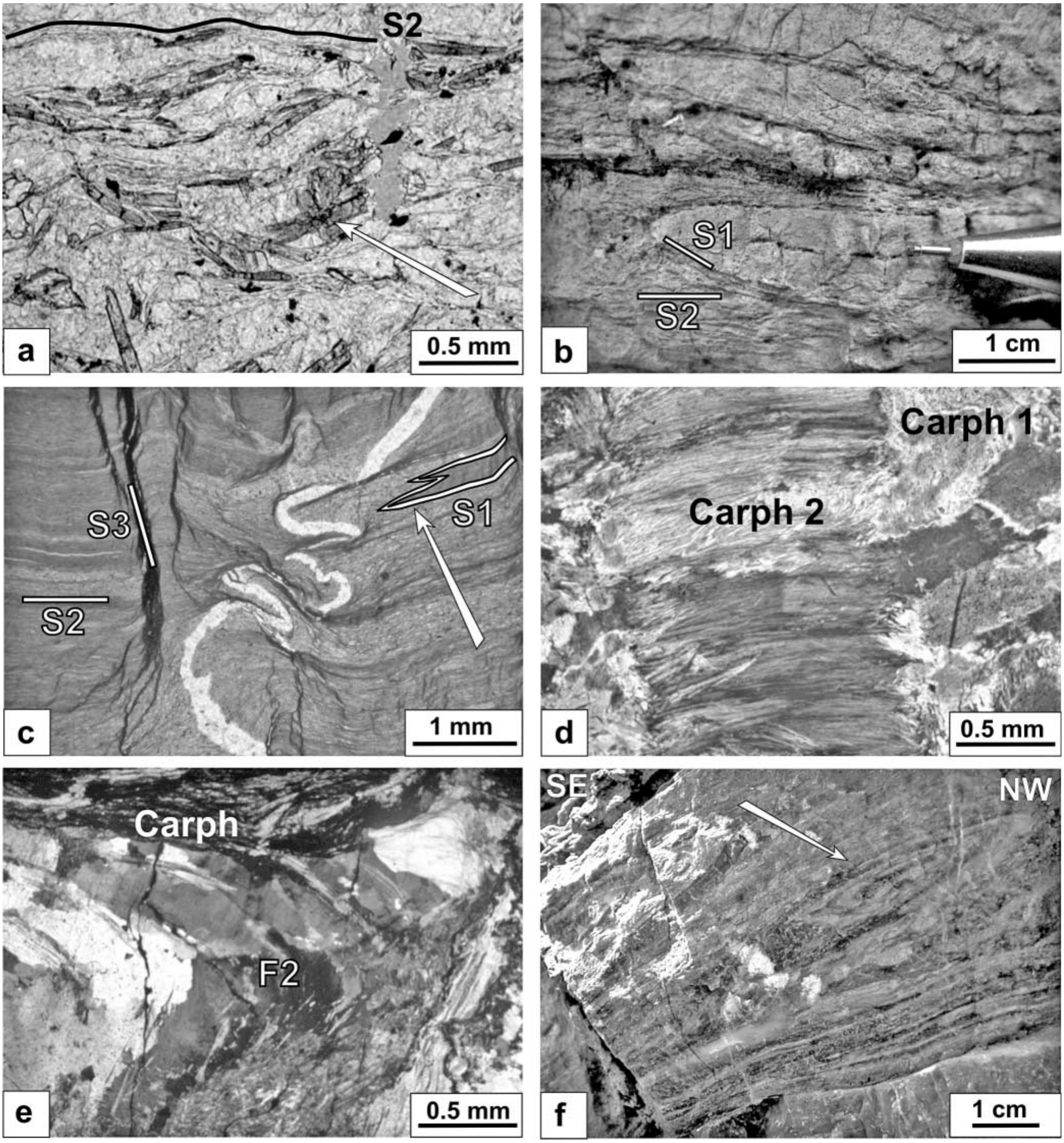

Fig. 7. Microstructures from the Lungro–Verbicaro Unit. (**a**) $(S_1)_{LVU}$-related chloritoid, partly preserved as rosettes (arrowed) and rotated parallel to the main $(S_2)_{LVU}$ foliation (Anisian phyllites, San Donato area). (**b**) Intrafolial $(F_2)_{LVU}$ fold and main $(S_2)_{LVU}$ foliation in Ladinian marly metalimestones (Abatemarco River valley). (**c**) Intrafolial $(F_2)_{LVU}$ fold (arrowed), main $(S_2)_{LVU}$ foliation and $(S_3)_{LVU}$ crenulation cleavage in Miocene metapelites (Verbicaro area). (**d**) Two generations of carpholite in quartz vein from Miocene metapelites (Maierà area). (**e**) Early carpholite fibres deformed by $(F_2)_{LVU}$ folds (Maierà area). (**f**) Sheath fold (arrowed) in calc-mylonite (San Nicola Arcella area).

by kilometre-scale, recumbent, SW-vergent backfolds and associated SW-facing parasitic folds; therefore, a gently WNW-plunging statistical (π) fold axis is obtained. Meso-scale $(F_3)_{LVU}$ folds affecting the main foliation $(S_2)_{LVU}$ in Triassic phyllites and the mylonitic foliation in calc-mylonites are mainly tight to isoclinal. A crenulation (microfold) lineation $(L_3)_{LVU}$ is also commonly associated with $(F_3)_{LVU}$ folds in Carnian and Anisian phyllites, whereas a related planar fabric $(S_3)_{LVU}$ is only discontinuously developed. Crenulation cleavage domains are defined by phengitic white mica (Wm III, mean Si content

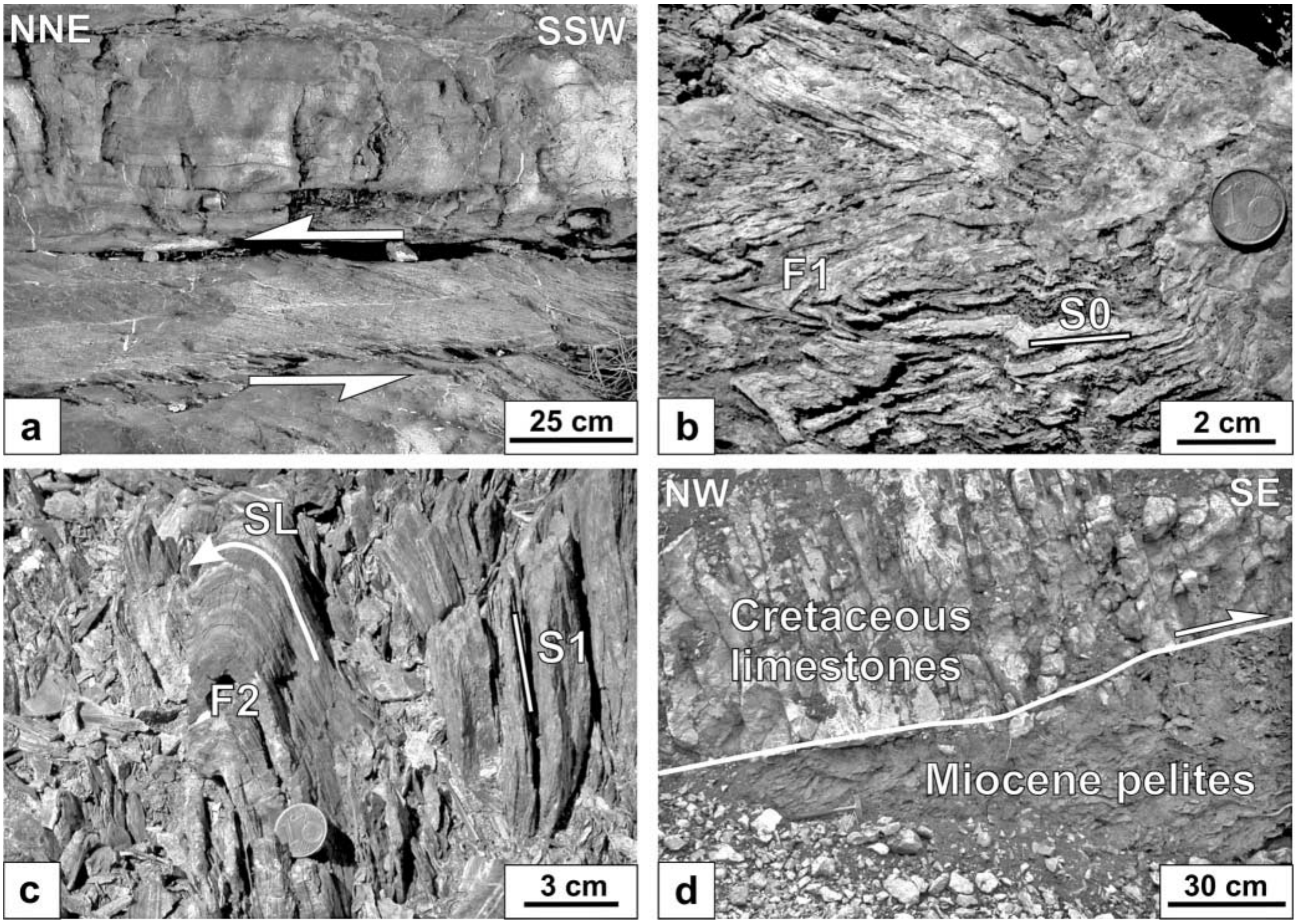

Fig. 8. Outcrop structures from the Pollino–Ciagola Unit. (**a**) Ductile shear zone in Jurassic marls intercalated with limestone beds (Aieta). (**b**) Tight $(F_1)_{PCU}$ folds in Palaeogene limestones (calcaires plaquettés; Cozzo Petrara area). (**c**) Stretching lineations $(SL_1)_{PCU}$ refolded by isoclinal $(F_2)_{PCU}$ folds in Palaeogene foliated limestones (Cozzo Petrara area). (**d**) Intensely cleaved Cretaceous limestones thrust onto Miocene pelites (Maratea area).

