SAPIENZA
UNIVERSITÀ DI ROMA

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PhD in Earth Sciences – XXXIV Cycle

Tectonic-stratigraphic evolution of the Longobucco Basin (NE Calabria), from Jurassic extension to Cenozoic compression: the Paludi Fm

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Introduction

The Longobucco Basin is part of the Calabria-Peloritani Terrane (CPT hereafter). The CPT is an exotic crustal fragment which, along with the Lesser and Greater Kabylias, Rifian-Betic and Alboran terranes, drifted from its original paleogeographic position and was accreted to the Apennine-Maghrebian chain. The Mesozoic paleogeographic affinity of these crustal fragments is still debated as they are considered as i) part of the southern European margin, ii) part of an independent microplate named AlCaPeKa (although the original meaning given by Boullin, 1986, was altered by subsequent authors) or Mesomediterranean Microplate (Guerrera *et alii*, 1993), and iii) part of the African margin. Reconstructing the geodynamic evolution of these crustal fragments has proved to be a tricky task because these elements were originally located at the boundary between the African and European plates, where the Alps, Apennines, Pyrenees and Betics developed. This boundary was the site of interaction between two opposite subduction systems, whose mode and timing of interplay is still debated in the literature (see Malusà *et alii*, 2018 for a review). The Cenozoic geodynamic puzzle at this plate boundary was further complicated by the opening of backarc basins, causing the fragmentation and drifting of AlCaPeKa from its original paleogeographic position. During their drifting, these allochthonous elements collided with the African plate, becoming part of the Apennine-Maghrebide accretionary wedge. This caused the coexistence and juxtaposition, in the innermost portion of the Chain, of radically different successions, with contrasting paleogeographic affinities (*i.e.,* European *vs* African), and a distinct tectonic/stratigraphic evolution with respect to the backbone of the Apennines. Northern Calabria can be taken as an example of such a complex scenario, as the Costal Chain (Western Calabria) is indeed a part of the Alpine orogen, which drifted and was eventually accreted to the Apennines (Rossetti *et alii*, 2001).

The timing of tectonic and stratigraphic events related to the Cenozoic evolution of the Arc are well constrained in its southern portion (*i.e*., the Peloritani Mts., north-eastern Sicily), following an in-depth study of the upper Eocene-lower Oligocene synorogenic siliciclastic deposits of the Frazzanò Fm., the upper Oligocene-Aquitanian Capo d'Orlando Fm. (Catalano *et alii*, 1996), the post-orogenic Burdigalian deposits of the Calcareniti di Floresta Fm. (Lentini & Carbone, 2014), and the completion of the official geological maps of the area (e.g., Catalano, 2009; Servizio Geologico d'Italia, 2009). The Sicilian authors (Lentini & Carbone, 2014) have highlighted the existence of an older tectonic phase (with respect to the canonical late Miocene thrusting phase of the Central Apennines), named "Balearic phase" (Lentini & Carbone, 2014), corresponding to the stacking of the Calabride Units, and their emplacement on top of the Liguride Complex.

In Northern Calabria, the tectonic evolution of the Coastal Chain (Liguride Complex), and of the so-called Northern Calabrese Units pertaining to the Panormide Complex, have been thoroughly investigated (see Vitale *et alii*, 2019). In contrast, the Cenozoic deposits of the Calabride Complex are only found in the Longobucco Basin and are underinvestigated (but see section on previous studies), which led in our view to an unsatisfactory interpretation of the evolution of the Arc by modern standards.

As demonstrated by the studies of the Sicilian Authors, an in-depth analysis of siliciclastic units, and consequent identification of pre-, syn-, and late-orogenic deposits, is crucial to the understanding of the geodynamic evolution of such troubled areas.

A geological mapping project in the Longobucco Basin proved essential for elucidating the tectonic-sedimentary evolution of northern Calabria, from Jurassic extension to orogenic compression. This paper will focus on the Cenozoic Paludi Fm, a poorly known multifaceted, unconformity-bounded siliciclastic unit, generically interpreted in the literature as an "Apennine flysch" (Magri *et alii*, 1965; Zuffa & De Rosa, 1978; Bonardi *et alii*, 2005).

By understanding the geological significance of the Paludi Fm a re-interpretation of the paleogeographic affinity of the Calabride Complex will be provided, shedding also light on the primordial phases of the Apenninic evolution.

Geological setting of the Calabria Peloritani Terrane

The Calabria Peloritani Arc (CPA) is an arcuate segment connecting the NW-SE trending Southern Apennines and the E-W trending Sicilian Maghrebides. As stated by Bonardi *et alii* (2001) the term CPA should be avoided, as it refers to the morphological feature of this part of the Apennines and not to its geological definition. In this light, authors use the term Calabria-Peloritani Terrane (CPT) indicating an exotic crustal fragment that is geographically part of the Apennines but differs significantly from the surroundings areas.

The CPT (Fig. 1) is bounded toward North and South by two tectonic lines: the Sangineto and Taormina lines, respectively. The nature of these lines is still debated, being both these structures made up of a complex fault system, where subsequent tectonic phases modified the original geometries of the elements. Along the Sangineto line the left-lateral strike-slip component prevails, while the Taormina line is considered to be a thrust fault. To the north-west the continental crust of the CPT gradually passes to the oceanic crust of the Tyrrhenian abyssal plain, while to the South-East it is bounded by the Calabrian subduction complex (Polonia *et alii*, 2011).

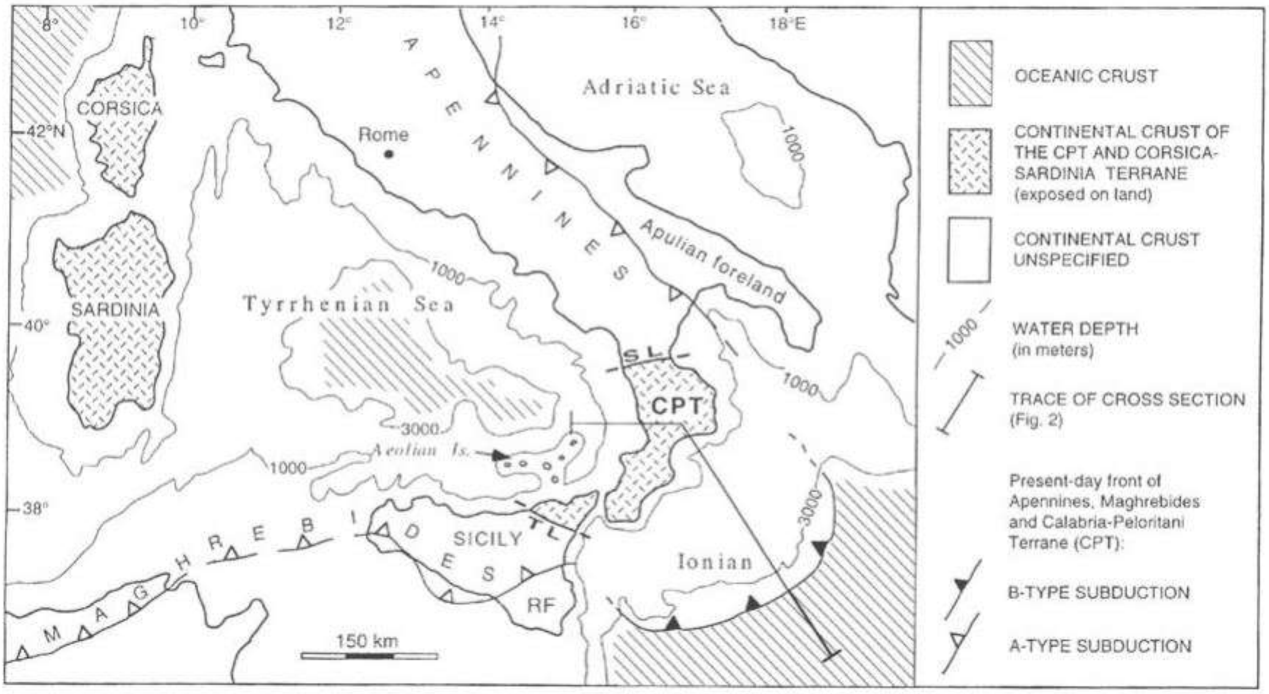

Figure 1: *The Calabria Peloritani Terrane. SL: Sangineto Line; TL: Taormina Line. After Bonardi* et alii *(2001).*

As already mentioned, the CPT is geologically different from the surrounding areas, as it is essentially characterised by an igneous or metamorphic basement with an uneven Meso/Cenozoic sedimentary cover.

Since the work of Ogniben (1973) the CPT has been commonly divided into three main juxtaposed Complexes (Fig. 2), from bottom to top:

- Panormide or Apenninic Complex: Apenninic units (mainly carbonates), locally metamorphosed, outcropping in a tectonic window or push-up transpressive structure in the northern sector of the Terrane;
- Liguride Complex: ophiolite-bearing units pertaining to the Ligurian branch of the Alpine Tethys;
- Calabride Complex: Hercynian basement, locally overlain by a Meso/Cenozoic sedimentary cover.

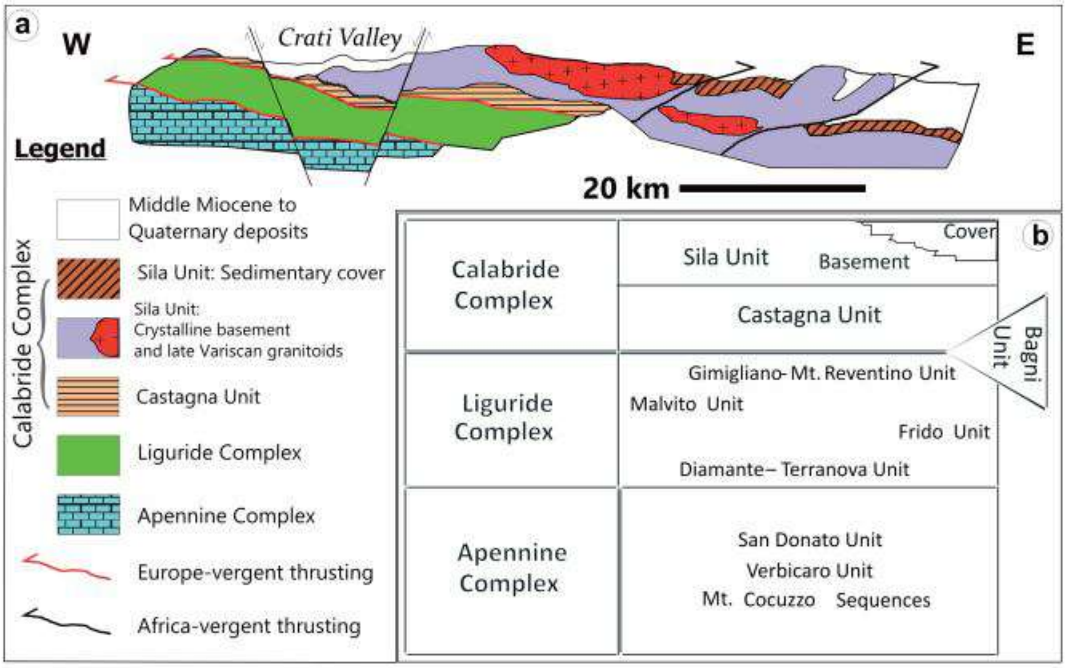

Figure 2: a) Geological cross-section of the northern sector of the CPT (Catena Costiera and Sila), representing the tectonic relationship existing among the main Complexes; b) tectonic units. After Cirrincione et alii (2015).

Each of these Complexes is further divided into tectonic nappes or units (Fig. 2). Despite the general agreement on the broad subdivision summarized above, disputes exist in the literature concerning the internal organisation of said Complexes, with lower-rank tectonic units and, above all, about the paleogeographic affinity of two of the three complexes. In particular, the paleogographic affinity of the crystalline basement, with its varying sedimentary covers, is alternatively considered to be:

- Part of the European continental margin accreted over the African margin together with ocean- derived fragments during the Miocene;

- Part of the African margin involved in the eoalpine orogeny along with the ophiolitic nappes, characterised by Europe- verging compressive structures. This older orogenic wedge would have subsequently been backthrust onto the Apenninic-Maghrebian chain in the Miocene;
- Part of a microcontinent (AlKaPeCa or Mesomediterranean Microplate; *sensu* Guerrera *et alii*, 1993), interposed between the southern European and the northern African margins.

All the paleogeographic reconstructions listed above require drastically different configurations of the Mesozoic/Early Tertiary oceans and continents. In the first hypothesis, only one branch of the Alpine Tethys would be required, and it would be located southward with respect to the CPT. The second interpretation would require a single ocean as well, but it would be located in the North with respect to the CPT. Finally, the third interpretation would require two oceanic branches of the Alpine Tethys, bordering the microplate both northward and southward.

The problem of the paleogeographic affinity of the CPT is not merely a theoretical subject for the scopes of the present thesis, as it deeply impacts the interpretations deriving from field data. At the same time, our new field data will hopefully contribute to the scientific debate. In this light, I will bore the reader with a brief and concise summary of the disputes that can be found in the literature.

AFRICAN MARGIN

All the earlier studies concerning the CPT (see Amodio-Morelli *et alii*, 1976) considered the HP/LT units outcropping in Coastal Chain, with metamorphism dated as Cretaceous/Eocene, as the product of an eo-alpine phase. This interpretation naturally implies that the Costal Chain should be considered as a part of the Alpine Chain, which drifted from its original paleogeographic position and was accreted to the Apennines *s.s.* This is perfectly mirrored in the paper of Amodio-Morelli *et alii* (1976), and in the attached geological map, which clearly subdivides the outcropping tectono-stratigraphic units into “Alpine” and “not Alpine” (plus some *incertae sedis* units). The deformation history of this sector was thus linked to a Europe-vergent phase (subduction directed to the South), connected to the closure of an ocean located between Sardinia (southern European margin) and Calabria (northern Africa margin) (Fig. 3). This interpretation was corroborated by the direct identification in the field (Amodio Morelli *et alii*, 1976) of East-dipping thrust planes in northwestern Sila and by north-westward kinematic indicators both in Sila and Catena Costiera.

This theory was largely abandoned after the middle ‘80s, though with some exceptions. Notably Cirrincione *et alii* (2015) collected structural data along the whole CPT, confirming the presence of westward directed kinematic indicators in the Liguride Complex and relating both the

Eocene metamorphism and the deformative phase to an Alpine-type subduction. Cirrincione and coauthors thus consider the northern CPT alternatively as a part of the African margin, or as an independent microplate (see below).

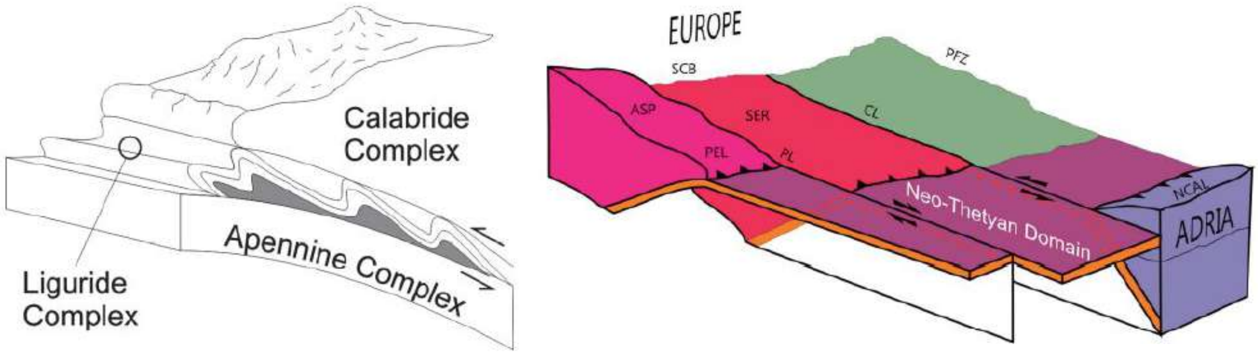

Figure 3: Paleogeographic reconstruction of the CPT in the Cretaceous (NCAL: Northern Calabria). After Cirrincione et alii *(2015).*

EUROPEAN MARGIN

The above interpretation was questioned by several authors (Bouillin, 1986; Knott, 1987) and more recently by Rossetti *et alii* (2001; 2004) and Vitale *et alii* (2019), who are the main proponents of the second paleogeographic hypothesis (*i.e.,* CPT as a part of the European Margin). Knott revised partly the stratigraphy of the Liguride Complex, questioning the division of the Cilento and Frido Units made by previous authors (see Bonardi *et alii*, 1982). These two units were considered to pertain to two distinct basins, bordering the CPT on opposite sides. Knott demonstrated that the Frido Unit is indeed a slightly metamorphosed equivalent of the Crete Nere Fm (pertaining to the Cilento Unit), meaning that the two units represented one sedimentary basin. Furthermore, at the Calabria-Lucanian border, the Author recognised eastward directed thrusts only within the Frido unit. Finally, all the above cited authors correlate the detrital evolution of the Liguride Complex with the progressive denudation of the Calabride Complex. All the above indicates that the Liguride oceanic basin had to be placed eastward with respect to the CPT, the latter being thus part of the Iberian plate. Deformation of this sector of the Tethys would have been consequently the product of a single subduction directed toward the North, taking place since the Late Cretaceous (Fig. 4).