3.17 p.f.u.) and chlorite in the Al-rich rock types. $(S_3)_{LVU}$ foliation in metalimestones consists of a spaced pressure-solution cleavage, also discontinuously developed.

In Miocene metapelites, the foliation $(S_2)_{LVU}$ is folded around open to closed box-shaped $(F_3)_{LVU}$ folds. An $(S_3)_{LVU}$ foliation is well developed in metapelites (Fig. 7c).

Mesoscopic hinges of third phase $(F_3)_{LVU}$ folds and related crenulation (microfold) lineations are subhorizontal to gently plunging; although they show a wide range of orientations, two dominant (NE and NW) trends can be recognized (Fig. 10i). Axial surfaces associated with these folds (Fig. 10j), as well as the related $(S_3)_{LVU}$ foliation (Fig. 10k), are dominantly gently to moderately inclined, although steep attitudes also occur.

Fe–Mg-carpholite occurrence

The occurrence of Fe–Mg-carpholite in the Lungro–Verbicaro Unit is significant in terms of the related *P–T* conditions and deserves further attention. Carpholite-bearing quartz veins and segregations (Fig. 7d and e) consist of centimetre-sized, calcite-rich, mostly lens-shaped bodies, strongly deformed by tight to isoclinal $(F_2)_{LVU}$ and $(F_3)_{LVU}$ folds. At outcrop, quartz veins and segregations exhibit coarse-grained, variably altered carpholite. The less altered carpholite occurs in white to greenish prismatic crystals, 1–10 cm long and up to 1 cm wide, whereas the strongly altered carpholite tends to be pseudomorphed by white mica and chlorite.

At the micro-scale, carpholite is mainly preserved in the least deformed parts of the veins, which consist mainly of quartz showing widespread subgrain development and undulose extinction. Carpholite occurs in different structural sites (the corresponding variations in composition of this index mineral will be described below). Pristine carpholite is rarely preserved in large white to greenish prismatic crystals (Cp I, mean X_{mg} value 0.43), mainly associated in radial aggregations, and included in coarse-grained, post-$(S_1)_{LVU}$ quartz and calcite domains. Pre-existing prismatic carpholite (Cp I) is recrystallized with subgranular structure (Cp II, mean X_{mg} value 0.48) along the main foliation $(S_2)_{LVU}$. Later needle-like crystals (Cp III, mean X_{mg} value 0.51) are also oriented along $(S_2)_{LVU}$, together with quartz and calcite.

Pollino–Ciagola Unit

The Pollino–Ciagola Unit experienced heterogeneous ductile deformation localized at several levels within the Jurassic–Miocene calcareous–pelitic succession (Iannace & Vitale 2004; Vitale & Iannace 2004). The best-developed ductile strain zone is localized in the upper part of the succession, ranging in

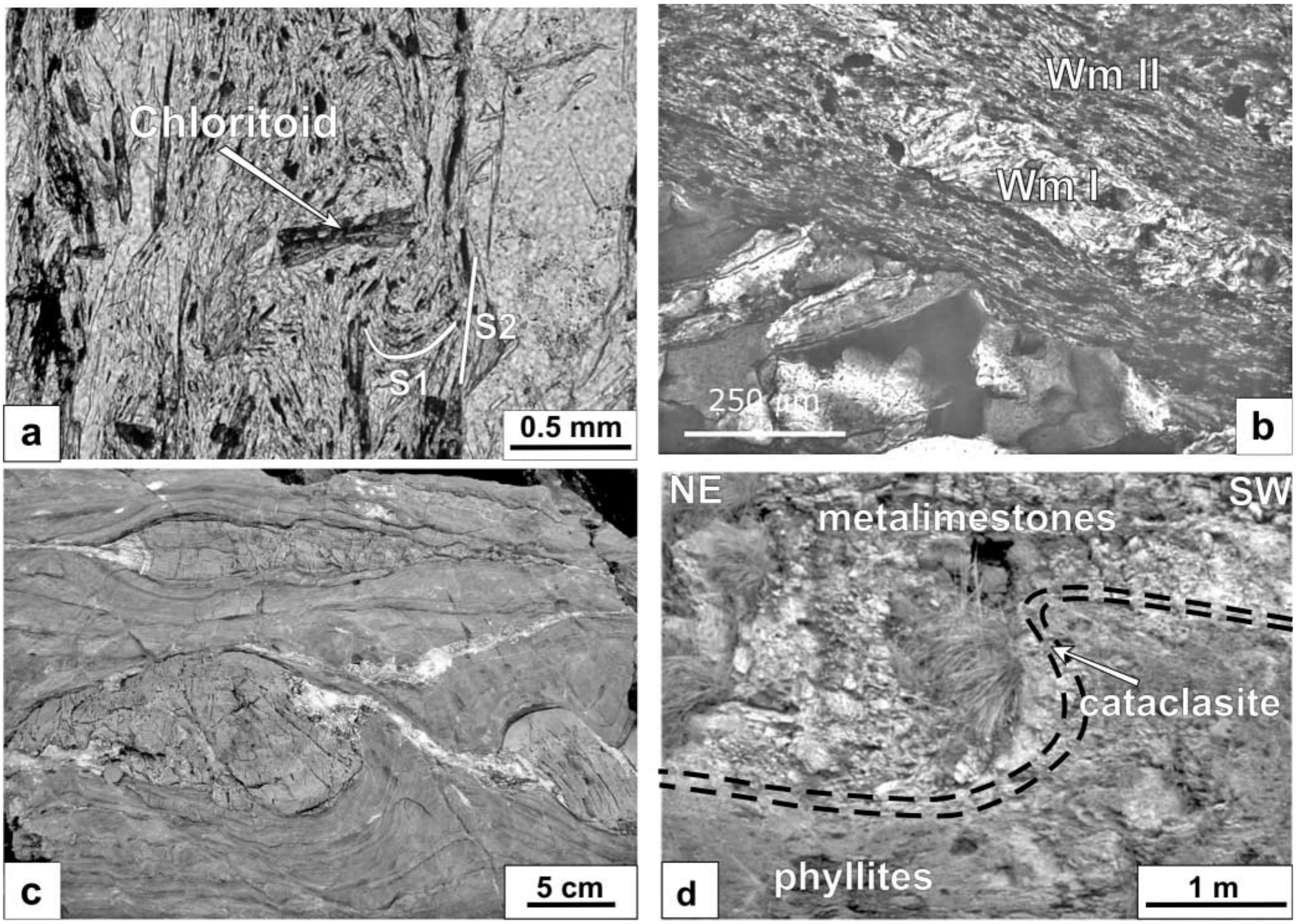

Fig. 9. Deformation features from the Cetraro Unit. (**a**) Microphotograph of $(F_2)_{CU}$ intrafolial fold in phyllites with preserved chloritoid in microlithon (Cetraro). (**b**) Coarse- and fine-grained phengite (Wm I and Wm II) related to $(S_1)_{CU}$ and $(S_2)_{CU}$ foliations, respectively (phyllites, Cetraro). (**c**) Dolomite boudins embedded in metalimestones (Scogliera dei Rizzi, Cetraro). (**d**) Brittle extensional fault folded by $(F_3)_{CU}$ structure (Scogliera dei Rizzi, Cetraro).

thickness from *c*. 200 m (Campotenese area) to *c*. 50 m to the NW (Maratea area). This deformation zone is characterized by variably deformed Jurassic–Cretaceous oncoidal packstones and Palaeogene conglomerates. These rocks are characterized by a foliation $(S_1)_{PCU}$ defined by a shape fabric of deformed oncoids and clasts (Vitale & Iannace 2004). Shears zones tens of centimetres thick occur in the Jurassic–Cretaceous interval, showing a consistent top-to-the-NNE sense of shear (Fig. 8a). In the most deformed rocks, a stretching lineation $(SL_1)_{PCU}$ also occurs. In Miocene marls intrafolial folds $(F_1)_{PCU}$ are developed (Fig. 8b), the hinge lines of which tend to be parallel to the stretching lineation. The main foliation $(S_1)_{PCU}$ is folded around close to isoclinal, frequently sharp-hinged folds $(F_2)_{PCU}$. In the marls the folds are strongly non-cylindrical and refold the pre-existing stretching lineation (Fig. 8c). A crenulation cleavage $(S_2)_{PCU}$ occurs in pelites and marls.

Poles to the main foliation $(S_1)_{PCU}$ show dominantly gently inclined attitudes (Fig. 10l), and the associated stretching lineation $(SL_1)_{PCU}$ is dominantly subhorizontal and north–south trending (Fig. 10m). The hinge lines of minor $(F_2)_{PCU}$ folds are mainly subhorizontal and ENE–WSW trending (Fig. 10n). The related axial planes range from subhorizontal to steeply NNW and SSE dipping (Fig. 10o), similarly to the associated $(S_2)_{PCU}$ discontinuous cleavage (Fig. 10p). Minor NE-directed thrust faults also occur within the Pollino–Ciagola Unit (Fig. 8d).

Cetraro Unit

The earliest recorded structure is a slaty cleavage $(S_1)_{CU}$ preserved in microlithons or in intrafolial folds $(F_1)_{CU}$ in the phyllite succession (Fig. 9a). Coarse-grained white mica (Fig. 9b, Wm I, mean Si content 3.20 p.f.u.) and chloritoid (Fig. 9a) are associated with this first foliation $(S_1)_{CU}$, which is mainly subhorizontal (Fig. 10q). Later $(F_2)_{CU}$ mesoscopic folds affecting the phyllites are characterized by subhorizontal, NE–SW-trending hinges (Fig. 10r) and axial planes ranging from subhorizontal to gently inclined (Fig. 10s). The related second foliation $(S_2)_{CU}$ is marked by fine-grained white mica (Fig. 9b, Wm II, mean Si content 3.09 p.f.u.). A crenulation lineation $(CL_2)_{CU}$ is also associated with $(F_2)_{CU}$ folds, being parallel to the $(F_2)_{CU}$ fold hinges (Fig. 10r).

In the metalimestones the main foliation is the second $(S_2)_{CU}$, mostly subhorizontal (Fig. 10t). The associated stretching lineation (SL_2) is gently SW plunging (Fig. 10u) and therefore mostly parallel to $(F_2)_{CU}$ fold hinges. NE–SW stretching is also marked by extensional brittle–ductile shear zones, mostly occurring as

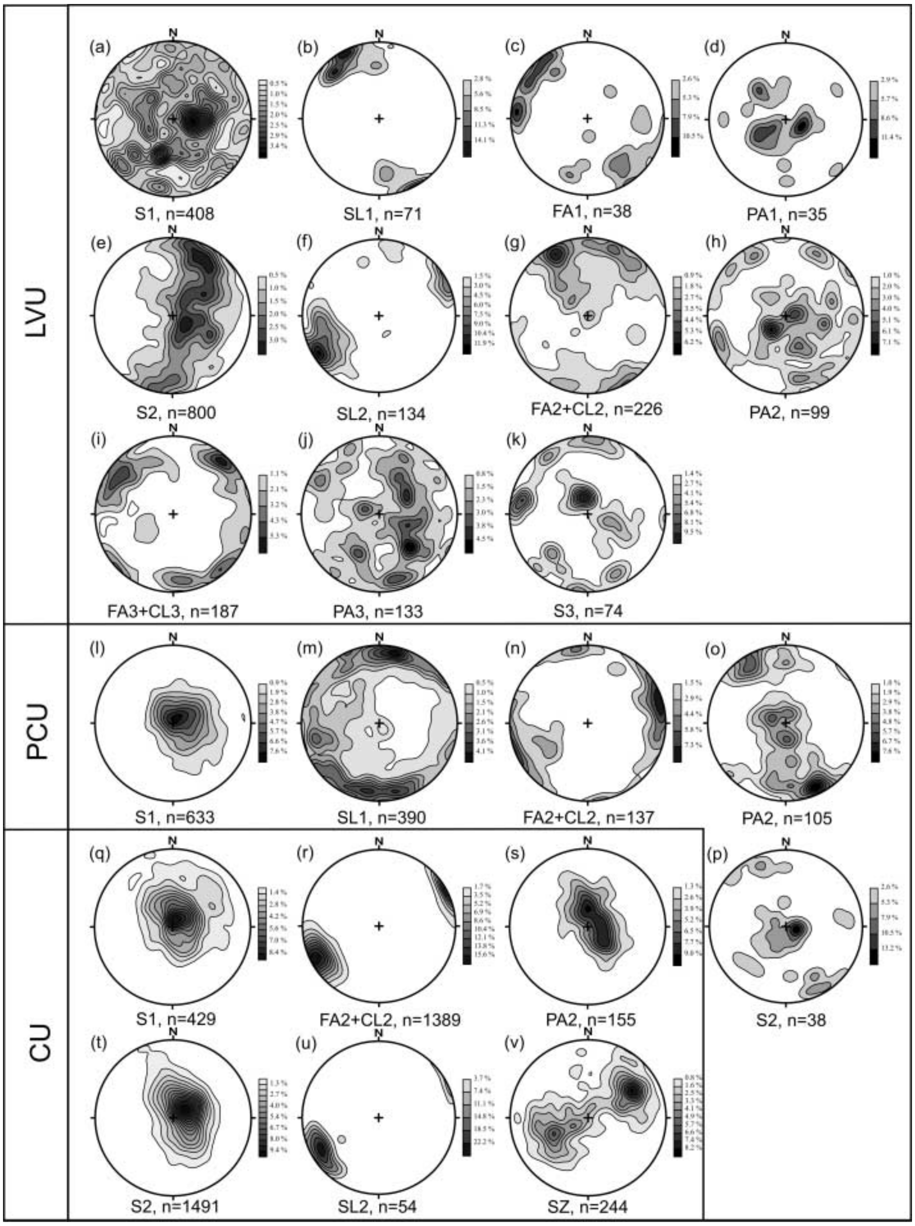

Fig. 10. Orientation data (lower hemisphere Schmidt projections) for the Lungro–Verbicaro Unit (LVU), Pollino–Ciagola Unit (PCU) and Cetraro Unit (CU) structural elements. S, poles to foliation; SL, stretching lineations; FA, fold hinges; CL, crenulation (microfold) lineations; PA, poles to axial planes; SZ, poles to brittle–ductile shear zones.

conjugate sets, which are moderately NE or SW dipping (Fig. 10v). The dolomite levels are affected by a severe boudinage (Fig. 9c), often characterized by en echelon boudins. $(F_3)_{CU}$ folds refold all former structures including brittle–ductile and brittle shear zones (Fig. 9d). These folds are asymmetric, mainly NW or SE vergent, and generally show subhorizontal hinges.

Petrology

In the Lungro–Verbicaro Unit, the main occurrences of metamorphic index minerals of the FeO–MgO–Al_2O_3–SiO_2–H_2O (FMASH) system are found in the slates and associated quartz veins of the Miocene metapelites (Scisti del Fiume Lao Fm), and in the Triassic phyllites. In Miocene metapelites, the metamorphic index minerals are Fe- to Mg-carpholite and chlorite, in addition to phengitic white mica. The main metamorphic assemblages are: (1) chlorite–phengite–quartz–calcite–dolomite and rarely Fe–Mg-carpholite in the slates (samples SC1, SC2.1, SC2.4A and B, MM1, MM2); (2) Fe–Mg-carpholite–quartz–calcite–chlorite–phengite in the quartz veins (samples SC1, SC2.1A and B, SC2.1, SC2.2, SC2.3, SC2.4, MM1, MM2). In Triassic phyllites of the Lungro–Verbicaro Unit, the main metamorphic assemblage is quartz–chloritoid–phengite.

In the Cetraro Unit, metamorphic index minerals of the FeO–MgO–Al_2O_3–SiO_2–H_2O (FMASH) occur in phyllites. The main metamorphic assemblage is quartz–chlorite–chloritoid–white mica.

Mineral chemistry

Mineral analyses on carpholite, chlorite, white mica, chloritoid, accessories and opaques have been obtained using a Cambridge Instruments Stereoscan 360 SEM equipped with an Oxford Instruments Energy 200 energy-dispersive spectrometer, at the Department of Mineralogic and Petrologic Sciences of the University of Turin (Italy). Carpholite and chlorite are normalized according to Azañon & Goffé (1997). Chloritoid is normalized according to Chopin *et al*. (1992) and Fe^{3+} calculated according to Azañon & Goffé (1997). Structural formulae of other minerals are based on the software of Ulmer (1993). Analytical data are available online at http://www.geolsoc.org.uk/SUP18287. A hard copy can be obtained from the Society Library.

Miocene metapelites of the Lungro–Verbicaro Unit. Fe–Mg-carpholite has been analysed essentially from quartz veins included within the Miocene metapelites. It ranges between Fe-carpholite (X_{Mg} 0.39–0.50) and Mg-carpholite (X_{Mg} 0.51–0.56). There is a tight correlation between microstructural sites and chemical composition of carpholite. Carpholite develops between $(S_1)_{LVU}$ and $(S_2)_{LVU}$ foliation, changing its composition, in a narrow range, with an increase of Mg and Ti. Mn contents are very low (0.0–0.03 a.p.f.u.).

In the slates, chlorite shows two compositions. Syn-$(S_1)_{LVU}$ (Chl I) is characterized by X_{Mg} ranging from 0.41 to 0.47, Al content from 2.78 to 2.88 a.p.f.u. and Si content from 2.71 to 2.81 p.f.u.; syn-$(S_2)_{LVU}$ (Chl II) displays X_{Mg} values between 0.36 and 0.39, Al ranging from 2.81 to 2.87 a.p.f.u. and Si from 2.67 to 2.71 p.f.u. Chlorite after carpholite in the quartz veins (X_{Mg} 0.38–0.42) shows a very similar composition to Chl II of the slates. White mica occurring in the slates along $(S_1)_{LVU}$ exhibits a phengitic composition, with Si content of 3.14–3.38 p.f.u., Fe^{2+} + Mg value of 0.07–0.26 and K/(K + Na+ Ca) ratio of 0.35–1.00. White mica occurring along $(S_2)_{LVU}$ shows a progressive decrease of celadonitic substitution.

In the quartz veins, post-S_2 to late or post-S_3 white mica after carpholite exhibits intermediate celadonitic substitution (Si content 3.13–3.23 p.f.u.) with respect to the white mica values recorded from slates.