Rossetti and co-authors, similarly, invoked the presence of a single North-directed subduction, questioning the tectonic transport detected along the shear zones affecting both the Liguride and Calabride complexes. Thanks to microstructural analyses, Rossetti *et alii* (2001) argued against the existence of any westward directed compressive structures (*i.e.*, Europe- verging), reinterpreting such elements as low angle normal faults with a top to the west sense of shear.

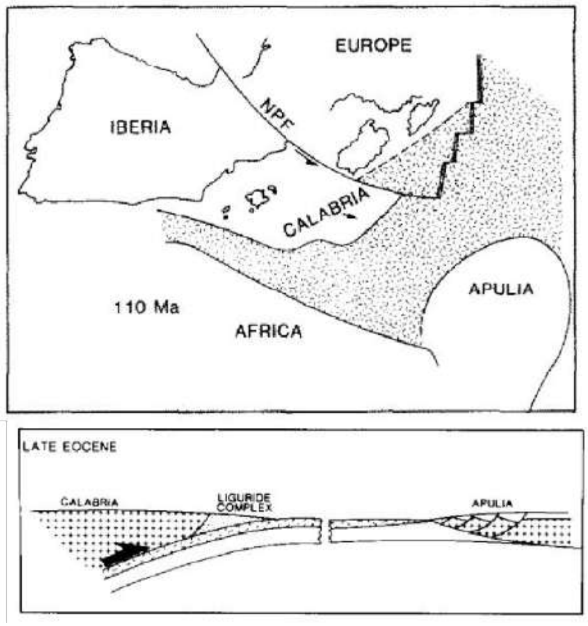

Figure 4: Paleogeographic reconstruction of Knott (1987). The continental unit of Calabria are part of the southern European margin and along with the oceanic derived units (Liguride complex), experienced a southward directed subduction since the Eocene.

For Rossetti *et alii* (2001) these can be interpreted as detachment faults, active at the rear of the orogenic wedge and causing the exhumation of HP/LT metamorphites, while at the front of the wedge thrusting was still active (Fig. 5).

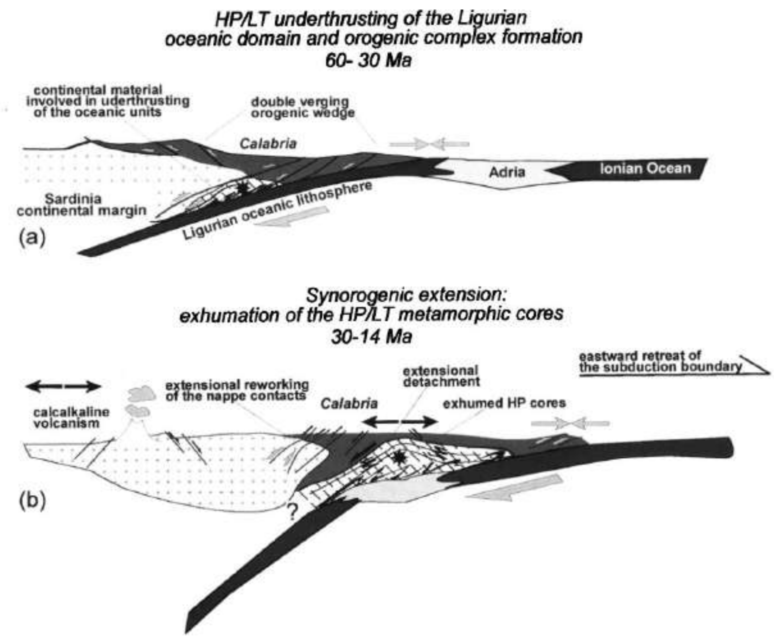

Figure 5: The Calabrian units were deformed by compressive structures directed toward the West since the Cretaceous. Such structures were later reactivated as extensional detachments with a top to the West sense of shear, due to collapse of the orogenic wedge. This extensional tectonic phase led to the exhumation of deep seated HP/LT units in the inner part of the wedge, while compression was still active in the external parts (after Rossetti et alii*, 2001).*

Within the framework of this interpretation, Vignaroli *et alii* (2012), studied the evolution of the eastern sector of the Sila Massif (Longobucco basin) concluding that, while in this sector of the CPT a thrusting phase was active, the innermost part of the chain (Coastal Chain) was being extended. This implies that the deformation of the CPT was a continuous process since the Paleocene/Eocene, migrating toward the East.

More recently Vitale *et alii* (2019), in their review on the Liguride Complex, discussed the petrologic affinity of the ophiolithic suites exposed in Basilicata and Northern Calabria with those of Northern Apennines. According to these authors, the two ophiolitic successions are largely comparable and can be considered as part of the same ocean. The authors also reviewed the tectonic vergences acquired in the field through the years by various authors for both the Liguride and Calabride complexes, highlighting the discrepancies present in the literature (*i.e.,* Europe vs Africa vergence). Vitale *et alii* (2019) show how the kinematic indicators are actually discordant, showing opposite trends, even within the same outcrop. They, however, propose as the most reliable kinematic indicators those indicating a top-to-the-East sense of shear (*i.e.* Africa vergency). All the above would indicate in their view that a single W-directed subduction was active since the Paleocene-Eocene (Fig. 6).

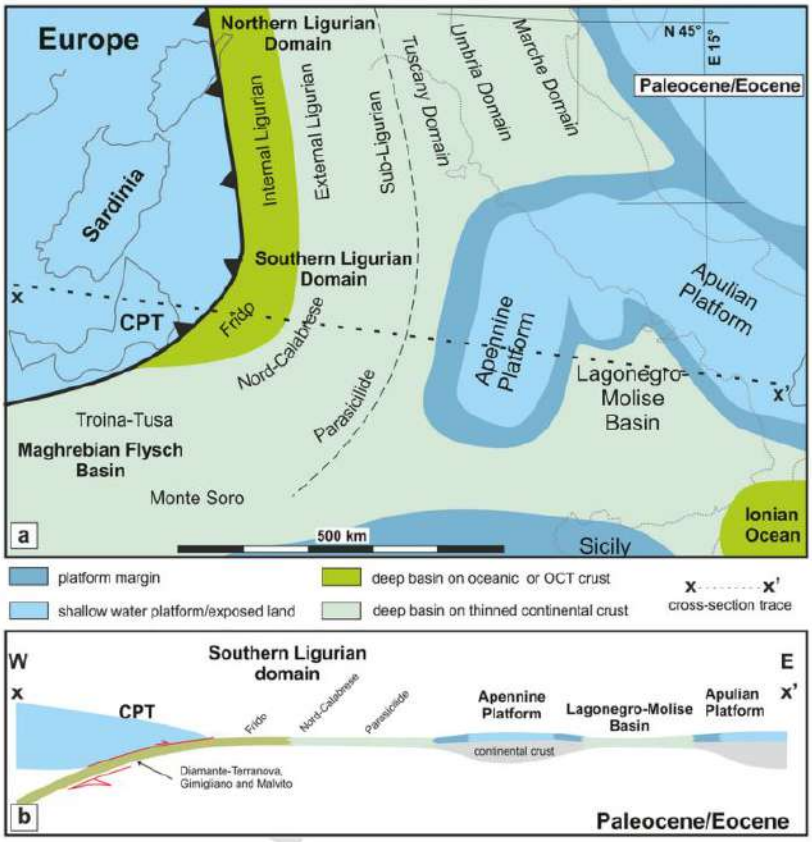

Figure 6: Paleogeographic reconstruction made by Vitale et alii *(2019). Note that the Authors believe that the subduction dipping toward the West since the Paleocene.*

It must be remarked that both Rossetti *et alii* (2001) and Vitale *et alii* (2019) consider the W-directed subduction active in Calabria as being the continuation of the Alpine subduction active in Alpine Corsica, which is however considered as E-directed in virtually all the existing literature.

ALKAPECA

The third hypothesis considers the CPT, along with Alboran, Rif and Kabylides, as a part of a small microplate bordered by two distinct branches of the Ligurian Tethys. This microplate has been improperly named AlKaPeCa, citing the paper of Bouillin *et alii* (1986), where this acronym was coined, which however considered such elements as part of the southern European margin until the upper Oligocene and as part of an independent microplate accreted to the Apennines only after that time (Fig. 7). In spite of this, the term AlKaPeCa is today widely used in the literature (Handy *et alii*, 2010).

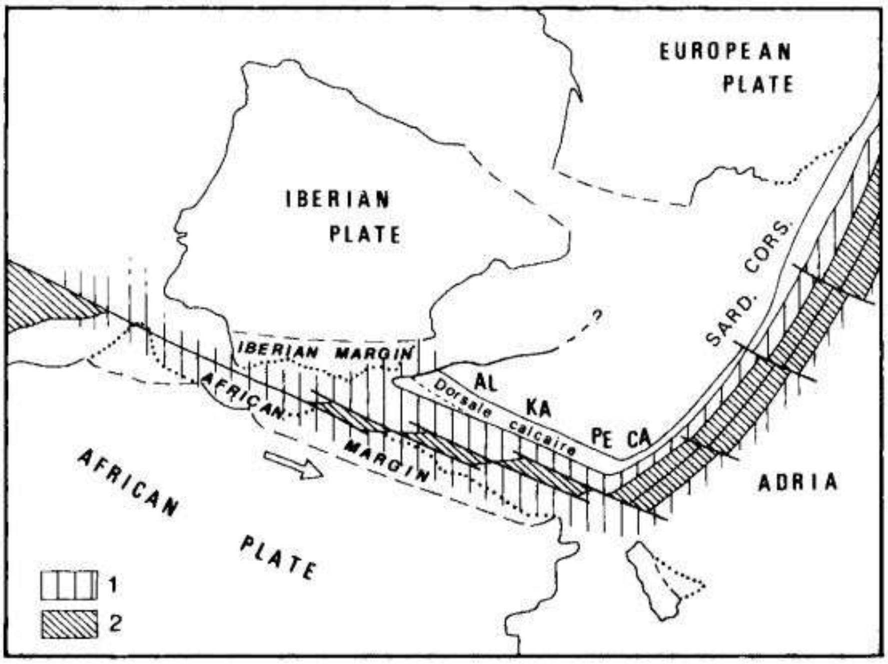

Figure 7: Paleogeographic reconstruction of Bouillin et alii *(1986). Please note that in the original definition of the AlKaPeCa domain, the crustal elements were located, in the Mesozoic, in the southern margin of the Iberian Plate.*

Guerrera *et alii* (1993) (Fig. 8) proposed the term Mesomediterranean microplate to indicate this double ocean-bounded element. The concept of an independent microplate was developed essentially through the revision of the tectonic-stratigraphic evolution of the abovementioned areas by different authors (Alvarez, 1991; Guerrera *et alii*, 1993), who highlighted key common features concerning the age of the main tectonic events, the position within the Apenninic-Maghrebian orogen, and the stratigraphy.

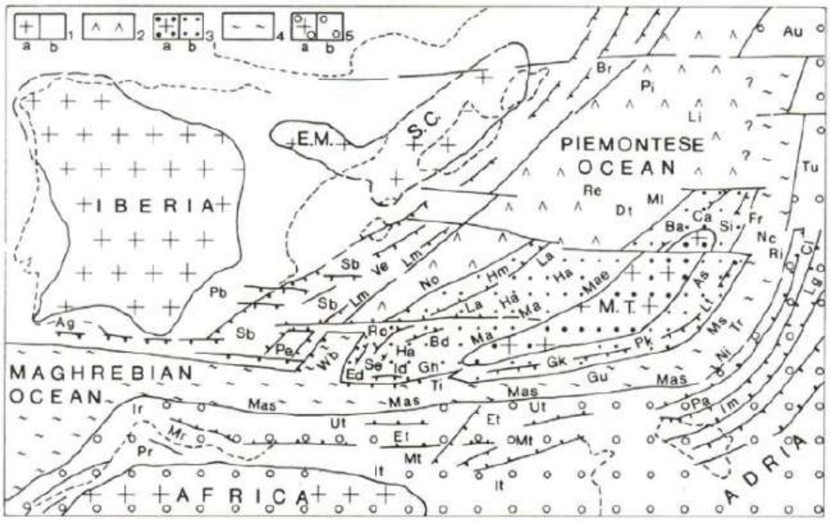

Figure 8: schematic configuration of the Mesomediterranean microplate by Guerrera et alii *(1993)*

Similarly, recent papers concerning the geodynamic evolution of the western Tethys, based essentially on geophysical constraints (Handy *et alii*, 2010; Van Hinsbergen *et alii*, 2014), include this crustal fragment in their paleogeographic reconstructions, without discussing in depth this decision, and arbitrarily only quoting the literature supporting such interpretation.

If we re-read critically the literature, it appears that the concept of AlKaPeCa suddenly became at some point a "widely accepted" theory in the evolution of the Western Mediterranean area, which it was not, despite the major open questions still remaining, concerning:

- Age of spreading of the two oceanic branches;
- Paleogeographic affinity of the individual elements of AlKaPeCa before the opening of the oceans (African vs European);
- Where are, today, these two oceanic branches?

An exception to the above is the study by Cello *et alii* (1996) (Fig. 9) in Northern Calabria, which is based on field data and not on a revision of the existing literature. The Authors consider the CPT as a part of a Jurassic-Paleogene independent microplate, an hypothesis put forward in an attempt to address the issue of the presence of East-verging vs West-verging structures in the ophiolite- bearing Units of the Liguride Complex. Cello and co-authors recognised both Africa-verging (in the Frido Unit of Knott) and Europe- verging (in the Diamante-Terranova Unit) structures, thus considering these two Units as pertaining to two distinct oceanic basins bordering the microplate. In the interpretation of Cello and co-authors, the thrusting event was linked with a southward directed subduction started in the Cretaceous and generating a double-vergent orogen. This is clearly in contrast with data pertaining the thrusting phase in the Longobucco Basin, that must be post-Eocene in age (this Thesis).