Triassic phyllites of the Lungro–Verbicaro Unit. Chloritoid occurring along $(S_1)_{LVU}$ shows a progressive increase of Mg from syn- or late-$(S_1)_{LVU}$ relict cores (mean X_{Mg} 0.20) to syn-$(S_2)_{LVU}$ rims of crystals (mean X_{Mg} 0.27). Mn is not detected. Al content ranges from 3.84 to 3.97 a.p.f.u., slightly lower than the ideal value of 4 a.p.f.u.; Fe^{3+} content ranges from 0.03 to 0.16 p.f.u. White mica exhibits the most phengitic values in rare post-$(S_1)_{LVU}$ flakes and preserved cores (Si content 3.21–3.55 p.f.u.; Fe^{2+} + Mg value 0.15–0.31; K/(K + Na) ratio 0.86–0.94). Low phengitic compositions, with Si content 3.05–3.13 p.f.u., Fe^{2+} + Mg 0.04–0.12 and K/(K + Na) 0.87–0.90, occur in syn-$(S_2)_{LVU}$ crystals. White mica after chloritoid exhibits intermediate celadonitic substitution (Si content 3.11–3.21 p.f.u.).

Phyllites of the Cetraro Unit. Chloritoid is an iron-rich variety with X_{Mg} between 0.08 and 0.13. There is a negligible zoning with a weak increase of Mg from cores to rims. Mn content, not detected in sample Pe2, is uniformly very low in sample Pe10 (0.05–0.07 a.p.f.u.). Al content ranges from 3.83 to 3.89 a.p.f.u., slightly lower than the ideal value of 4 a.p.f.u.; Fe^{3+} content is 0.11–0.17 p.f.u.

Chlorite from two analysed samples shows different compositions, indicating a different chemistry of the rocks. However, within each sample, chlorite of different generations (Chl I and Chl II) displays homogeneous compositions (X_{Mg} 0.35–0.38, Al content 2.80–2.88 a.p.f.u. and Si 2.57–2.60 p.f.u., in sample Pe2; X_{Mg} 0.42–0.43, Al content 2.82–2.93 a.p.f.u., Si 2.54–2.57 p.f.u., in sample Pe10).

The composition of white mica shows a slight progressive decrease of the celadonitic content from $(S_1)_{CU}$ to $(S_2)_{CU}$ related laminae. The most phengitic values are exhibited by the post-$(S_1)_{CU}$ flakes with preserved relict cores, showing a Si content of 3.18–3.21 p.f.u., an Fe^{2+} + Mg value of 0.13–0.19 and a K/(/K + Na) value of 0.91–0.93. Very low phengitic compositions, with Si content 3.05–3.14 p.f.u., Fe^{2+} + Mg 0.06–0.10 and K/(/K + Na) 0.76–0.90, are shown by crystals occurring along $(S_2)_{CU}$.

P–T *paths*

The occurrence of metamorphic Fe–Mg-carpholite in the Miocene metapelites of the Lungro–Verbicaro Unit provides evidence for high-pressure and low- or medium-temperature conditions for this unit. Fe- and Mg-carpholite are exclusively known in metasediments from low-grade to high-grade blueschist facies (e.g. Goffé *et al*. 1988; Jolivet *et al*. 1996; Azañon & Goffé 1997). Further evidence for high-pressure metamorphism in the Lungro–Verbicaro Unit is the presence of Mg-rich chloritoid (with X_{Mg} 0.25–0.29). These data are in good agreement with the reported occurrence of blue amphiboles in the 'limburgites' (Pierattini *et al*. 1975) and in metabasalts contained in the Triassic phyllites (Macciotta *et al*. 1986).

An estimate of *P–T* conditions for the Lungro–Verbicaro Unit, according to the chemical composition of analysed mineral phases, indicates a multistage evolution (Fig. 11). Based on microstructural observation, the formation of carpholite veins occurred between the development of the first foliation $(S_1)_{LVU}$, characterized by phengite with high celadonitic content (Wm I, mean Si content 3.26 p.f.u.) and that of the main foliation $(S_2)_{LVU}$, with phengites showing a lower celadonitic content (Wm II, mean Si content 3.09 p.f.u.). During this interval, Fe–Mg-carpholite is always stable. This is demonstrated by the fact that deformation of early quartz veins is associated with the growth of new carpholite showing increasing Mg content (from that of early veins, with a mean X_{Mg} value of 0.43, to that included in latest veins, with X_{Mg} 0.51). *P–T* trajectories taking into account both the decrease in celadonitic content of phengite as well as the increase in Mg of carpholite imply decompression accompanied by increasing *T* (Fig. 11). A similar trajectory can be reconstructed for the lower part of the Lungro–Verbicaro Unit taking into account the Mg content of chloritoid. Following Rossetti *et al*. (2001), the Mg increase from chloritoid core (mean X_{Mg} 0.20) to rim (mean X_{Mg} 0.27) is consistent with a *T* increase during exhumation and decompression from 1.4 MPa $(D_1)_{LVU}$ to 1.1 MPa $(D_2)_{LVU}$.

The first stage of metamorphism, corresponding to formation of the first foliation $(S_1)_{LVU}$, is characterized, in the Lower Miocene Scisti del Fiume Lao Fm, by a maximum *P* value of 1.3 GPa, with *T* around 320 °C (within the pyrophyllite stability field), as constrained by phengite (Wm I) and chlorite (Chl I, mean X_{Mg} 0.44) in the slates, and by Fe–Mg-carpholite (mean X_{Mg} 0.43) in quartz veins. For Triassic rocks, the maximum reconstructed *P* value is 1.4 GPa, with *T c*. 390 °C, as indicated by phengite (Wm I) and chloritoid core (mean X_{Mg} 0.20) assemblages in Anisian–Ladinian phyllites. Therefore, recorded peak *P–T* values are typical of blueschist-facies metamorphism.

A second stage of metamorphism, corresponding to $(D_2)_{LVU}$ deformation and characterized by decreasing *P* values, is recorded in the Lungro–Verbicaro Unit by the mineral assemblages preserved along the main foliation $(S_2)_{LVU}$. The exhumation path begins with a weak decompression and an increase of temperature reaching about 340 °C for the upper part of Lungro–Verbicaro Unit and 400 °C for the lower part. This is indicated by a decrease of the celadonitic content of white mica (Wm II,

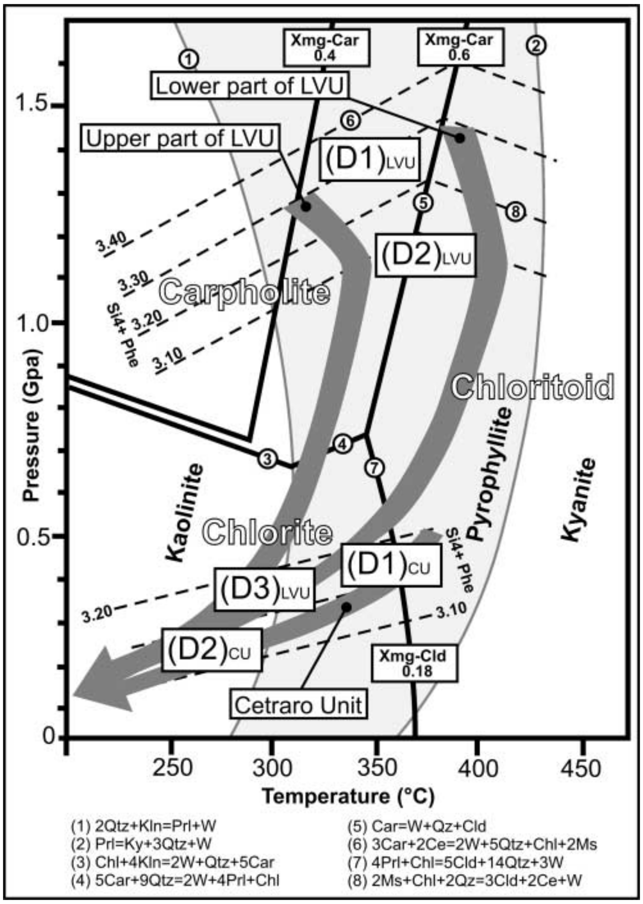

Fig. 11. *P–T* paths calculated from petrological and compositional data on: (i) carpholite, chloritoid and phengite contained in the Triassic phyllites and Miocene metapelites of the Lungro Verbicaro Unit, and (ii) chloritoid and phengite in the Triassic phyllites of the Cetraro Unit. The grid is after Rossetti et al. (2004, fig. 4). Analytical data are in Supplementary Publication (see p. 1178).

mean Si content 3.09 p.f.u.) within both the Scisti del Fiume Lao Fm and Triassic phyllites as well as by the increase of Mg content in the latest carpholite veins and in the chloritoid rim of Triassic phyllites.

Exhumation continued after $(S_2)_{LVU}$ and during $(S_3)_{LVU}$ foliation development, with a retrograde decompression, starting at $P < 1.1$ GPa (within the pyrophyllite field) both in the Lower Miocene rocks (Scisti del Fiume Lao Fm) and in the Anisian–Ladinian phyllites, as indicated by the growth of chlorite and phengitic white mica as alteration products of both carpholite and chloritoid (chlorite, mean X_{Mg} 0.40; Wm III, mean Si content 3.17 p.f.u.). The reconstructed evolution is similar to that recorded in subducted continental crust-derived metasedimentary units of other Mediterranean regions (e.g. Jolivet *et al*. 2003). Rossetti *et al*. (2004) depicted a similar thermobaric evolution based on perfectly overlapping data obtained from Fe–Mg-carpholite found at several localities in northern Calabria. They attributed their samples to the oceanic-derived Frido Unit. However, taking into account the location given in their paper and our remapping of the area, the samples actually came from the Scisti del Fiume Lao Fm of our Lungro–Verbicaro Unit, and are therefore fully compatible with our data.

According to the mineral assemblage of the phyllites, a different thermobaric evolution characterizes the Cetraro Unit. The estimated *P–T* conditions for this unit indicate a polyphase greenschist-facies evolution (Fig. 11). There is no evidence of HP conditions, in agreement with the data of Rossetti *et al*. (2004). In fact, chloritoid has X_{Mg} values of 0.08–0.13, white mica I has a maximum Si content of 3.21 p.f.u., and the rocks contain chlorite. The mineral assemblage related to syn- and post-$(D_1)_{CU}$ developed at *P c*. 0.4 GPa and *T c*. 380°C (within the pyrophyllite field). The evolution continues during $(S_2)_{CU}$ and $(S_3)_{CU}$ foliation development with a slight retrograde path (Wm with mean Si content of 3.09 p.f.u.).

Thermochronology

New radiometric ages were obtained by apatite fission-track analysis on Lungro–Verbicaro Unit rocks, and $^{40}Ar/^{39}Ar$ dating of white micas from the Cetraro Unit. Analytical data are available as a Supplementary Publication (see p. 1178).

Important timing constraints for the tectonic evolution of the former are provided by biostratigraphic data, discussed in previous sections, which show that the stratigraphic succession of the Lungro–Verbicaro Unit extends to the early Miocene. Therefore, almost the whole duration of the burial–exhumation cycle may be obtained by integrating these stratigraphic constraints with cooling ages obtained by very low-*T* thermochronology (apatite fission-track dating). In contrast, the stratigraphic succession of the Cetraro Unit is much more reduced and seemingly comprises only post-Triassic rocks. Consequently, there are no direct stratigraphic relationships to constrain its tectonic evolution. For the Cetraro Unit, $^{40}Ar/^{39}Ar$ data were obtained from white mica, which complement those already available for this unit (Borsi & Dubois 1968; Rossetti *et al*. 2004).

Apatite fission-track dating was carried out on two samples from metabasalts ('limburgites') of the Lungro–Verbicaro Unit collected at Alberosa near Verbicaro. Eighteen and 20 grains have been dated respectively in samples MDE245 and MDE246. In both samples, the χ^2 value is passed, thus demonstrating a very low dispersion in grain age distribution. The calculated central age therefore should reflect the timing of cooling below the *c*. 110 °C isotherm. However, relating a fission-track age to closure temperature is plausible only when it can be demonstrated that no significant annealing has occurred post-closure time. This requires rapid cooling, which should be reflected in the track length distribution. Unfortunately, in the samples analysed here we could not measure a statistically meaningful amount of track lengths. However, stratigraphic constraints (i.e. thickness of the overlying deposits) suggest that no major burial affected the sampled rocks after their exhumation. We can therefore conclude that these data indicate that the Lungro–Verbicaro Unit was exhumed through the closure temperature (*c*. 110 °C) of the apatite system during the Tortonian. The age difference between the two samples is not significant given that ± 1 Ma errors are reported.

Ar data were collected from two aliquots of white mica handpicked from a crushed whole-rock sample (sample Pe2; 39°32′17″N, 15°54′24″E) collected at Cetraro. White mica age spectra are reported with 1σ errors and plotted in Figure 12 with respect to cumulative ^{39}Ar%. Each step-heating profile represents the analysis of a single mica aliquot. White mica data from the two aliquots yield similar results, with the ages generally increasing with increasing ^{39}Ar% released (a function of temperature; Fig. 12). In both cases, the age range is similar (15–35 Ma and 12–32 Ma) and no plateaux can be identified within either of the

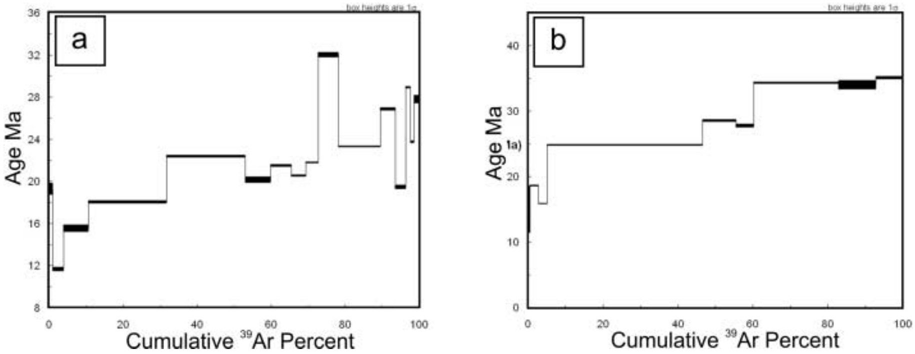

Fig. 12. $^{40}Ar/^{39}Ar$ age spectra. (**a**) Sample SM1. (**b**) Sample SM2. Analytical data are in Supplementary Publication (see p. 1178).

two aliquots. Correlation plots of $^{39}Ar/^{40}Ar$ v. $^{36}Ar/^{40}Ar$ show no systematic trends (data not shown) that can be related to the heterogeneous presence of excess ^{40}Ar.

Discussion

Structural styles and timing of the deformation

The stratigraphic, structural, petrological and radiometric data presented above may be integrated to determine the tectonic history of the Lungro–Verbicaro Unit and Pollino–Ciagola Unit. Despite common strain localization phenomena, a coherent structural evolution can be obtained for each tectonic unit as defined here. This represents significant progress for an area that for a long time has been considered a geological puzzle (for a review, see Iannace *et al*. 2005b). Only for the Cetraro Unit is a complete reconstruction of the tectonic evolution hindered by the lack of stratigraphic constraints and the not fully conclusive radiometric data (Ar data show large age ranges, probably representing mixed detrital–growth ages).

The main deformation of the Lungro–Verbicaro Unit occurred under blueschist-facies conditions during $(D_1)_{LVU}$ and $(D_2)_{LVU}$ deformation stages, as shown by the mineral assemblages associated with the related structures. These include: (1) a dominantly flat-lying foliation $(S_2)_{LVU}$, with which a mineral or stretching lineation $(L_2)_{LVU}$ is locally associated; (2) metre- to millimetre-sized, recumbent, isoclinal intrafolial folds $(F_2)_{LVU}$; (3) boudinage of competent (mostly dolomite) beds; (4) low-angle to moderately dipping extensional shear zones. Later, $(D_3)_{LVU}$, deformation is characterized by the development of spectacular regional backfolds and associated parasitic structures $(F_3)_{LVU}$, with which a crenulation cleavage $(S_3)_{LVU}$ is associated.

$(D_1–D_2)_{LVU}$-related structures appear to be associated with small-scale partitioning of coaxial and non-coaxial strain. Coaxial strain components are consistent with dominantly vertical shortening and horizontal extension. Gently SW-dipping extensional shear zones, recording non-coaxial strain, also occur in some calcareous levels. $(D_3)_{LVU}$ structures include both regional and minor SW-vergent folds. Similar structures have been observed in other HP–LT terranes (Searle *et al*. 2004) and have been associated with back-shearing related to return flow within the subduction channel, an interpretation that appears to be consistent also in our instance. On the other hand, these hinterland-vergent folds could also be related to a late phase of backthrusting or -folding within the thrust belt.

It seems likely that much of the Lungro–Verbicaro Unit deformation records the early part of the exhumation-related tectonic evolution, as it retains kinematic coherence with structures developed under progressively lower pressure conditions. It should be noted that the stability field for blueschists is such that substantial decompression may be achieved before greenschist-facies conditions are reached. Based on stratigraphic constraints, the age of Lungro–Verbicaro Unit internal deformation (stages $(D_1–D_3)_{LVU}$) can be bracketed between the Aquitanian (age of the Scisti del Fiume Lao Fm, i.e. stratigraphic top of the Lungro–Verbicaro Unit succession) and the latest Tortonian (age of the post-orogenic deposits unconformably overlying the exhumed terranes).

Whole-rock K–Ar data from eight metabasite (‘limburgites’) samples from the Lungro–Verbicaro Unit yielded a best-fit isochron providing an age of 18.55 ± 0.09 Ma (Pierattini *et al*. 1975). A significantly older ($^{40}Ar/^{39}Ar$) age (33.5 ± 0.1 Ma) was obtained by Rossetti *et al*. (2004) from one sample (CLB8) that they claimed was taken from the Triassic phyllites included in our Lungro–Verbicaro Unit. Re-examination of their data from this sample indicates that the older ages of their spectra (steps 6–19) fall on a mixing line with $^{40}Ar/^{36}Ar$ of *c*. 1250 and a radiogenic component corresponding to an age of *c*. 30 Ma, slightly younger than their bulk age and significantly younger than the 33–38 Ma age interpreted as minimum estimates of the timing of mica crystallization during nappe formation and crustal thickening. Indeed, an age of 30 Ma was interpreted by Rossetti *et al*. (2004) to reflect the onset of exhumation. Although this may well be the case for older cooling stages occurring within the oceanic subduction-related accretionary prism, subduction and subsequent exhumation of continental margin-derived rocks is clearly much younger (i.e. Miocene), as indicated by the stratigraphic constraints provided in this paper. As the sampling area of sample CLB8 of Rossetti *et al*. (2004) is geologically rather complex and includes outcrops of phyllites that are actually cover rocks to the ophiolitic units, a detailed discussion of their radiometric data would require a more precise sample location.