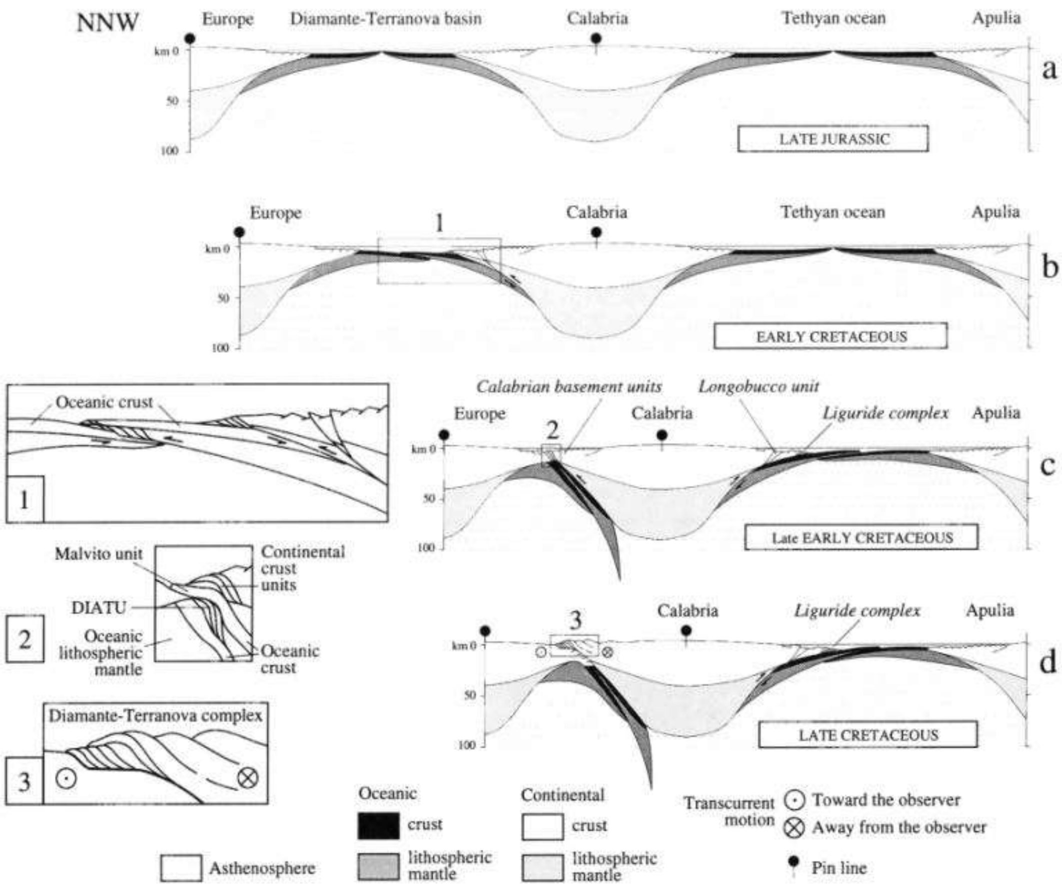

Figure 9: Paleogeographic reconstruction of Cello et alii *(1996) representing the CPT as bordered by two distinct oceanic branches and characterised by double vergent thrusts since the Cretaceous.*

A comprehension of the geodynamics of the CPT is further complicated by the fact that the previously described schematic subdivision in major Complexes does not perfectly mirror the internal architecture of the CPT, which displays a number of differences along its longitudinal extent. The dissimilarities, summed up by Bonardi *et alii* (2001), concerning the northern and southern sectors and are:

- Absence of Panormide and Liguride Complexes in southern Calabria and north-eastern Sicily;
- Both Europe- and Africa- verging compressive structures in the North, solely Africa-verging structures in the South (this point is questioned by many authors, *e.g.*, Rossetti *et alii*, 2001);
- HP/LT metamorphic units found in northern Calabria only;

- Oligocene metamorphic event in southern sector only;
- Continuous sedimentation from Triassic to Oligocene in the southern sector (actually, this consideration is not true, as only fragmentary outcrops in the Longi-Taormina Unit would document this).

Regarding the abovementioned discrepancies, since the '80s some authors (Bonardi, 1980; Tortorici, 1982; Scandone, 1982) proposed that the CPT could be the product of the amalgamation of two terranes (named the Northern and Southern Calabria/Peloritani Subterranes). The contact between these two sectors has not been recognised in the field thus far. Bonardi *et alii* (1994) tentatively place such crustal suture in the northern Serre Massif. Although this subdivision has been questioned over time, it is still commonly used to highlight differences among these two sectors of the CPT.

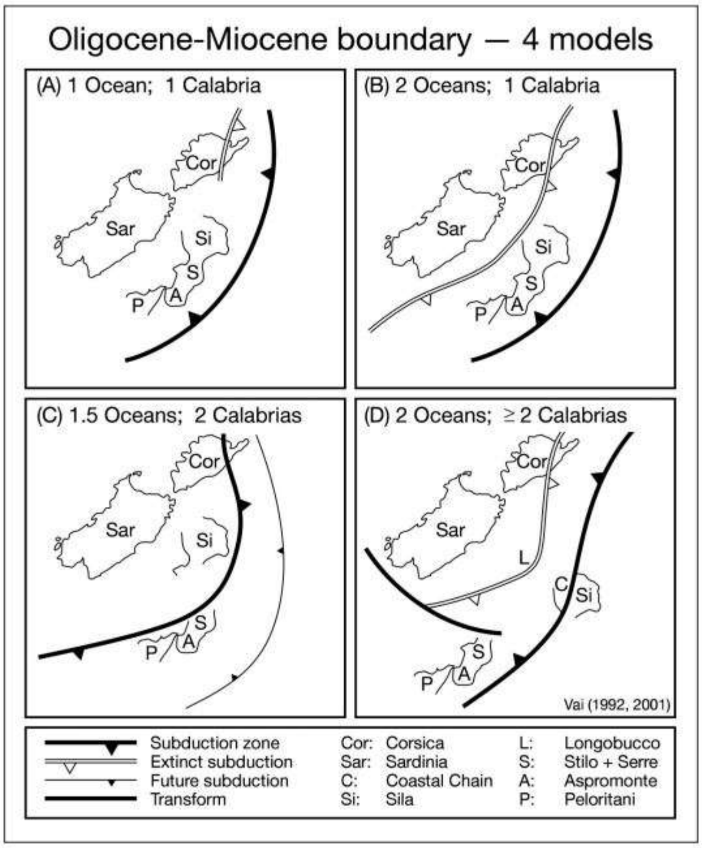

Figure 10: Tentative paleogeographic scenarios proposed by Alvarez & Shimabukuro (2009).

Clearly, this hypothesis further complicates the paleogeographic reconstructions, implying the presence of a transform fault separating the two sub-terranes (Fig. 3 and Fig. 10).

The post-Oligocene evolution of the CPT is conversely well constrained and accepted, and is related to the eastward retreat of the Apenninic slab due to the opening of back-arc basins (Carminati *et alii*, 2012), paired with fragmentation and dispersal of crustal fragments. Such elements were described by Alvarez (1976) as "microplates", as they acted independently from the major plates (Europa-Africa). In this light Bouillin *et alii* (1986) coined the name AlKaPeCa, to describe the common post-Eocene history of these elements. The concept of "microplate", as already explained, was then implicitly stretched to the Mesozoic.

The Oligo/Miocene evolution can be divided into two phases related to the diachronous opening of the Liguro-Provencal and Tyrrhenian basins respectively (Fig. 11).

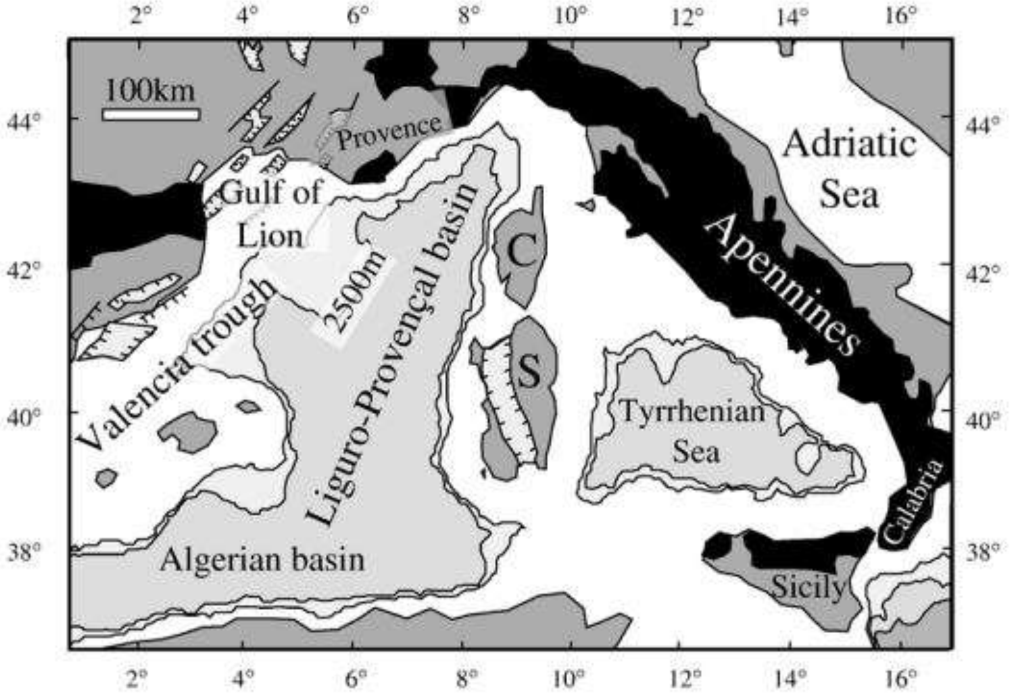

Figure 11: Current configuration of the Western Mediterranean by Gattacceca et alii*, 2007.*

First phase → Opening of the Liguro Provençal basin and rotation of the Sardinia-Corsica-Calabria Block

The paleogeographic position of the Sardinia-Corsica Block is well constrained: this element was located in the southern European margin, attached to Provence. The CPT was surely close to this Block, and these two areas acted and moved as one starting in the Oligocene and until the opening of the Tyrrhenian Basin. Since the late Oligocene, syn-rift deposits both in the Catalan-Provençal and Sardinian margins record the onset of extensional tectonics related to the opening of the Liguro-Provençal Basin. In the early Miocene (20.5 Ma; Gattacceca *et alii*, 2007) the Sardinia-Corsica block drifted and started to rotate CCW by 45°, until 15 Ma. This phase corresponds to litospheric break-up and sea-floor spreading in the Liguro-Provençal Basin. A further 10° rotation could have possibly occurred between 21.5 and 20.5 Ma (since the end of the rifting phase), resulting in a total CCW rotation of about 55° for the Corsica-Sardinia Block. In this time span, evidence of compressive tectonics is detected in the CPT (this Thesis). The CPT collided in the Aquitanian with the Adria-Africa paleomargin (with the Panormide Complex to the North and Sicilide Units to the South of the CPT, respectively). This tectonic phase is well constrained by the dating of syn- late- and post-orogenic siliciclastic deposits in the CPT. In particular, in southern Calabria and north-eastern Sicily, the Stilo-Capo D'Orlando Fm, whose younger elements are dated as Burdigalian, seals the compressive structures marking the end of this orogenic phase.

Second phase → Opening of the Tyrrhenian Basin

After a 5 Ma long tectonic and volcanic quiescence, crustal extensional jumped east of the Sardinia-Corsica block with the opening of the Tyrrhenian basin. The oldest evidence of rift-related tectonics in the Northern Tyrrhenian basin can be ascribed to the Tortonian (8 Ma), while oceanic spreading occurred in two distinct events: at 4 Ma in the Vavilov Basin and after 2 Ma in the Marsili basin (South). In this phase the CPT drifted away from the Sardinia-Corsica block reaching its current position. The presence of oceanic crust in the Ionian basin is inferred to have driven the very fast south-eastern migration of the CPT. During this younger phase the CPT, deformed in the previous phase, was passively transported at the rear of the newly formed chain. Major transcurrent faults deformed the compressive ones, and new thrust sheets developed at the front of the chain.

Starting in the Tortonian in the Tyrrhenian side of the CPT, high angle normal faults started to develop due to back-arc extension, generating subsiding basins such as the Paola, Amantea, and Gioia basins (Ghisetti, 1979).

The Sila Massif and Longobucco Basin

The Longobucco Basin is located in north-eastern Calabria in the Sila Massif (Fig. 12)

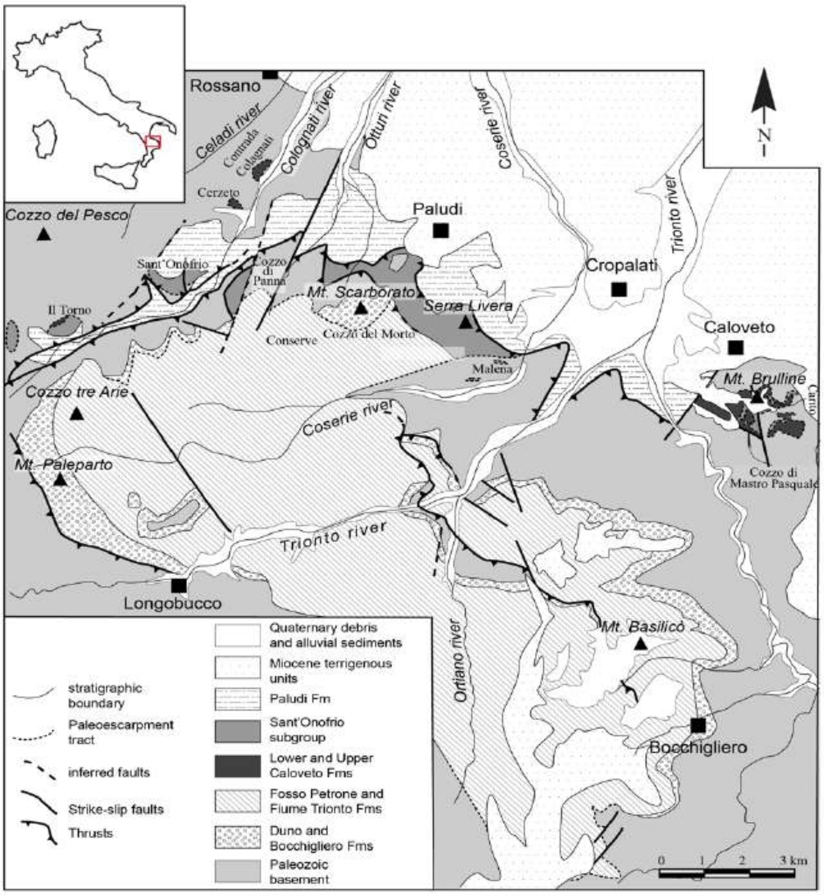

Figure 12: Schematic geological map of the Longobucco Basin.

The Sila Massif shows a nearly complete late Hercynian crustal section composed from bottom to top by:

- High-grade metamorphic rocks (lower to middle crust) outcropping to the West and represented by migmatitic paragneiss, marbles and metabasites. Age of metamorphism ranges from 305 to 296 Ma (Festa *et alii*, 2006).
- Late Hercynian granitoids (middle crust) forming a huge batholite (600 km^2). The composition of such granitoids is represented essentially by monzogranite and tonalite, cut by mafic and leucogranitic dykes. The age of intrusion, assessed by Ayuso *et alii* (1994) by means of Ar/Ar thermochronology, spans from 293 to 289 Ma.
- Low-grade and very low-grade metamorphic rocks (upper crust) represented by metapelite, metarenite, metavolcanic rocks, rare marbles pertaining to the Mandatoriccio Complex (low-grade) and essentially phyllites pertaining to the Bocchigliero complex

(very low-grade) (Langone *et alii*, 2010). The Meso/Cenozoic succession rests unconformably both on the igneous and very low/low-grade metamorphic Hercynian basement. The Mesozoic succession records the evolution of a continental margin affected by extensional tectonics related to the opening of the Western Tethys. This phase dismembered the submarine paleotopography in structural highs and lows. Within this frame two stratigraphic groups were recognised (Santantonio & Teale, 1987) describing the successions pertaining to hangingwall and footwall blocks of Jurassic faults: the Longobucco and Caloveto groups, respectively (Fig. 13).

The sedimentary cycle starts in the upper Triassic with continental deposits resembling the Verrucano facies of the Southern Alps, which unconformably cover the Hercynian basement. These deposits have been named Torrente Duno Fm by Young *et alii* (1986), while Bouillin *et alii* (1988) considered this lithostratigraphic unit as two distinct formations:

- The Monte Paleparto Fm: polymictic conglomerates with clasts represented essentially by metamorphic quartz pebbles in a reddish/purple matrix;
- Torrente Duno Fm: alternations of sandstone, often cross- bedded, siltite and clay, with trace fossils and abundant plant remains.

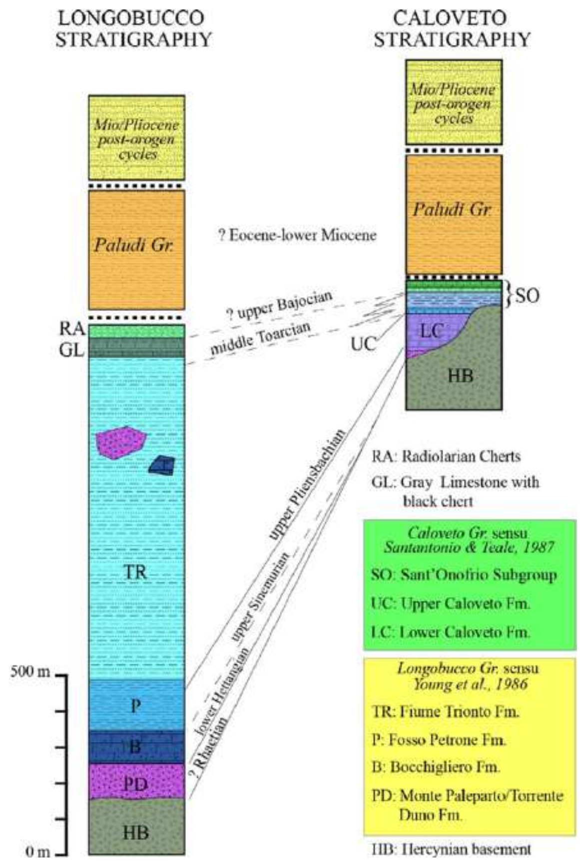

Figure 13: Stratigraphy of the Longobucco Basin and main Jurassic tectono-sedimentary events (Santantonio & Fabbi, 2020).

These two formations can be interpreted as the product of deposition in a braided fluvial system and in an alluvial plain. The Monte Paleparto Fm does not outcrop evenly across the basin, so that the Torrente Duno Fm is commonly seen resting directly on the Hercynian basement. This can be explained assuming that the basal unit essentially filled pre-existing morphological depressions of the Hercynian substrate. The two continental units display thickness variations across the basin, ranging from 10 to 100 metres. Their age is not well constrained but is considered to be Rhaetian (?) *p.p.* – Hettangian *p.p.* Contrary to other sectors of the Western Tethys, where Triassic deposits are up to some km in thickness, the Longobucco basin represented a non-subsiding morphologic high in the basement (Santantonio *et alii*, 2016), where accommodation space was scarce.

Overlying the continental red beds, the Bocchigliero Fm is made of dark limestones with brachiopods, bivalves and gastropods, documenting a marine transgression and the onset of a mixed carbonate/siliciclastic shelf. This shelf displays dominant subtidal facies, represented by muddy bioturbated levels, rich in oncoids, hybrid calcarenites with wave ripples, and cross-bedded ooid shoals. The passage from the Duno to the Bocchigliero Fm is transitional, as the lower part of the Bocchigliero Fm is characterised by common sandstone interbeds, rich in plant debris, that gradually decrease upsection. The presence of such intervals testifies to the presence of an emergent continent backing the shelf. The unit is 60 meter thick, and its age is Hettangian *p.p.* - Sinemurian. Both the Torrente Duno and Bocchigliero Fms display small syn-sedimentary faults predating the first major extensional tectonic phase starting in the earliest Pliensbachian, when the deposition of the marls of the Fosso Petrone Fm, characterised by nektonic faunas (cephalopods), marks the deepening of the basin. This paleoenvironmental turnover is ascribed to tectonics, which generated a complex submarine topography where basins were bordered by rift faults. Such margins have been identified in the field and mapped in recent years and occur in the form of steep submarine paleoescarpments onlapped by huge, laterally continuous clastic bodies, interpreted as the result of repeated episodes of catastrophic collapse of the footwall blocks. These bodies have been dated as latest Sinemurian/earliest Pliensbachian by means of ammonite and belemnite biostratigraphy (Santantonio *et alii*, 2016; Innamorati & Santantonio, 2018).

Basinward, intercalated within the marls, conglomeratic intervals are common, representing the distal expression of such mass transport deposits (Fig. 14). Further evidence of this extensional phase is the presence of neptunian dykes densely cutting the Hercynian basement.

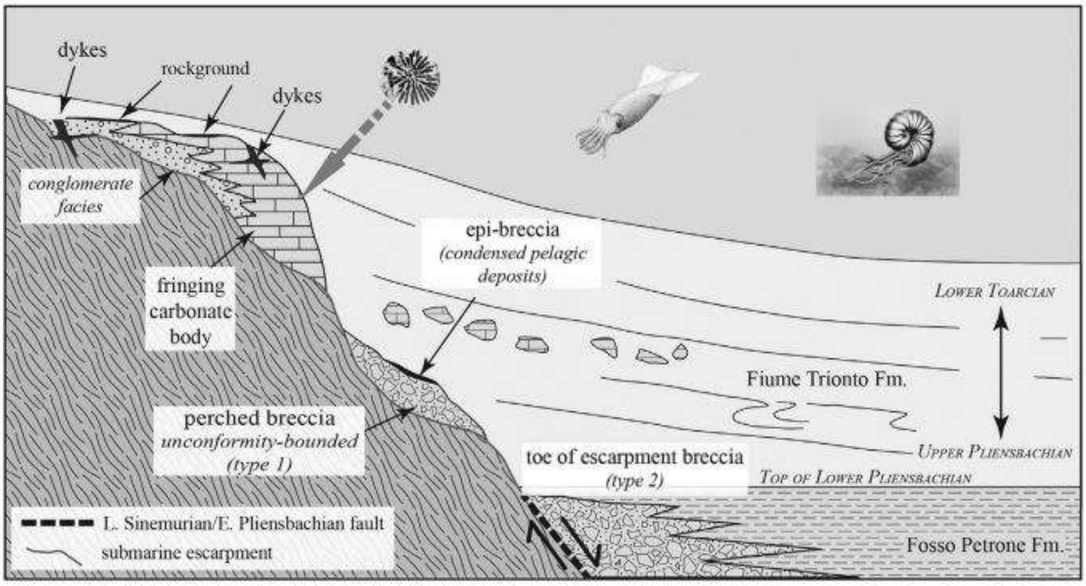

Figure 14: schematic configuration of the basin margin succession. Note the presence of distinct types of clastic deposits, resting either on the escarpment or at its toe. Both the basement and the fringing carbonate bodies are cut by neptunian dykes. The Trionto formation is characterised by olitoliths and slumps. From Innamorati & Santantonio (2018).

Tectonic subsidence characterised hangingwall blocks throughout the Pliensbachian, as documented by the siliciclastic turbidites of the Trionto Fm. This unit is 1 km in thickness and in its upper part (latest Pliensbachian-early Toarcian) bears huge (> 100 m across) olistoliths made of basement rocks or of the pre-rift sedimentary units. Besides megaclasts, the unit displays also frequent slumps, interpreted as a sign of sea-bottom instability which was triggered by the rejuvenation of basin margins due to a second tectonic phase. A Toarcian extensional phase is exceptionally well documented through mappable synsedimentary faults and neptunian dykes observed in the Caloveto sector (see below).

The evolution of the Longobucco Group was considered by previous authors (Young *et alii*, 1986) to end with the top of the Trionto Fm. Recently, Santantonio & Fabbi (2020) (Fig. 15) highlighted that sedimentation continues upward with gray pelagic Posidonia limestone with black chert bearing carbonate turbidites with clasts from a shallow-water platform source. This unit can reach up to 200 m in thickness and its age is Toarcian *p.p.* – Bajocian *p.p.*, and it passes upward to radiolarian cherts (Bajocian *p.p.* - Oxfordian). These two lithostratigraphic units were attributed by Bouillin *et alii* (1988) to a distinct tectonic-stratigraphic unit named “Unità di Sant’Antonio” and considered to belong to a separate basin.

Thanks to recent field evidence, Santantonio & Fabbi (2020) demonstrated that the upper part of the *Zoophycos* marls (Sant'Onofrio sub-group, see below) is locally replaced by the gray limestone with black chert, thus demonstrating a likely physical connection between the Longobucco and Caloveto Groups (Fig. 15).

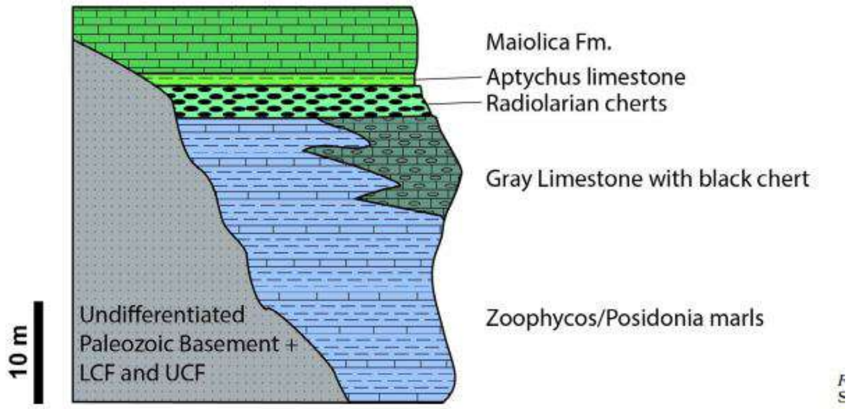

Figure 15: Schematic representation of the Sant'Onofrio sub-group by Santantonio & Fabbi (2020).

As mentioned above, around the Sinemurian/Pliensbachian boundary the rift-related extensional tectonics dismembered the seafloor. This event marks also the beginning of sedimentation of the Caloveto Group, which characterises footwall blocks. The Caloveto area represented a rocky island or a promontory (Santantonio & Fabbi, 2020) in the Pliensbachian. Sedimentation started with the Lower Caloveto Fm, represented by shallow-water carbonate bodies fringing the exhumed Hercynian basement. These bodies were characterised by silicliclast-rich limestones, with oncoids, gastropods, bivalves, rare oolitic shoals and coral assemblages. At the Pliensbachian/Toarcian boundary, the second tectonic phase caused the drowning of the shallow-water limestones, as testified by normal faults and neptunian dykes. Sedimentation then continued with the Upper Caloveto Fm, a typical Rosso Ammonitico facies, represented by thinly bedded reddish marls and nodular limestones. This unit ranges in thickness from few decimetres to 10 metres and its age is lower Toarcian *p.p.* – upper Toarcian *p.p*. The drowning can be either represented by a drowning succession or by a drowning unconformity (*sensu* Marino & Santantonio, 2010). The former (Santantonio & Fabbi, 2020) is represented by a brachiopod coquina, grading into a packstone rich in gastropods, ammonites, coated grains and sponge spicules. This interval is followed by red encrinites. Wherever the drowning unconformity is present, the top of the Lower Caloveto formation is moulded to form a low-relief mineralised hardground surface, and the Upper Caloveto Fm directly rests on the shallow-water bodies. Sedimentation in the Caloveto Group continued with the Sant'Onofrio subgroup, that resembles the classic middle/upper Jurassic Tethyan successions, despite the common occurrence of siliciclastic debris. Due to being predated by a tectonic phase, the base of the sub-group is locally an angular unconformity.

Sedimentation starts with *Zoophycos* marls (Toarcian *p.p.* – lower Bajocian) characterised by reddish bioturbated micaceous marls and marly limestones rich in thin shelled bivalves (*Bositra* and *Lentilla*), rare ammonites and belemnites. Locally incipient hardgrounds are testified by dark mineralizations. As already mentioned, the *Zoophycos* marls can either pass upward to Gray Limestone with black chert or to the Radiolarian chert (Bajocian *p.p.* – Oxfordian). This can be explained by inferring that the limestones, typical of the depocentral areas of the basin, pinch out toward the structural highs/basin margins where they locally replace the top of the *Zoophycos* marls. Sedimentation continues with Aptychus limestone, characterised by purple-color thinly bedded marly limestones and cherty limestone with abundant aptychi.

The age of the unit is Kimmeridgian – Tithonian *p.p.* The Sant'Onofrio sub-group ends with thinly bedded white limestone with chert pertaining to the Maiolica Formation (Tithonian *p.p.* – Hauterivian), containing radiolarians and calpionellids. The base of the formation is characterised by debrites, with clasts pertaining to the basement and to the older Mesozoic units. The uppermost part of the Formation is represented by bioturbated marls and marly limestones.

It must be noted that as of today it is not known in northern Calabria a Jurassic shelf that had to feed the carbonate turbidites rich in resedimented platform-derived grains typical of the middle/upper Jurassic unit of the Longobucco basin. Similarly, the shelf that had to source the Toarcian turbidites of the Trionto Fm is not found in outcrop, probably having been eroded due to Quaternary or older uplift.

The Mesozoic succession (both the Longobucco and the Caloveto Groups) and the Hercynian basement are both unconformably covered by the Cenozoic Paludi Fm, which is the object of this thesis. Although the age of this unit is not yet robustly constrained (see below), a hiatus of at least 80 Ma occurs between the Hauterivian Maiolica-like pelagites and the Paludi Fm.

The Paludi formation is an extremely varied lithostratigraphic unit interpreted by previous authors (Zuffa & De Rosa, 1978) as a witness of the dismantling of a rising orogen. The Paludi Fm is made of quite an assortment of different lithologies, from hemipelagic marls to turbidites, to megaclastic deposits, documenting sedimentation in a basin which received material eroded from its tectonically active margins.

The Paludi Fm, the Mesozoic Units and the basement are cut by NW-SE striking thrusts verging toward the North-East. Thrusts are arranged forming an imbricate structure, with the innermost structures dipping more than the external ones. Four tectonic units, more or less continuous along the basin, were recognised, and are generally characterised by hangingwall anticlines, with an overturned to sub-vertical forelimb, paired with very tight footwall synclines with an overturned to subvertical backlimb. Both the anticlines and synclines host dense parasitic folds. All the above-mentioned Units are unconformably overlain by late-orogenic siliciclastic units, ranging in age from Tortonian to Pliocene (Muto *et alii*, 2014; Roda, 1967). These late-orogenic units are not cut by any compressive structures but rather seal them, and they gently dip toward the North/Northeast forming a monocline. These Miocene units are interpreted to represent wedge-top deposits of the Calabrian foredeep-basin system. This system is divided in three main depocenters, the Rossano, Cirò and Crotone basins, whose infills differ slightly in age and composition. In the Longobucco Basin the late orogenic deposits outcrop at low elevation in the eastern sector only, except for the Bocchigliero area, where they outcrop at ~800 m a.s.l. The Rossano Basin infill shows a transgressive trend, and the sedimentary cycle starts with alluvial and fan-delta conglomerates representing the Conglomerati Irregolari Fm. The succession continues upward with fossiliferous sandstones rich in *Clypeaster*, *Pecten* and *Ostrea* characterising a nearshore area, forming a unit named “Arenaceo-conglomeratica Fm”. This unit is then followed by the “Argilloso-marnosa Fm”, characterised by marls, clays, and thin sandstones beds at it base, interpreted as an outer shelf/slope deposit.

A recently published paper by Siravo *et alii* (2022) analyses paleomagnetic data from the Longobucco Basin. The Authors estimate comparable rotations from the various Mesozoic lithostratigraphic units (in particular from Rosso Ammonitico-like facies and the circa 25 Ma younger Aptychus limestone), amounting to 160° CCW with respect to stable Europe. The Authors, consider the Longobucco Basin as attached to Sardinia until the opening of the Tyrrhenian Sea. Considering that, since early Eocene times, the Sardinia-Corsica block underwent a 90° CCW rotation, Siravo *et alii* (2022) conclude that the Longobucco Basin underwent a main rotational event of about 70° CCW between the Cretaceous and Eocene, linking the whole event with an Albian phase, due to transcurrent tectonism related to left-lateral shear between Africa and Europe. In this frame and in their reasoning the Authors never consider the Paludi Fm (although they state they did sample it) to further constrain the age of rotation.