Apatite fission-track data may be used to refine the timing defined above based on stratigraphic criteria, as synmetamorphic deformation $(D_1–D_3)_{LVU}$ must have taken place at temperatures that are significantly higher than the closure temperature (*c*. 110 °C) of the apatite system. Therefore, such deformation stages are older than 10 Ma. Apatite fission-track data also show that the latest part of the Lungro–Verbicaro Unit exhumation path (from a depth of *c*. 4 km to the surface) occurred entirely during the Tortonian. These ages are also in good agreement with available fission-track data from the crystalline basement tectonic units of northern Calabria. These are characterized by apatite fission-track ages in the range of 15 ± 2 to 10 ± 1 Ma (Thomson 1994, 1998).

Deformation of the Pollino–Ciagola Unit took place at submetamorphic conditions, with temperatures never exceeding 300 °C (Vitale *et al*. 2007) and records top-to-the-north or -NE non-coaxial shearing, consistent with north- or NE-directed thrusting of the overlying Lungro–Verbicaro Unit. The final overthrust emplacement of the Lungro–Verbicaro Unit onto the Pollino–Ciagola Unit was younger than Langhian (age of the flysch deposits capping the Pollino–Ciagola Unit succession).

As far as the deformation of the Cetraro Unit is concerned, our structural and petrological data confirm that it occurred under greenschist-facies conditions, as suggested by previous studies (Amodio Morelli *et al*. 1976; Rossetti *et al*. 2004). Similarly to the Pollino–Ciagola Unit, it records top-to-the-north or -NE non-coaxial shearing, consistent with north- or NE-directed thrusting of the overlying Lungro–Verbicaro Unit. Dominant top-to-the-north or -NE sense of shear, related to thrusting and crustal thickening, was also suggested by Rossetti *et al*. (2004) for Cetraro Unit structures. However, the reconstruction of the relationships between the Cetraro Unit and the Lungro–Verbicaro Unit and Pollino–Ciagola Unit is difficult not only because of the characteristics of the metamorphism, but also in terms of its timing. In fact, $^{40}Ar/^{39}Ar$ ages on white micas (see Fig. 12), similarly to those obtained by Borsi & Dubois (1968) and Rossetti *et al*. (2004), are best interpreted in terms of mixing between two different white mica generations (see Fig. 9b), with older ages representing the outgassing of older, larger (detrital?) white mica, whereas the younger ages represent the finer, younger component. The progressive change in ages through the release spectra represents an increasing contribution of the coarser white mica fraction at increasing temperature. The three high-temperature steps from aliquot 1 yield an unweighted mean age of 34.5 ± 0.5 Ma, which is interpreted to represent the age of the coarser white mica generation. This age could correspond to the time of mica crystallization, if the crystallization temperature is less than the closure temperature of the mica. Alternatively, the age may represent the time at which the mica passed through the closure temperature following a thermal event. The grain diameter, parallel to the (001) basal plane, is *c*. 250 μm, implying closure temperatures around 360 °C, although this is not well constrained because of uncertainties in diffusion characteristics and cooling rates (Reddy & Potts 1999).

The fine-grained mica generation within the Cetraro Unit comprises grains of *c*. 25 μm with a closure temperature of the order of 300 °C. The mean ages for the first three steps of aliquot 1 and 2 are 16.5 ± 1.5 Ma and 15.1 ± 2.6 Ma, respectively, and the best estimate of isotopic closure is at *c*. 16 Ma. Again, this could correspond to either crystallization of mica within the D_2 fabric or isotopic closure, depending on the temperature of the D_2 deformation. Given that D_2 fabrics in the Cetraro Unit are associated with the contact of the Cetraro Unit and Lungro–Verbicaro Unit, and that this took place at temperatures around 300 °C (see Fig. 11), the age of 16 Ma is interpreted to approximate the time of this juxtaposition. The original geometry and kinematics of the related tectonic contact have been obscured by subsequent structural reworking and tectonic excision associated with extensional deformation (as described above).

Geodynamic evolution

The stratigraphic, structural and petrological data presented above indicate that, in the studied sector of the Apennine belt, various stages of the evolution of a continental margin are recorded, from passive margin formation to subduction (predating full continent–continent collision) and subsequent exhumation of HP–LT rocks. This evolution can be framed within the history of plate fragmentation, oceanic spreading and later closure taking place from Mesozoic to Neogene time in the central Mediterranean region.

Mesozoic–Palaeogene continental margin architecture. The palaeogeographic reconstruction presented in this paper, summarized in Figure 13, is best interpreted as a record of progressive extension and subsidence of the continental lithosphere up to the final formation of oceanic lithosphere and development of a passive continental margin. An analogous process has been quantitatively analysed in other segments of chain; namely, the Austro-Alpine system in the Alps *sensu stricto* and the South Alpine chain, where the evolution of the Adriatic margin is recorded. A complete section of the basement and sedimentary cover is exposed in the latter. This quantitative assessment of the spatial and temporal evolution of the area of crustal extension has shown that rifting has migrated from east to west from Early Triassic to Middle Jurassic time, a process that has been tentatively ascribed to a progressive strengthening of the extending lithosphere as a result of post-rifting cooling (Bertotti *et al*. 1993; Manatschal & Bernoulli 1999). Such an evolution can be clearly recognized also in the Mesozoic record of northern Calabria.

Iannace *et al*. (2005*b*) have already pointed out that from Anisian to Rhaetian time there is a clear shift in the areas of extension from (present-day) NE to SW. In fact, the Middle Triassic facies trend indicates that the margin of a large shelf was located in the Monte Caramolo area (see Fig. 13a), characterized by persistent synsedimentary tectonics and connected to emergent land to the SW. This shelf flanked a basin area for which we have no direct outcrop evidence in our study area. However, all regional-scale palaeogeographical reconstructions for the Middle Triassic point to the presence of a large basin to the east of the Apennine carbonate platform domain and oceanographically connected to the Palaeotethys area (or East Mediterranean). The remnants of this basin are preserved north of our study area, in Lucania, where they form the Lagonegro thrust sheets.

Starting from Norian time, active extensional tectonics is documented in the southwestern part of the study area (Fig. 13b) by the organic-rich, platy dolomites interpreted as deposited in intraplatform, restricted troughs. These facies have their exact equivalent throughout the Italian peninsula (Cirilli *et al*. 1999) and have been interpreted as the signature of the rifting phase preceding the opening of the Jurassic oceanic areas. The facies distribution shows that at least two basins with an intervening swell were present in the SW, whereas the Maratea and Pollino areas were sites of persistent shallow-water sedimentation. During the Rhaetian (Fig. 13), shallowing in the northeastern areas is witnessed by prograding back-reef megalodontid limestones over slope successions, and even by emersion (Monte Cifolo). In contrast, the pre-existing Norian swell was drowned (Zamparelli *et al*. 1999; Perri *et al*. 2003), suggesting that the area of maximum subsidence was located to the SW. This trend is further confirmed by the Jurassic facies pattern (Fig. 13c): bed thickness and grain sizes of the resedimented facies both decrease toward the SW in the Calcari con Selce Fm. Moreover, radiolarites, which presumably represent the deeper facies of the whole stratigraphic record, occur only in the southwestern area.

In the Lungro–Verbicaro domain during the Late Jurassic–Early Cretaceous a transition occurred to successions deposited on oceanic crust. More precisely, as suggested by Spadea (1982), Bonardi *et al*. (1988) and Cello & Mazzoli (1999), by analogy with Iberian-type continental margins, an area with exhumed continental crust and mantle rocks was present. These are now