The Paludi Fm, despite its debated age, is a key unit for the following reasons:

- If it is considered as Aquitanian (Bonardi *et alii*, 2005), it must have registered the 45° CCW rotation event occurring after the drifting stage of the Liguro-Provencal Ocean (Gattacceca *et alii*, 2007);
- If it is considered as Eocene (as proposed in this work), it must have registered part of the abovementioned rotation of 90°CCW of the southern Iberian margin (Advokaat *et alii*, 2014) and,

furthermore, a different amount of rotation with respect to the Mesozoic units of the Longobucco Basin.

Finally, Siravo *et alii* (2022) use their data to disprove the existence of the AlKaPeCa microplate as a distinct terrane separated from the southern European margin by an oceanic branch of the Alpine Tethys, as no rotation related to oceanic spreading and complete terrane detachment is documented by their data. They propose that the AlKaPeCa domain was instead a continental ribbon separated from Europe by thinned continental crust.

Objectives

The Longobucco basin is located in one of the innermost sectors of the Apennines and, as highlighted by recent studies on the exhumation history of the basin (Vignaroli *et alii*, 2012), it documents an older and different history with respect to the backbone of the Apennine chain. It is furthermore part of one of the most geologically problematic areas of our peninsula: the CPT.

As recalled in the Geological Setting, unanswered questions about the CPT are plenty, in particular concerning its paleogeographic affinity, its Cenozoic tectonic evolution and its internal organization. Sicilian Authors (see Lentini & Carbone, 2014) highlighted a polyphase deformation history for the SCPT, which certainly makes the comprehension of this sector of the Apennines a challenging task.

As demonstrated by the studies of these Authors (Lentini & Carbone, 2014), an in-depth analysis of siliciclastic units, and consequent identification of pre-, syn-, and late-orogenic deposits, is crucial to the understanding of the geodynamic evolution of such troubled regions as the CPT. This work has never been carried out in the NCPT, as the only exposed Meso/Cenozoic basin is the Longobucco basin.

In this light, the present PhD thesis focuses on the Cenozoic portion of the Longobucco basin and, in particular, on the geological significance of the Paludi Fm.

In the literature two different interpretations exists for the Cenozoic evolution of our basin (as for the whole NCPT), that can be summarised as follows: “Alps” *vs* “Apennines” (Dubois, 1970 *vs* Bonardi *et alii*, 2005 *vs* Vignaroli *et alii*, 2005).

These conflicting views are due to the fact that the age of thrusting in the Longobucco Basin is poorly constrained, being generically considered as “post-Paludi” and “pre-Tortonian”. At the same time, the age of the Paludi Fm is itself currently debated (Cretaceous *vs* Eocene *vs* Aquitanian, see chapter “Previous studies”), which therefore led to widely different interpretations of the geological significance of this unit (Bonardi *et alii*, 2005 *vs* Dubois, 1970; Zuffa & De Rosa, 1978). While sedimentation of the Paludi Fm is intimately linked with syn-orogenic dynamics, few studies have dealt with the sedimentology, general geometries, relationships with the older units, and biostratigraphy of the Paludi Fm.

To further complicate the situation, as highlighted in the Geological Setting section, two main stratigraphic gaps exist, which leaves much of the Longobucco Basin history undocumented:

- between the Mesozoic succession and the Paludi Fm: ranging between 80 Ma *vs* 107 Ma as a function of the considered age of the Paludi Fm (Eocene *vs* Aquitanian). This gap is puzzling as it occurs separating deep marine Mesozoic deposits and other deep marine

Cenozoic deposits, consequently it cannot be explained by only invoking uplift and emersion, followed by subsidence;
- between the Paludi Fm and the late orogenic siliciclastic deposits (38 Ma *vs* 11 Ma).

During the present PhD thesis, a geological mapping project was aimed at elucidating the Cenozoic evolution of the Longobucco Basin, and frame it within the more general context of Western Mediterranean geodynamics.

Field data concerning the nature and geometries of the basal contacts separating the Paludi Fm from the older units have been collected. This was a crucial step as the local stratigraphy is characterised by multiple unconformity-bounded units. This information is missing in the literature, as the Paludi Fm has always generically been described as "unconformable on the older units". Field work was also aimed at defining the structural setting of the Basin, mapping the main thrusts, and defining the relationships existing among different tectonic units. This provided the physical framework hosting the expected flow of new micropaleontological, sedimentological and lithostratigraphic data. Last, one of the goals of this study was unveiling the relationships existing between the Jurassic rift basin and the Cenozoic Paludi Basin.

Concerning the sedimentology of the Paludi Fm, besides the work of Zuffa & De Rosa (1978), which is essentially focussed on the petrology of arenites, the sedimentological features and depositional geometries of the formation have never been investigated, being this unit generically described as a "flysch". We measured 9 stratigraphic sections located in distinct tectonic units and oriented both down-current and cross-current. Sections have been interpreted using the recent studies concerning the down-current evolution of bipartite turbiditic flows (Mutti *et alii*, 2003; Tinterri & Tagliaferri, 2015; Tinterri & Piazza, 2019) with special reference to the importance of Mass Transport Deposits in tectonically active basins (Ogata *et alii*, 2012; 2014). It must be highlighted that Calabria is a logistically impervious, wild region, where vegetation has pervasively conquered and covered hillsides, also due to the progressive abandonment of pastoralism by local inhabitants. Outcrops are often covered, and the measurement of continuous stratigraphic logs is often made impossible by the dense network of contractional faults and related folds. Despite this, besides measured logs, a wealth of additional information was gathered in isolated fragmentary outcrops discovered during the field survey.

As micropaleontological analyses were mainly aimed at assessing the age of the Paludi Formation, two major conceptual issues arose:
- The possible reworking of faunas, as invoked by Bonardi *et alii* (2005);

- The possibility that we were dealing with separate siliciclastic bodies actually pertaining to different orogenic stages (syn-orogenic *vs* post-orogenic – foredeep deposits *vs* wedge-top deposits).

The first challenge was tackled with thin section analysis, based on sedimentology and on biostratigraphy (benthic foraminifera, planktonic foraminifera and nannofossils).

The second problem-area was addressed i) by means of pairing lithostratigraphy with our subdivision into tectonic units, namely by recognising in the field how lithostratigraphic units populate the different tectonic units, and ii) by assessing if same stratigraphic units display a different age in different tectonic units.

Lastly, in order to establish the age of the compressional phase, and with the more general goal of reconstructing the burial and exhumation history of the Basin, X-ray diffraction analyses of clay minerals, organic matter optical analysis, apatite fission tracks and U/Pb dating on calcite veins have been performed.

To summarize, the objectives that we tried to accomplish are:

- Age of the Paludi Fm (Eocene *vs* Aquitanian);
- Sedimentology of the unit and understanding of its depositional setting;
- Origin of the unconformities and stratigraphic gaps;
- Age of thrusting and burial/exhumation history of the Basin;
- Relationships with the Jurassic rift basin.

The answers to these open questions have a broader impact on our understanding of the evolution of the Alpine/Apenninic system, as they will shed light on the evolution of one of the most internal sectors of the chain, that, as we will see, carries the evidence for a very complex polyphase geodynamic history.

Previous studies about the Paludi Fm

The Paludi formation was never formalised as an official lithostratigraphic unit by the ICS (Italian Commission of Stratigraphy), and not even analysed preliminarily for the "catalogue of the Italian geological formations". The name "Paludi Fm" was proposed by Dubois (1976) as a part of a stream of studies by the "French school" that led in the 1970's to a synthesis of the geodynamic evolution of the CPA. In his PhD thesis, Dubois (1976) ascribed the whole unit to the lower-middle Eocene.

The rocks which would later become the Paludi Fm were known since the end of the 19^{th} century due to research of Italian geologists and palaeontologists like Di Stefano, Fucini and Cortese (Fucini, 1886; Di Stefano, 1904; Cortese, 1895). The work of these authors was pioneering because the Longobucco Basin was essentially unknown at the time (see Lovisato, 1878-1879 for an exception). These studies were linked with the production of the official geological map of Italy on the 1:100.000 scale by Cortese. It is worth mentioning that as of today the sheet #230 "Rossano", surveyed by Cortese, is still the sole official map of the area.

The attention of these early papers was focused on the fossiliferous intervals of the Jurassic succession, and, for what concerned the Paludi Fm, the Authors focussed mainly on the calcarenitic beds rich in resedimented benthic foraminifera (see below). The Authors tentatively referred the Formation to the Eocene according to the presence of "Nummulites".

The next research concerning the Paludi Fm came more than 50 years later. Magri *et alii* (1965), during their geological mapping of the Longobucco basin, aimed at the finding of uranium-bearing ores, proposed an age ranging from the latest Cretaceous to the early Eocene based on planktonic foraminifera assemblages. Being the late Cretaceous assemblages discovered only in few localities, the authors hypothesised that the flyschoid sedimentation began in a restricted basin in the Maastrichtian, becoming widespread only later (early Eocene).

Lanzafame & Tortorici (1980) studied the lithostratigraphy of the whole Longobucco basin, analysing both the Mesozoic and Cenozoic deposits. In their work the Authors addressed the planktonic and benthic foraminifera associations, assessing a "middle" Paleocene – "middle" Eocene age to the Paludi Fm.

In the two outstanding syntheses of the CPT by Ogniben (1973) and Amodio Morelli *et alii* (1976), the Paludi Fm received very little attention. Ogniben only spent few words about this unit, by merely mentioning the preliminary data of Dubois (1970). Ogniben (1973) simply treated it as a follow-up of the Jurassic succession of the Longi Unit, a part of the Calabride Complex. In the work of Amodio Morelli *et alii* (1976), the space devoted to the Paludi Fm is even smaller, and the Authors merely quote generic "Eocene flyschoid deposits".

Many years later, in 2005, Bonardi and co-authors, in their study on the geodynamics of the Calabria-Peloritani Arc (Bonardi *et alii*, 2005), measured stratigraphic sections covering the Paludi Fm and proposed an Aquitanian age for the whole unit, based on a study of nannoplankton associations and considering as reworked all the older planktonic and benthic foraminifera associations, thus excluding an Eocene age.

The lithostratigraphy of the Paludi Fm has been first outlined by Magri *et alii* (1965), who subdivided the unit into four main lithofacies: i) basal conglomerates; ii) reddish marls; iii) arenaceous-marly flysch; iv) cherty limestones and varicoloured jaspers. Conversely, Zuffa & De Rosa (1978) in their study on the petrology of the Unit, and Lanzafame & Tortorici (1980), proposed a three-fold lithofacies subdivision: basal conglomerates and breccias, reddish marls, and pelitic and calcareous-arenaceous turbidites.

The studies of the end of the 19th century and of the '60s and '80s are true milestones in the comprehension of the evolution of the Basin because, as already mentioned, these Authors discovered and described localities that still today serve as key reference points. On the other hand, none of the above cited papers describes i) the nature and geometries of the basal contacts existing between the Paludi Fm and the older units, and ii) the position of sampling localities with respect to the local stacking pattern of tectonic nappes. In the present study, we tackle these two weaknesses, based on the results of a recent mapping project in Calabria, and convert them into the foundations of our first-order tectonic-stratigraphic interpretations.

Lithofacies description and age assessment

LITHOFACIES 1 – REDDISH MARLS

This lithofacies resembles the classic Tethyan "Scaglia Rossa" facies, as it is represented by reddish marls and shales with whitish/greenish "flames", in thin to medium beds, rich in planktonic foraminifera.

This lithofacies has a very limited extent being exposed in the third tectonic unit only, along the Colognati River valley (westernmost sector of the basin), resting unconformably on the Aalenian *Zoophycos* marls and on granites through a NE-SW striking escarpment. The basal part of the lithofacies is not exposed, its minimum thickness is 30 m.

This lithofacies represents essentially the background hemipelagic sedimentation before the beginning of the gravitative instabilities and is everywhere overlain by the third lithofacies. As highlighted in the sedimentological chapter, it is better preserved in the distal area of the basin (far from the NW-SE directed escarpment) where the strong erosive fluxes arrived later. Despite being under-represented in the basin (especially because good outcrops of the base of the formation are missing), we decided to separate this lithofacies from the third one due to its geological significance (see below). The occurrence of the first arenaceous bed has been chosen as a reference to place the boundary between the first and third lithofacies. The pelitic/arenaceous ratio has been discarded as a criterion for separation, because it cannot be objective in the field, especially where outcrops are extensively covered and poorly exposed. Furthermore, the occurrence of such sandy interval it has not only a lithostratigraphic meaning but also a precise geological one, being related with the start of the gravitative instabilities.

LITHOFACIES 2 – (RUDITIC-ARENACEOUS-PELITIC ALTERNATION) BRECCIAS

This lithofacies is represented either by matrix-supported or by clast-supported conglomerates and breccias alternating with a variable amount of arenaceous/pelitic intervals. It crops out continuously along a W/NW – E/SE- directed belt from the Laurenzana River to Vallone Sant'Elia, and is missing West of Paludi, where only the lithofacies 2 and 3 are exposed. Conglomerates/breccias can be mapped only in the lowermost tectonic unit, while they are missing in the higher tectonic units.

This lithofacies as a whole forms a megaclastic wedge that can be divided in minor clastic intervals (Key Beds) separated from each other by arenaceous/pelitic intervals.

Being the fluvial valleys where the lithofacies is exposed parallel to its dip (~N-S), lateral (along strike) variations in a cross-current direction are poorly detectable. The only remarkable exception is the Laurenzana Valley, as it cuts the lithofacies obliquely with respect to its dip, and due to this unveils a complex sedimentological scenario which is not visible elsewhere.

Basal and upper contacts

The basal clastic interval covers the Hercynian metamorphic basement through an unconformable onlap contact. The conglomerates are typically unconformably overlain in turn by the third lithofacies (hybrid arenites). This passage is not sharp as the third lithofacies still contains ruditic beds, that are thin, matrix-supported and with clasts not exceeding the size of pebbles.

Along the Trionto Valley, where the river cuts deeply through the hillside, the contact between this lithofacies and the basement can be followed along a difference in altitude of 200 m. Here the basement forms a south-westward dipping escarpment (Fig. 16) on which the conglomerates rest with good lateral continuity, being only locally replaced by sandstones. Along the river valley the conglomerates are thicker than uphill, where they pinch out to 0 m, evidencing their general wedge geometry. The same scenario can also be appreciated along the Vallone Sant'Elia and Coserie River.

Toward the pinch-out culmination of the conglomerates (where their thickness tends to 0), an

Figure 16: Contact between the lithofacies 2, 3 and the basement. Conglomerates pinch-out toward the north up to 0 meters, with hybrid arenites directly resting on the basement.

angular unconformity between this lithofacies and the arenites is evident, as the conglomerates dip more steeply than the turbidites.

A well preserved meso-scale outcrop of the basal contact is found along the Coserie River valley. Here, a paleosurface can be observed in detail, carved in the phyllites, exhibiting a dip towards the SW and hosting the onlap contact with the conglomerates.

The contact surface is irregular, displaying a small-scale “spurs and grooves” erosional pattern, with a relief of up to 10-20 cm (Fig. 17). As proposed by Sobiesiak *et alii* (2018), this type of basal interaction can be classified as “no slip flow”, encompassing those flows that have a strong interaction with the underlying substrate. This interaction can occur in the form of both erosion and substrate deformation. The latter sub-category can be *a priori* excluded in our case because it only regards unlithified substrates. The styles of basal interaction of MTDs are generally poorly documented in outcrop (see Sobiesiak *et alii*, 2018, for a list of outcrop examples vs seismic derived ones) so our Calabrian example adds to the field-based dataset, aiding in the interpretation of processes that occur at the base of a MTD.