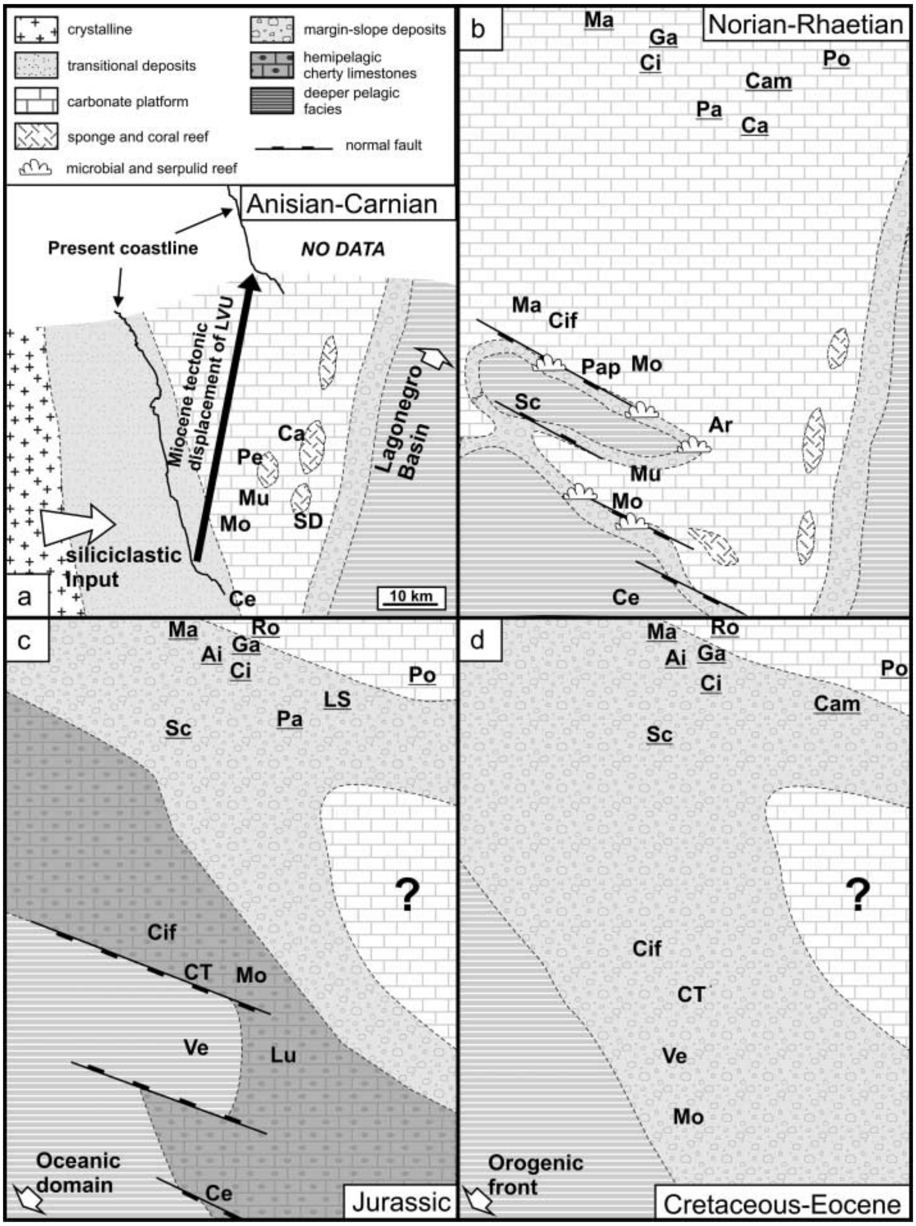

Fig. 13. Interpreted palaeo-environmental evolution of the study area in four key stages of the Meso-Cenozoic history. Localities: Ai, Aieta; Ar, Argentino River; Ca, Caramolo Mt.; Ci, Ciagola Mt.; Cif, Cifolo Mt.; Ce, Cetraro; CT, Colle Trodo; Ga, Gada Mt.; LS, La Serra Mt.; Ma, Maratea; Mo, La Montea Mt.; Mu, La Mula Mt.; Pa, Palanuda Mt.; Pap, Papasidero; Pe, Cozzo del Pellegrino Mt.; Ro, Rossino Mt.; Sc, Scalea; SD, San Donato; Ve, Verbicaro).

partly preserved, as relict blocks, in a tectonic mélange cropping out NE of the Pollino Massif interposed between the ophiolitic Frido and North Calabrian Units (Bonardi *et al*. 1988), where the sedimentary cover extends to Upper Oligocene and Aquitanian sequences, respectively.

The facies pattern of the Pollino–Ciagola Unit also consistently indicates that the transition to the margin and slope areas was toward the west (Fig. 13d). The lagoonal facies occurring in the Pollino succession are replaced, in the Ciagola and Aieta area, by high-energy facies with frequent breccia horizons. The age of the resedimented carbonates points to a progressive backstepping of the platform toward the NE. This depositional margin was basinward replaced by a by-pass margin, as indicated in the LVU by the large stratigraphic gap existing between the Calcari con Selce and Colle Trodo Fms, the latter representing toe-of-slope sediments in Palaeogene time. The similarity between the calcaires plaquettés of the Pollino–Ciagola Unit and the breccias of the Colle Trodo Fm of the Lungro–Verbicaro Unit indicates that, by the end of the Late Palaeogene, the sediments of the Pollino–Ciagola succession also were deposited in a lower slope environment. Within this tectonosedimentary evolution, the basic metavolcanic rocks occurring in the Jurassic part of the Lungro–Verbicaro Unit succession should be viewed as a manifestation of alkaline intraplate magmatic activity roughly contemporaneous with the late rifting stages of the continental margin. However, as explained above, this point requires further investigation.

Neogene tectonic evolution. The high-pressure (1.3–1.4 GPa), relatively low-temperature metamorphism of the Lungro–Verbicaro Unit rocks is inconsistent with burial beneath overthickened continental crust during the collision of two continental margins. Instead, it is interpreted as a result of continental subduction, with basement–cover detachment and return flow within the subduction channel (Fig. 14a–d). This model requires the hanging wall to behave in an essentially rigid fashion. In

Calabria the overriding plate is strongly deformed and highly extended; it does not appear to have behaved entirely rigidly. However, intense deformation and thinning of the upper plate in Calabria occurred during the latest stages of convergence and post-dated a substantial part of the exhumation of the Lungro–Verbicaro Unit. Therefore, the overriding plate was essentially rigid at the time return flow is envisaged to have occurred within the subduction channel (see Fig. 14). It is worth noting that slab roll-back appears to characterize the retreating subduction zone in Calabria (Malinverno & Ryan 1986). In these conditions, the subduction channel is likely to be open, allowing for fast downward and upward circulations and detachment of sediments from the subducting basement (Beaumont *et al*. 1999). Subducted sediments might act as lubricants of the channel and prevent severe shear deformation of the basement, which therefore is not included in the return flow (Jolivet *et al*. 2003). In our instance, the occurrence of low-viscosity material (phyllites) both at the base (Triassic) and top (Miocene) of the Lungro–Verbicaro Unit stratigraphic succession appears also to have prevented severe deformation of the carbonate rocks included within this succession.

Burov *et al*. (2001) have shown that several levels of circulation can be modelled in continental subduction zones, with an upper level corresponding to the accretionary complex and a lower level to the subduction channel. According to Jolivet *et al*. (2003), fast exhumation of HP rocks occurs in the upper part of the subduction channel; rocks are scraped off the down-going plate by shearing along the base of the overriding plate, whereas the positive buoyancy of the continental crust essentially drives exhumation. The displacement of rocks inside the low-viscosity subduction channel induces a component of shear along the contact with the overriding plate at depth that has an ‘extensional’ geometry.

A possible evolution of the Lungro–Verbicaro Unit can be proposed, which integrates available data with the conceptual model described above. During the Aquitanian (Fig. 14a), while the closure of the contiguous oceanic area was going on, the Lungro–Verbicaro Unit succession was still exposed on the sea floor, and was characterized by siliciclastic sedimentation. Subsequently, the Lungro–Verbicaro Unit was subducted to depths in excess of 40 km, reaching *P–T* peak conditions typical of blueschist-facies metamorphism (Fig. 14b). Partitioned coaxial and non-coaxial deformation affected the Lungro–Verbicaro Unit during the early stages of tectonic exhumation (Fig. 14c), as witnessed by $(D_1–D_2)_{LVU}$-related structures. Coaxial strain components are consistent with a maximum shortening (*z*) axis

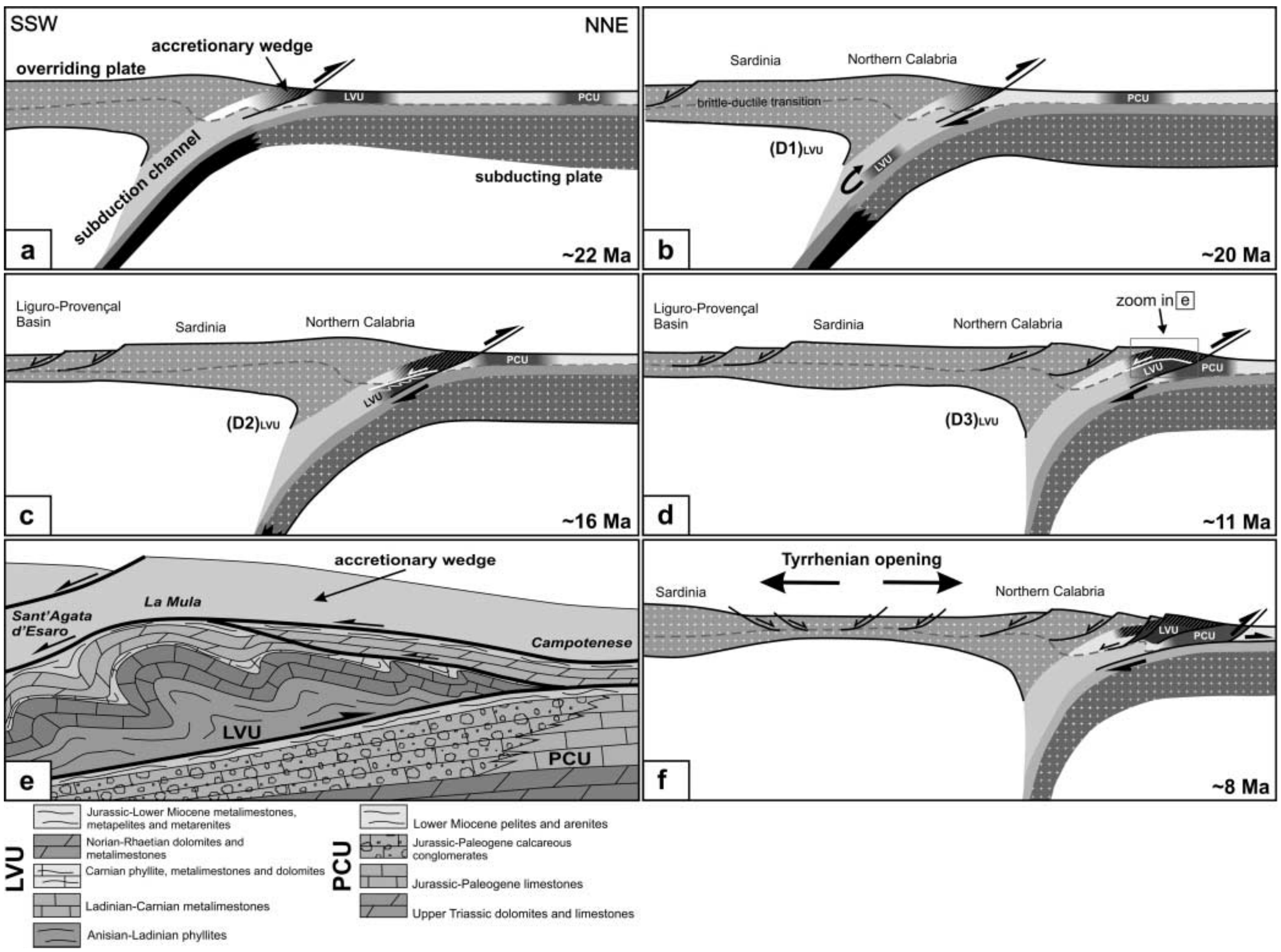

Fig. 14. Schematic reconstruction of the subduction–exhumation history of the Lungro–Verbicaro Unit and Pollino–Ciagola Unit. (**a**–**d**) Early to Middle Miocene stages. (**e**) Enlargement of box in (**d**), showing structures developed during $(D_3)_{LVU}$ deformation stage (see text). (**f**) Late Miocene stage.

perpendicular to the subduction zone, which is then reoriented to z vertical following exhumation along the subduction channel (Fig. 14d). Stretching lineations are aligned NE–SW, consistent with exhumation of footwall HP rocks along the SW-dipping subduction zone. Non-coaxial shearing resulted in D_3-related, SW-facing regional folds. These major structures may be interpreted as antithetic backfolds reflecting relative extension during upward movement of HP rocks (Fig. 14e). The extension is relative (between NE-directed upward motion of deeper, HP footwall rocks and a static hanging wall) in a convergent subduction zone setting (Searle *et al*. 2004).