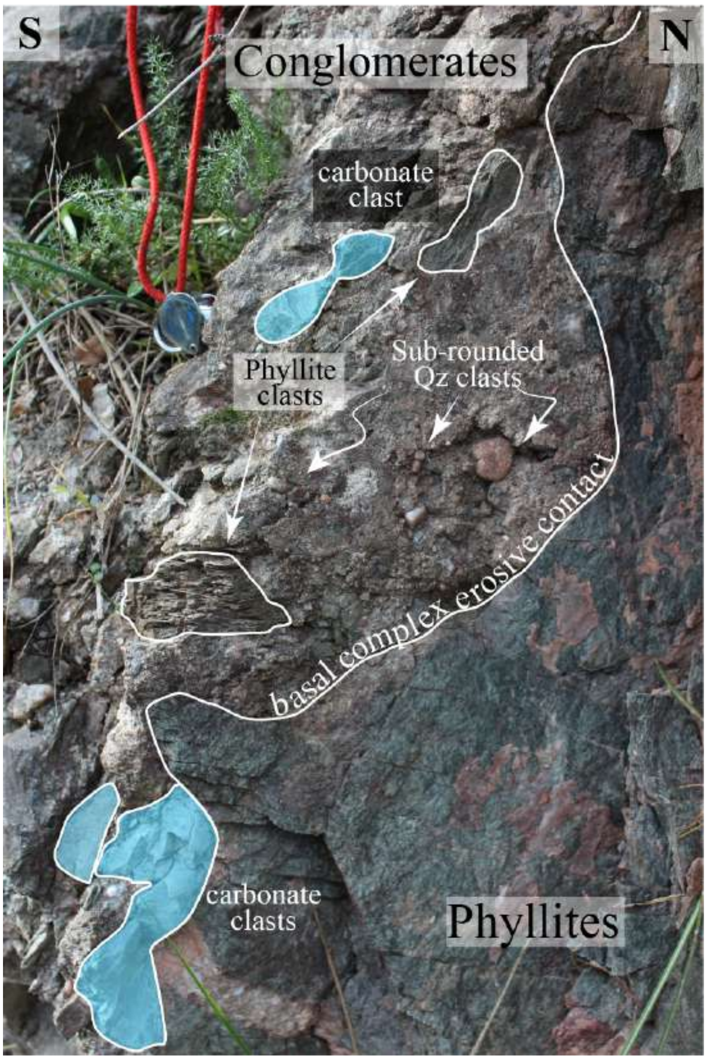

Figure 17: Detail of the contact between the Lithofacies 1 and the phyllites. Please note the sub-rounded quartz grains (arrowed).

Description

As it will be highlighted in the description of the stratigraphic sections, the lithofacies shows a down-current variability.

The basal and most proximal (with respect to the NW-SE oriented onlap unconformity) intervals are represented by clast-supported polymictic breccias. These deposits are very thickly bedded (up to 10 m), and the conglomerates/breccias appear as massive and chaotic bodies, without any internal organization (Fig. 18).

A muddy matrix is essentially missing, being the finer component represented by very coarse sands.

Grain size populations are highly variable, ranging from sand grade to large boulder size (up to 1,8 m across clasts). Very poor sorting is the norm, as mm-size grains coexist with clasts several decimetres across within the same bed, with no internal organisation whatsoever. A typical Paludi Fm conglomerate bed appears as cobbles floating in a pebbly matrix.

The conglomerates and breccias are typically polymictic, being made of clasts pertaining to the basement (both igneous and metamorphic) and to various Formations of the Caloveto Group, including the Sant'Onofrio sub-group.

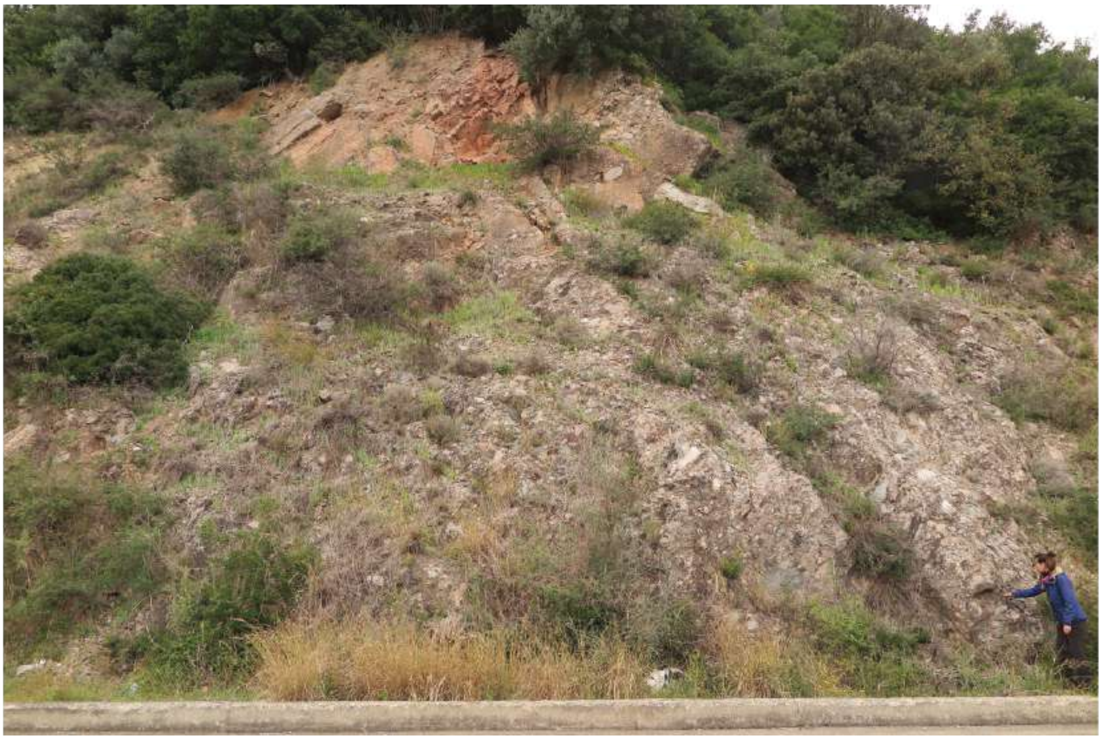

Figure 18: Typical appearance in outcrop of the very thickly bedded, clast-supported breccias.

Moving away from the onlap surface (and consequently in the stratigraphically higher parts of the interval) the lithofacies changes a little. Bedding decreases in thickness with the appearance of 30 cm- thick beds in association with very thick beds (up to 7 m, on average 1 m). Progressively the arenaceous intervals start to increase in frequency and thickness, being represented by medium bedded, coarse, structureless sand intervals, often eroded by the clastics. The pelitic component gradually increases too, being represented by thinly/moderately bedded intervals of reddish marls that are anyway deeply eroded. The upper part of the basal lithofacies can be thus considered a ruditic/arenaceous/pelitic alternation with the ruditic part prevailing among the others.

In this part of the lithofacies the fine muddy matrix is still completely absent, but a marked increase in the sandy component can be appreciated (Fig. 19).

Figure 19: a) Breccias/conglomerates alternating with arenaceous beds; b) general aspect of breccias rich in a coarse sandy component.

Locally, in the interval with a major sandy component, a coarsening upward trend can be observed. This is related to the contrast in density between the coarser clasts and the matrix that generates buoyance forces: when the flows halt, their down-stream larger clasts remain in the upper part of the deposits due to frictional freezing.

Grain size and composition are essentially the same of the basal part of the lithofacies, with the gradual increase in basement-derived clasts with respect to the sedimentary clasts. Furthermore, in the upper part of the lithofacies, also plastically deformed arenaceous and marly clasts are present. These elements were scraped off from the substrate during the emplacement of the clastic intervals.

An exception to the above is the Laurenzana outcrop where the megaclastic interval is matrix-rich and characterised by plastically deformed elements pertaining to the other two lithofacies.

As a general rule, clast shape depends essentially on their lithology. For example, phyllite and chert blocks are angular or subangular, while limestone and marl clasts are generally sub-rounded. A remarkable exception is represented by granite and quartz clasts, which are all very well rounded (see detail in Fig. 16). In our opinion this implies that they may have experienced multiple episodes of deposition/recycling.

Plastically deformed clasts can also be observed in the higher stratigraphic part of the lithofacies, representing three typologies:

1. Elements made of reddish marls (Fig. 20). These elements range in size from few millimetres to 1 meter (longer axis). They display an array of plastic deformations: asymmetric rootless folds, pseudo-SC structures, pseudo-sigma structures, hydroplastic folds, asymmetric boudinage and obviously mechanical lithologic mixing;

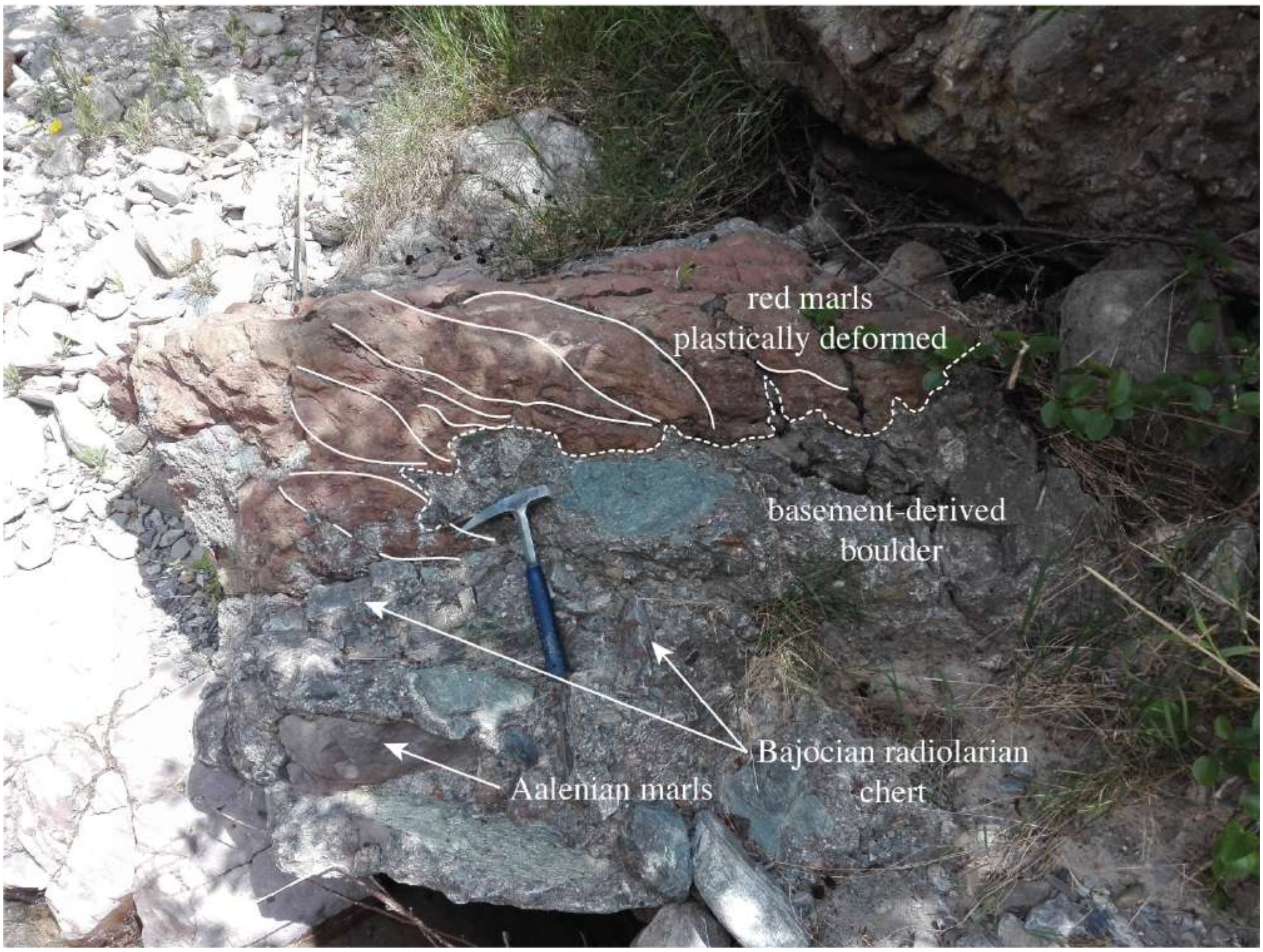

Figure 20: Example of plastically deformed clast of reddish marls showing internal deformation. The clast is compenetrated by a basement-derived boulder. The deposit hosts also Mesozoic clasts.

2. Megablocks (bed-stacks) pertaining to the Maiolica Fm (Lower Cretaceous). The Maiolica clasts display a mixture of brittle/ductile deformation, as described in the following chapter;
3. Blocks made of coarse/very coarse sands. They exhibit essentially hydroplastic folds, rootless folds and boudinage;
4. Blocks pertaining to unknown lithostratigraphic units covering the time span of the Lower Cretaceous-Paleogene hiatus.

Limestone clasts are locally silicified. Flat clasts are typically randomly oriented within any given clastic body.

In some localities clasts of various sizes (from pebble to boulder) exhibit fractures that can be filled either by reddish marls or by silt (Fig. 21). Such fractures, in the larger blocks, can be up to 10 cm. This feature is not common and has been detected only in the intervals where a finer matrix is present (see thin section description). This can be interpreted as an evidence of overpressure in the matrix that led to hydrofracturing.

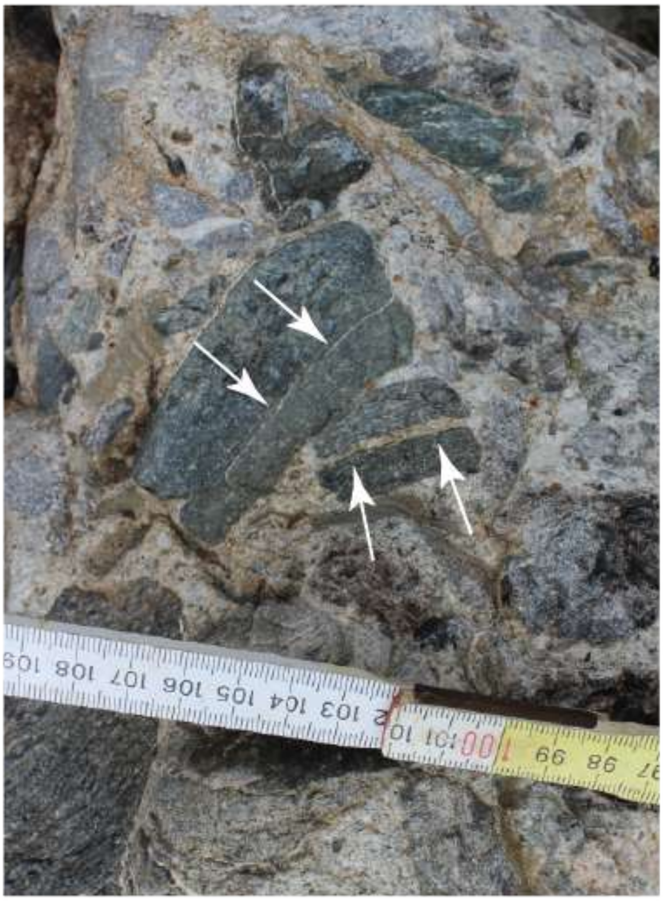

Figure 21: Example of hydrofracturing in the conglomerates.

On the whole lithofacies scale, we can observe a thinning and fining upward trend, the latter related to a general increase in the coarse sand component and a general prevalence of cobble-size clasts with respect to boulder-size clasts.

Thin section analysis

In thin section, additional observations can be made, especially concerning the matrix. Typically, breccias/conglomerates are densely packed, clast-supported and matrix-poor. In thin section a further subdivision can be made considering the matrix/clasts ratio:

i) Intervals where a clayey/silty matrix is completely missing. The groundmass is itself represented by a microbreccia with moderately sorted elements ranging in size from >

1 mm (coarse to very coarse sand). Clasts are sub-angular and composed by Qz, phyllites etc. Where present, carbonate clasts are sub-rounded and the siliciclastic elements (like phyllites) penetrate them with sutured contacts due to pressure solution. Among Qz and phyllite clasts, contacts are both tangential and sutured. A dark clay matrix is sometimes present around clasts, also due to pressure solution. In thin section, as also observed in the field, any fitting between clasts is missing or very rare. Parallel sets of dissolution seams are visible within carbonate clasts producing a stylolaminated fabric; they develop parallel to the long axis of the clast (Plate 1);

ii) Intervals with a moderate percentage of silty matrix. In thin section we can observe a poor sorting with clasts ranging in size from fine sands to granules, dispersed in a reddish silty matrix. Clasts are rarely in contact, although tangential contacts can be observed, while sutured contacts are extremely rare. Fitted fabrics among clasts and hydrofracturing can be observed in the field as well as in thin section (Plate 1);

iii) Intervals with abundant marly/silty matrix. This typology has been detected only in the Laurenzana River valley at the very base of the megaclastic interval. Samples are moderately sorted. Clast contacts are essentially missing (Plate 1).