In the latest stages of synorogenic exhumation the Lungro–Verbicaro Unit overthrust the Pollino–Ciagola Unit, and the latter underwent $(D_1–D_2)_{PCU}$ deformation. Starting from the late Tortonian, the study area experienced post-orogenic extension associated with opening of the Tyrrhenian Sea. Extension caused the formation of basins filled by clastic sediments produced by progressive unroofing of rocks of (1) crystalline basements nappes, (2) ophiolitic nappes, and later (3) the Lungro–Verbicaro Unit (Perrone *et al*. 1973).

Structural development and evolution occurred throughout the synorogenic exhumation history of the Lungro–Verbicaro Unit, starting from high-pressure (blueschist-facies) conditions and continuing during decompression. Our analysis suggests an early development of structural features and their continued modification during exhumation. Although narrow zones of relatively high, localized non-coaxial strain also occur, there is a general good preservation of sedimentary and palaeontological features, a characteristic that has been observed in other continental margin successions involved in underthrusting to considerable depths (Stöckhert 2002).

Synorogenic exhumation and internal deformation of the Lungro–Verbicaro Unit occurred essentially prior to its final emplacement on top of the Pollino–Ciagola Unit, as the latter does not record metamorphic conditions. The contact between the Lungro–Verbicaro Unit and the Pollino–Ciagola Unit is best interpreted as a thrust, which emplaces older on younger rocks, essentially with a footwall flat geometry, and relatively high-pressure units on top of lower-pressure ones.

It can be envisaged that, while experiencing bulk ductile thinning, the Lungro–Verbicaro Unit rock body was also affected by localized non-coaxial deformation induced by 'extensional' shear along the boundary between the subduction channel and the upper plate, accommodating the upward flow of exhuming material. Dominant top-to-the-SW shearing recorded by calc-mylonites could be associated with this process, as well as the development of the spectacular regional backfolds characterizing the Lungro–Verbicaro Unit. Later stages of exhumation took place at upper crustal levels.

The Lungro–Verbicaro Unit is at present bounded by a thrust at the base, whereas extensional tectonic contacts are exposed above it. A major (synorogenic) extensional detachment occurs within the ophiolitic units overlying the Lungro–Verbicaro Unit. In these units, a major, 'normal-sense' pressure break is known to occur and has been recently interpreted as a regional extensional detachment, based on structural data gathered south of our study area (Rossetti *et al*. 2001).

Rapid cooling occurring in the immediate footwall of an extensional detachment could also explain the good preservation of Fe–Mg carpholite in the topmost part of the Lungro–Verbicaro Unit (Jolivet *et al*. 1996). The subsequent evolution of the system occurred at shallow levels, leading to the development of NE-directed thrusts, as it is typical of the Apennines fold and thrust belt, followed by strike-slip and extensional faulting related to recent Tyrrhenian margin evolution (e.g. Butler *et al*. 2004).

Subduction and exhumation rates. A first-order approximation for subduction and exhumation rates can be obtained using the stratigraphic constraints reported above: considering an approximate time span of 15 Ma (from *c*. 22 to *c*. 7.5 Ma) between the younger age of subducted sediments and the older sediments unconformably overlying the exhumed metamorphic rocks, and an approximate maximum burial depth of 40 km, a vertical average subduction or exhumation rate of 2.7 mm a^{-1} is obtained. However, apatite fission-track data may be used to obtain an average exhumation rate between *c*. 10 Ma (cooling through the 110 °C isotherm, i.e. *c*. 4 km depth) and *c*. 7.5 Ma (age of unconformable deposits). This yields an exhumation rate of 1.6 mm a^{-1}, therefore suggesting that vertical rates in excess of 3 mm a^{-1} must have characterized previous exhumation stages. This is in good agreement with regional studies suggesting that the deformation of the Apulian margin did not proceed at a constant rate during the Miocene (Sgrosso 1998). According to Faccenna *et al*. (2001), a substantial lack of subduction occurred between *c*. 15 and 10 Ma, following a period of relatively fast subduction at rates in excess of 30 mm a^{-1}. Integrating this information with the constraints on the timing of deformation and metamorphism discussed above permits an enhanced reconstruction. This involves: (1) beginning of underthrusting of Lungro–Verbicaro Unit rocks at around 22 Ma; (2) peak metamorphism at around 18.5 Ma; (3) fast exhumation within a subduction channel until 15 Ma; (4) slow exhumation during the subduction stasis (15–10 Ma); (5) final exhumation (from a depth of *c*. 4 km) between 10 and 7.5 Ma. For a subducting slab mean dip of 40° as suggested by Faccenna *et al*. (2001) for the early stages of continental subduction, a subduction rate of *c*. 25 mm a^{-1} results from our study, which is compatible with that obtained by Faccenna *et al*. for Calabrian subduction between 22 and 15 Ma. Assuming exhumation occurring at a similar rate (resulting in a vertical rate of 16 mm a^{-1}) within the subduction channel (e.g. Rubatto & Hermann 2001), at 15 Ma a depth of about 9 km is reached. Such rapid exhumation occurred without substantial thermal re-equilibration, as indicated by petrological data (isothermal decompression). Apatite fission-track data indicate that Lungro–Verbicaro Unit rocks lay at a depth of *c*. 4 km at around 10 Ma, therefore providing an average exhumation rate of 0.9 mm a^{-1} during the 15–10 Ma 'quiet' period. The final (i.e. Tortonian) stages of exhumation (between 4 and 0 km) occurred at an average rate of 1.6 mm a^{-1}, as discussed above.

Based on the foregoing discussion, the first stages of deformation of the hinterland part of the Apennine carbonate platform in southern Italy appear to have occurred during the latest stages of rotation of the Corsica–Sardinia block and the coeval opening of the Liguro-Provençal basin (30–15 Ma). These deformation stages were associated with continental subduction immediately following the final consumption of the oceanic lithosphere originally interposed between the Adria and European continental palaeomargins. During the subduction stasis between 15 and 10 Ma, the Lungro–Verbicaro Unit rocks were slowly exhumed within the accretionary complex, being finally emplaced, together with the ophiolitic units, on top of the Pollino–Ciagola Unit. Late Tortonian final exhumation of the Lungro–Verbicaro Unit occurred in the hinterland (i.e. Tyrrhenian) side of the orogen, while active thrusting had moved towards external domains to the east.

Conclusions

Relatively high-pressure (HP) metamorphic assemblages occur in the study area within the metasedimentary succession of the Lungro–Verbicaro Unit, which formed the oceanward (i.e. western) edge of the Apennine carbonate platform of the southern Apennines. This suggests that the early stages of involvement of the Apulia continental margin in convergence-related tectonics were associated with continental subduction. Internal deformation of the HP–LT metamorphic rocks took place by a combination of roughly subduction-perpendicular coaxial shortening and development of narrow zones of localized non-coaxial strain, as well as simple shear-related, regional backfolding. Structural evolution appears to have occurred continuously during decompression, as indicated by index mineral composition associated with progressively younger tectonic fabrics.

Despite their tectonic and metamorphic evolution, the rocks of the Lungro–Verbicaro Unit display a good preservation of sedimentary and palaeontological features. A continuous Triassic to Lower Miocene succession is well exposed in the study area, allowing stratigraphic and facies correlations to be made with the carbonate platform succession included in the structurally lower thrust sheet (Pollino–Ciagola Unit). These correlations indicate that the Lungro–Verbicaro and Pollino–Ciagola Units originated from the innermost part of the Apulia margin, corresponding to the transition between the wide shallow-water domain of the carbonate platform and a branch of the Jurassic Neothetys realm.

According to our new stratigraphic constraints, this portion of the Adria continental palaeomargin was involved in thrusting not earlier than the Aquitanian. Integrating all available information with the known regional framework for the Calabrian Arc, subduction and most of the subsequent exhumation of the rocks of the Lungro–Verbicaro Unit probably occurred, up to late Langhian time (15 Ma), at a rate in excess of 25 mm a^{-1} (vertical rate of 16 mm a^{-1}).

Our results confirm that, similarly to other Mediterranean examples such as Crete (Stöckhert 2002), dominantly carbonate successions may be well preserved even when they have undergone underthrusting to great depths, and thus allow the continental palaeomargin architecture to be reconstructed with some confidence.

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