Remarkably, in thin section also Larger Benthic Foraminifera have been observed (Plate 1). They are clearly damaged, as it will be described extensively in the chapter related to hybrid arenites.

LITHOFACIES 3 – HYBRID ARENITES (ARENACEOUS/PELITIC/RUDITIC ALTERNATIONS)

This is the most widespread facies of the three. In the lower tectonic unit, it unconformably covers breccias of lithofacies 2 at lower altitudes, while at higher altitudes it onlaps the metamorphic basement directly. West of Vallone Sant'Elia, in the lowermost tectonic Unit, the turbidites onlap the metamorphic basement directly, as the conglomerates are missing. In both cases a South-dipping escarpment carved in the basement can be mapped with good lateral continuity. In the higher tectonic units, in the western sector of the Basin, this lithofacies follows the reddish marls, or it rests unconformably on the late-Hercynian granites or on various Jurassic Units. Often it is preserved only as a thin cover, as at Serra Livera, at the top of eroded anticlines.

Lithofacies 3 is made up of an alternance of arenites, pelites and rudites, with the arenites dominating, and it displays a clear horizontal and vertical organization.

At the toe of the escarpment (proximal area) thickly bedded, structureless, very coarse sands/pebbly sandstones prevail. Beds are often amalgamated showing plastically deformed marl chips. The pelitic beds are few and often deeply eroded. Some true ruditic beds (up to 3 m thick) are sporadically found. Rip-up clasts up to 10 cm across, made of reddish marls, are also very common, displaying plastic deformation as pseudo-sigma structures. In thicker beds amalgamation surfaces are present. Locally, some laminated beds (parallel and cross lamination) are observed.

In the distal areas and moving up-section thick bedded coarse sands prevail with respect to the thickly bedded very coarse sands/pebbly sandstones. Simultaneously, the marly interbeds start to increase showing less evidence of erosion. Similarly, at higher elevation close to the contact with the basement, marly interbeds and faintly bedded coarse sandstones prevail.

Reddish marls are dominant in the lower few tens of metres, being replaced upsection by thinly bedded yellowish/greyish marls.

Two additional features related to mass transport deposits are remarkable:

- Slumped intervals: recognised in the basal part of the lithofacies, very close to the toe of the escarpment. They occur at different scales, forming centimetre- to decametre-thick intervals, and are typically non cylindrical (sheath folds) (Fig. 22):
- Chaotic beds: recognised in the upper and distal part of the lithofacies, below thickly bedded coarse sands or pebbly sandstones. Such intervals are the product of the interaction of the flows with the substrate, which they deform. Such beds are characterised by shear-surfaces in the marly layers at the contact with the coarser deposits. Such intervals develop pseudo-SC structures. More competent arenaceous layers are deformed by means of hydroplastic folds and boudinage, while marly intervals are deformed by pseudo-sigma structures.

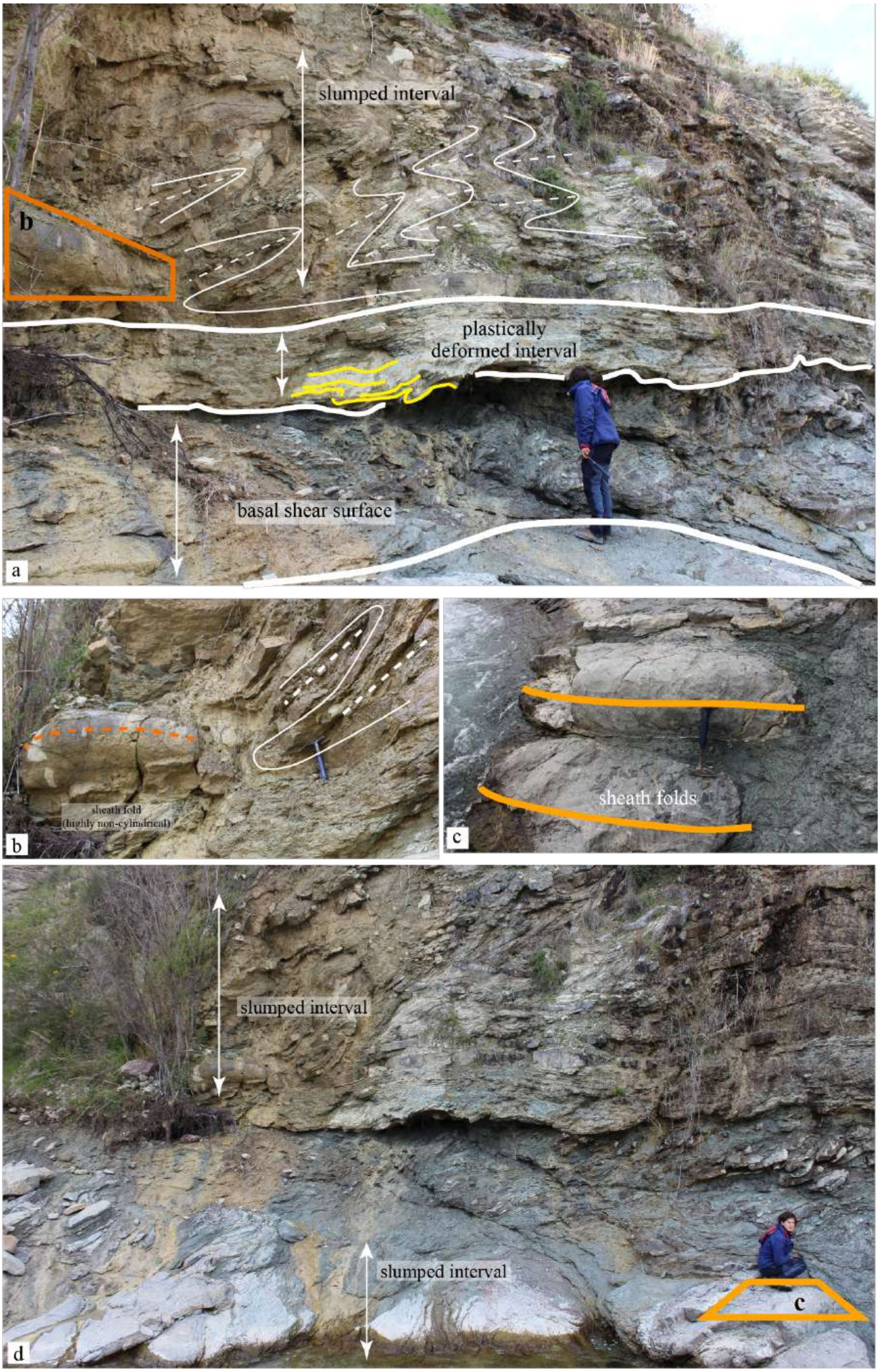

Figure 22: a) and d) Examples of metric slumped intervals within the third lithofacies; b) and c) detail of a sheath fold.

Thin section analysis

From thin section analysis it can be observed that these deposits are typically clast-supported, poorly/moderately sorted. The matrix is represented by reddish/grayish marls. The siliciclastic fraction is represented by angular phyllite clasts and quartz, ranging in size from 0.15 mm to 2 mm. As already evidenced in other outcrops, Qz can be sub-rounded. Carbonate clasts pertaining to the Mesozoic cover can be referred to two typologies:

- lithified fragments, very rare, ranging in size from very coarse sand to granule;
- rounded marl clasts, characterised by well-defined boundaries, either bearing micro-foraminifera or barren.

The carbonate fraction and especially the rip-up clasts are penetrated by siliciclastic elements being still plastic and not yet lithified at the time of their emplacement.

An abundant bioclastic component is represented by larger benthic foraminifera, while glauconite is locally observed as the authigenic component. Based on their composition, these intervals can be classified as hybrid arenites (Zuffa, 1980).

An important component of these hybrid arenites are larger benthic foraminifera. Their tests are badly damaged both in densely and loosely packed intervals, showing in the first case conspicuous pressure-solution at the contact with siliciclastic grains. The marginal part of tests is generally fragmentary or missing, and the external whorls are mostly abraded. In densely packed and matrix-poor intervals, tests compenetrate each other. The most damaged specimens are the flat-shaped ones, such as the orthophragmines, to the point of being barely identifiable. Flat tests are typically randomly oriented in thin section. A key observation is that all the benthic foraminifera clearly occur as loose elements within the matrix, and not as part of larger lithoclasts, as it is instead the case with other microfossils (Plate IV-f). Larger benthic foraminifera thrive in shallow-water carbonate environments. An in-depth discussion about the Eocene shallow water depositional systems of the Tethys is beyond the objectives of this thesis, but a brief discussion will be presented, in order to fix sedimentological constrains. The Eocene is a period of re-organization of carbonate platforms, as, after the PETM (Paleocene-Eocene Termal Maximum, Scheibner & Speijer, 2007), coral reefs start to demise. Simultaneously, larger benthic foraminifera, started to spread becoming the dominant carbonate producers with red algae, molluscs and bryozoans (Pomar *et alii*, 2017). In the early Eocene (during the so-called Early Eocene Climatic Optimum) Tethyan carbonate platforms were almost exclusively characterised by LBF-dominated ramps. These depositional settings can be referred to heterozoan carbonate factories and in particular to the foramol skeletal association. Although an heterozoan factory shows slower diagenetic rates with respect to a photozoan factory (10^4 vs 10^1 years; Fluegel, 2010) and in such settings early diagenesis plays a minor role, in modern environments it has been demonstrated that aragonitic sediments undergo significant dissolution in about 30.000 years, leading to the precipitation of cements in skeletal pores (see Rivers *et alii*, 2008) (see Knorich & Mutti, 2006 and Nelson & James, 2000 for fossil examples). Besides these sedimentological quantitative studies, field examples provide evidence for relatively fast diagenetic rates. The Brecce della Renga Fm, for example, is a huge megaclastic body, developed in the early Tortonian/early Messinian in the hangingwall blocks of bulge related normal faults (Compagnoni *et alii*, 1990; Critelli *et alii*, 2007). Such rudites bear lithified clasts and huge olistoliths of the "Calcari a Briozoi e Litotamni" Fm, that is an heterozoan association of latest Burdigalian/early Tortonian *p.p.* in age (Civitelli & Brandano, 2005). The presence of foraminifera as loose elements therefore strongly suggests that they are penecontemporaneous with the gravity flows and must be considered as resedimented and not reworked bioclasts. Consequently, they can be considered as reliable tools for biostratigraphic assessments. As it will be described in the following chapter, foraminiferal assemblages are indeed age-consistent, with no detectable faunal mixing, suggesting that the larger foraminifera are not reworked, and their poor preservation is the result of downslope transport in the company of coarse clastic material. Stylolitic clast contacts are the norm, with the development of dissolution seams.

Even in the matrix-rich samples, the reddish marl matrix exhibits complex irregular/conjugate sets of stlylolites. These can be considered as the product of burial- or compression-related compaction rather than of gravity-driven pressure solution.

General Considerations

The clasts in both Lithofacies 2 and 3 pertain to the Paleozoic basement and to the Mesozoic succession, which therefore had to have been exhumed. This is confirmed by the nature of the basal contact of the two lithofacies with the substrate. The basal contact of the Paludi Fm is everywhere unconformable, as it can be found resting on different stratigraphic units (radiolarian cherts, *Zoophycos* marls, Maiolica) as well as on the basement. Furthermore, in some localities (Coserie, Trionto and Il Torno for example), mappable onlap contacts mark the existence of a submarine paleoescarpment dipping toward the south-west. All the above indicates that the Paleozoic/Mesozoic substrate was already, and/or was being actively deformed at the time of sedimentation of the Paludi Fm, being subjected to uplift and erosion. The Paludi Fm as a whole is a Mass Transport Complex, being characterised by a variety of distinct mass transport processes ranging from debris flows to slumps (Fig. 23).

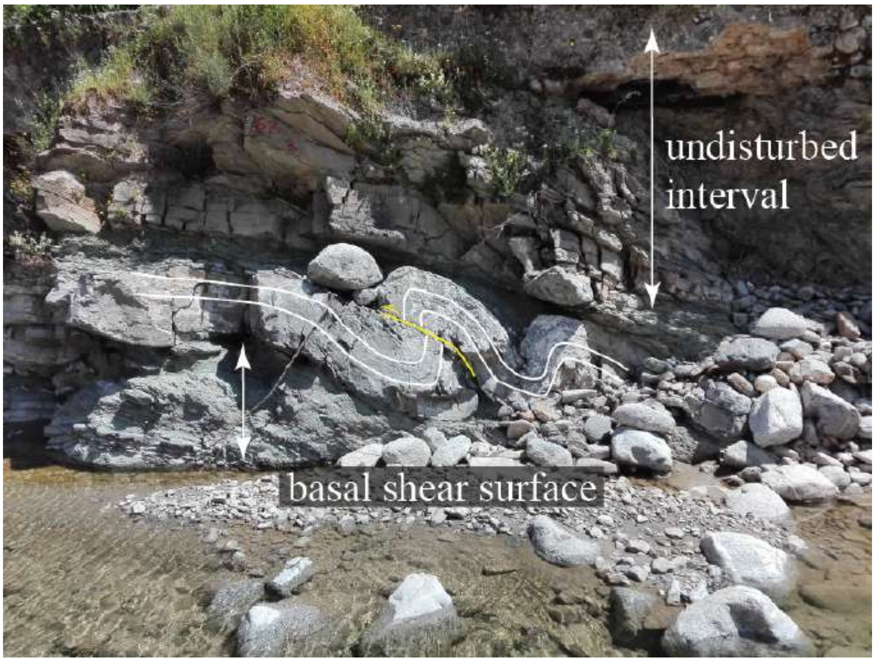

Figure 23: Slumped interval in the third lithofacies.

OLISTOLITHS

As mentioned above, several mappable olistoliths are locally found embedded within the arenites and marls. We can recognise two different types of reworked objects: i) Jurassic elements: lithified/semilithified huge blocks/bedstacks; ii) plastically deformed Cretaceous clasts.

The first type was discovered in the neighbourhoods of Paludi, in a narrow area of about 1km^2, at Cozzo Rango, in the unnamed valley south-west of Paludi and at Cozzo dello Scivola. The olistoliths are huge, their longer axis being up to >70, their internal bedding is still recognizable, and they are all made of radiolarian chert. MAGRI *et alii* (1965) considered these elements as a facies of the Paludi Fm, but they are clearly embedded in the arenites. The Authors compared these olistoliths with the Jurassic radiolarian cherts outcropping in the western part of the Basin at Sant'Onofrio and Torno (Colognati Valley). Here an *in situ* stratigraphic succession pertaining to the Sant'Onofrio sub-group is exposed (*Zoophycos* marls, Radiolarian Cherts, Aptychus/*Saccocoma* Limestone and Maiolica), which is covered by the hybrid arenites of the Paludi Fm (lithofacies 3). Magri *et alii* (1965) did not recognise the *Zoophycos* marls as being a Jurassic formation and mistook them for the reddish marls of the Paludi Fm, consequently considering the radiolarian cherts as a lithofacies of the Paludi Fm.

Figure 24: a) and b) General aspect of the interval characterised by plastic deformation. It is possible to observe the rheological contrast between the more competent part of the stack and the more shaly one, respectively deformed by pseudo-Sigma and pseudo-S-C structures. Trionto Valley.

The second type of olistoliths is widespread, but at the same time they are harder to identify in the field, as these elements are cosmetically similar to the mudstone interbeds of the third lithofacies. They can be detected only in well exposed sections where their sedimentological features, like plastic deformations, are clearly visible, otherwise they can go unnoticed due to their elusive nature, mimicking a "normal" bedded succession. The most instructive example of such typology of olistoliths comes from the Trionto Valley, where a six metres thick interval stands out for being different from the encasing rocks, despite being sub-concordant with them. It is made of alternations of whitish/greyish limestones, calcareous marls and black shales. The more competent calcareous parts are deformed by metre- to centimetre-sized pseudo-sigma structures, while the less competent shales are deeply foliated and deformed by pseudo S-C structures (Fig. 24). All the above is thus in contrast with the regular bedding of the surrounding arenites/pelites. In thin section this interval reveals an Aptian/Albian planktonic foraminifera assemblage (*Globigerinelloides*, *Hedbergella*) (Plate I). Nannofossil analysis corroborates this age assessment (*Nannoconus* and *Watznaueria*). This interval can be consequently interpreted as an olistolith made of a detached Cretaceous part of the succession, cannibalized by the down streaming flows, and plastically deformed during its emplacement.

Similarly, more Cretaceous planktonic foraminifera were discovered in comparable contexts, that is in exotic clasts. The difficulties in identifying these allochthonous elements in the field can explain why some Authors considered the Paludi Fm as being Cretaceous in age.

AGE ASSESSMENT

Samples were collected both from the measured sections and from fragmentary exposures across the whole outcrop area of the Paludi Fm, and has been analyzed in collaboration with Prof J. Pignatti (Sapienza).

Larger foraminifera-bearing samples yield rather uniform assemblages, consisting of *Nummulites* spp., *Discocyclina* spp., *Orbitoclypeus* sp., *Asterigerina cayrazensis* Sirel & Devecíler, 2017, *Ornatorotalia granum* Benedetti, Di Carlo & Pignatti, 2011, *Ornatorotalia spinosa* Benedetti, Di Carlo & Pignatti, 2011, *Cuvillierina* cf. *vallensis* Ruiz de Gaona, 1948, *Granorotalia sublobata* Benedetti, Di Carlo & Pignatti, 2011 and *Medocia blayensis* Parvati, 1970. As to the age indicated by the larger foraminifera in our samples, there are two options (Fig. 25):

- Combining the known ranges for these taxa and considering that the known range of *Asterigerina cayrazensis* corresponds to biozones SBZ 11-SBZ 12 of the Shallow Benthics Zonation of Serra-Kiel *et alii* (1998), considered until recently as latest Ypresian. After the establishment of the Global Stratotype Section and Point (GSSP) of the base of the Lutetian stage in the Gorrondatxe section (Molina *et alii*, 2011) and the recent recalibration of these biozones by Rodríguez-Pintó *et alii* (2022), the Ypresian/Lutetian boundary occurs within SBZ 12. Thus, according to the current knowledge, an SBZ 11-SBZ 12 age indicated by the larger foraminiferal assemblages is latest Ypresian to basal Lutetian, spanning from the upper part of chron C23n to the lower part of chron C21n, and corresponding to the upper part of NP12 to the upper part of NP14 nannofossil zones. According to the age model of the GTS 2020, this interval corresponds to ca. 50.5 to 47 Ma.
- If instead a more restrictive approach is followed, i.e., that there is a single assemblage, and only the concurrent ranges of these taxa are taken into account, because *Cuvillierina vallensis* does not extend into SBZ 12, the age is that of SBZ 11, *i.e.*, latest Ypresian, spanning from the upper part of chron C23n to the lower part of chron C21r, and corresponding to upper part of NP12 to the

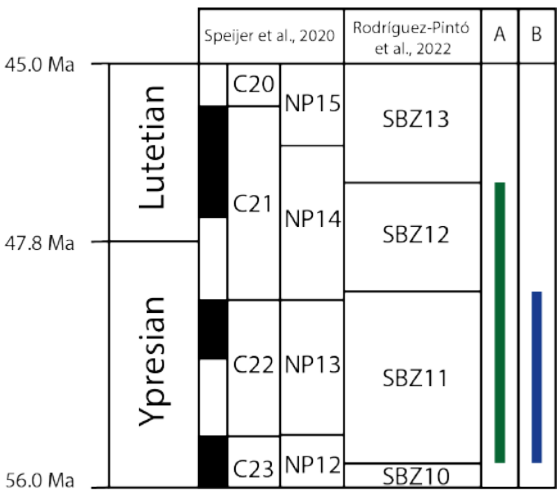

Figure 25: Age assessment of the Paludi Fm: the two proposed attributions are evidenced in columns A and B.

base of NP14 nannofossil zones. According to the age model of the GTS 2020, this interval corresponds to ca. 50.5 to 48.5 Ma.

Thin sections from the marly intervals contain abundant planktonic foraminifera. Samples are typically characterized by the genera *Morozovella*, *Akarinina* and *Subbotina*. In particular the occurrence of *Morozovella* gr. *formosa*, ranging between the E4-E6 zone of Breggren & Pearson (2005), along with the abovementioned LBF taxa point to an Ypresian age for the analyzed samples.

Because neither the genus *Hantkenina*, nor the genus *Globigerinatheka*, whose first occurrences are in the middle Lutetian (E8 – top of the SBZ13), have never been found in the sections, we can point to an early Eocene age (late Ypresian/base of Lutetian) as already stated with LBF.

The Jurassic Longobucco rift basin and the Cenozoic Paludi basin: evidence for one welded element

This thesis has greatly benefited from the knowledge collected in the past years by our research group. In particular, the Jurassic evolution of the basin has been thoroughly investigated in the last years, and the main Jurassic elements have been identified and mapped in the field (Santantonio *et alii*, 2016; Innamorati & Santantonio, 2018; Fabbi & Santantonio, 2020). This premise is mandatory as the Cenozoic basin is superimposed above the Jurassic rift basin that is characterised, as already explained, by an extreme paleotopographic complexity.

For the present thesis, the basin has been divided in four tectonic units, in order to verify the consistency between the age assessments, the facies and their position in the tectonic edifice (Fig. 26).

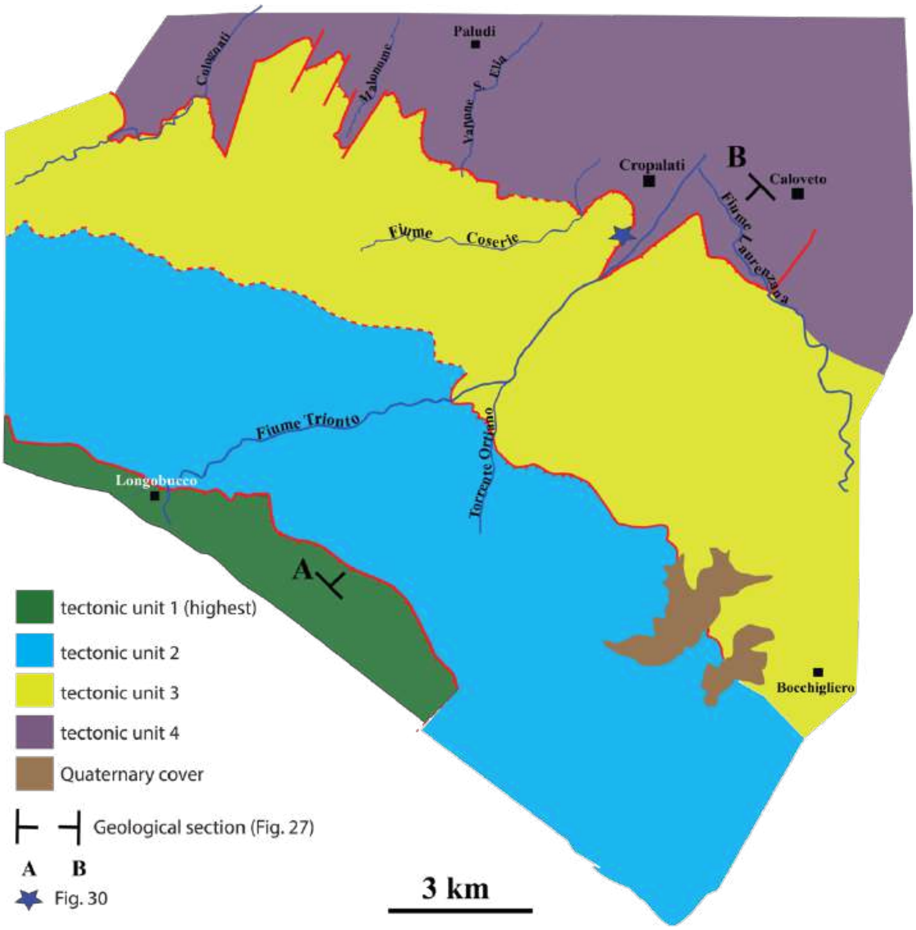

Figure 26: Tectonic units of the Longobucco basin.

The units are separated by three south-west dipping thrusts with a top-to-the Northeast sense of shear. These elements display high lateral continuity, as they can be mapped for more than 10 km along their strikes.

It is worth noting that the description of the tectonic units is further complicated by the extremely complex stratigraphy of the Longobucco Basin, inherited after the Early Jurassic rifting phase, that dismembered the paleotopography in structural highs, lows and basin-margin complexes (Fig. 27). This architecture is mirrored by abrupt lateral lithofacies variations and unconformable stratigraphic contacts (onlap, downlap, ecc). Another important feature, which cannot be underestimated, is the role of inherited submarine paleoescarpments during the growth of the orogenic wedge. When basement-made structural highs (or basin margins) are subjected to compressive stresses, the strong rheological contrast between the basement and the thin-bedded pelagic units favors the development of buttressing, producing disharmonic parasitic folds, with discontinuous axial planes. This is the case with the north-western sector of the basin, where a >14 km long preserved Jurassic basin-margin paleoescarpment has been mapped along the Torrente Colognati Valley.

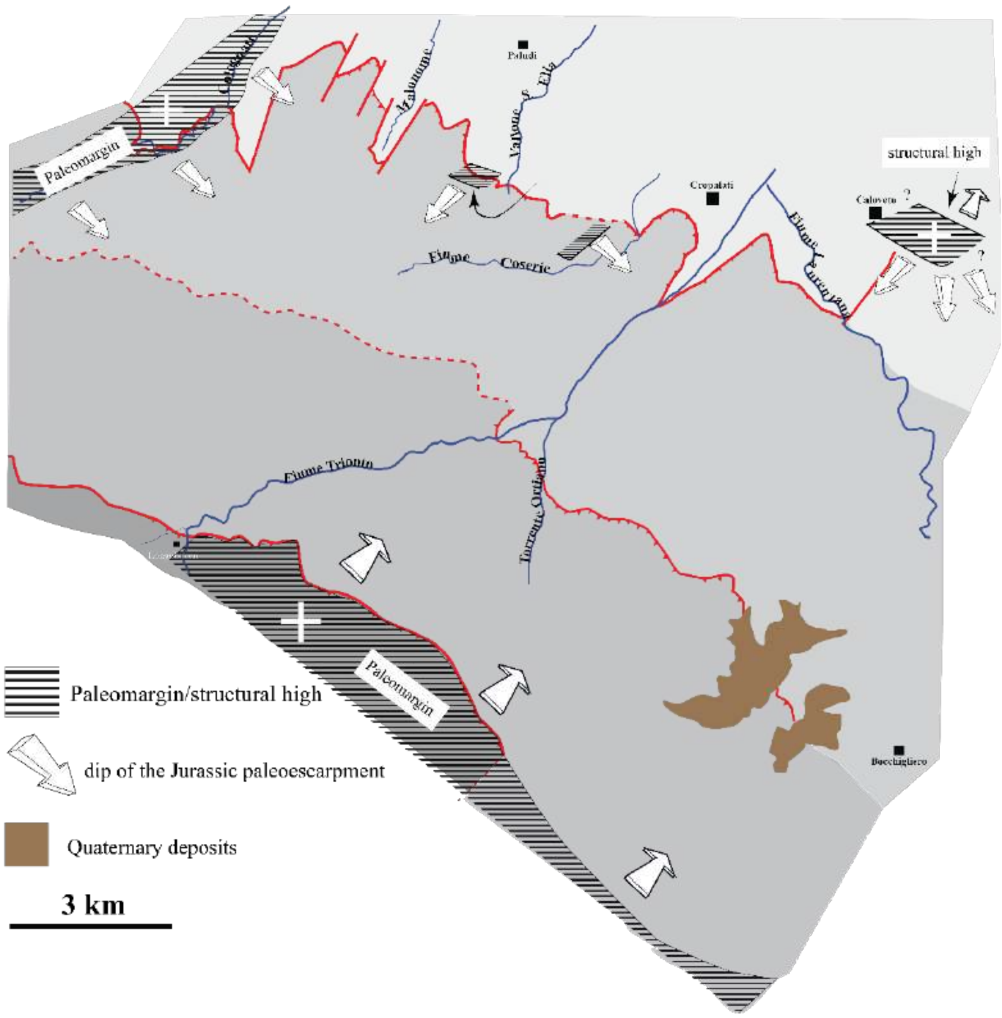

Figure 27: Relationships between tectonic units and Jurassic paleotopography.

A representative transect along the Trionto Valley will be described below, where the tectono-stratigraphic configuration of the basin is clear (Fig. 28).

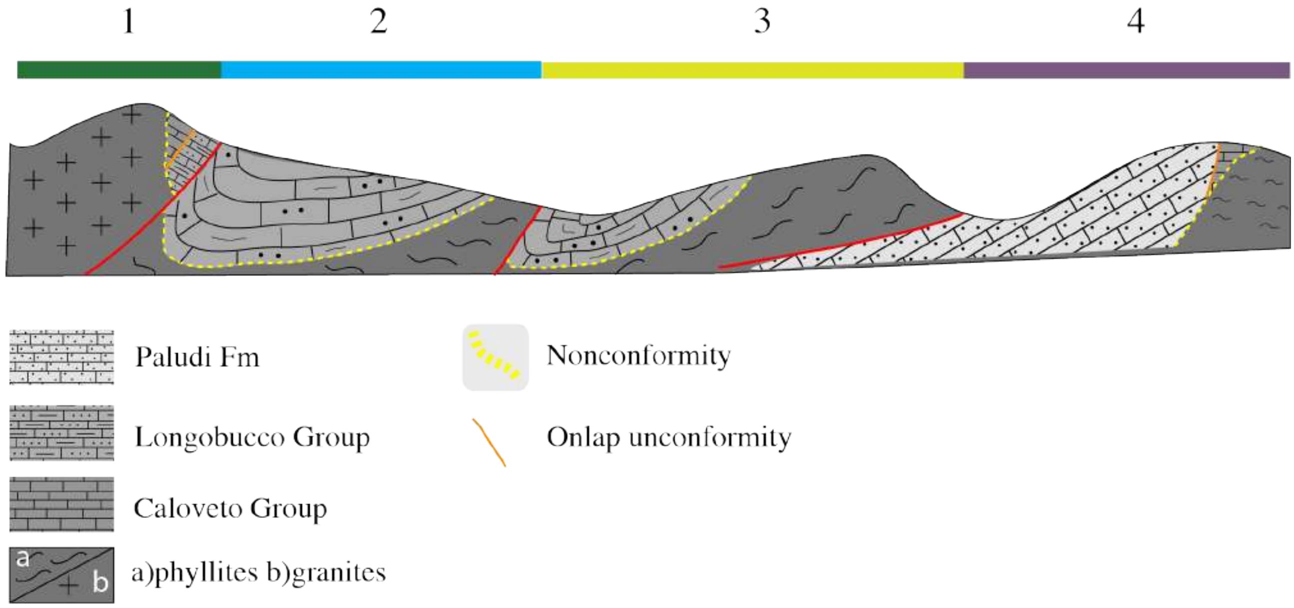

Figure 28: Geological section across the Longobucco basin.

The **First (highest) tectonic unit** is located at an altitude of more than 1,000 metres, within the Sila Plateau. It is characterised by an hangingwall anticline with an overturned to sub-vertical forelimb. This sector has been interpreted as the southern paleomargin of the Jurassic basin and mapped for about 20 km from Longobucco to Bocchigliero, and is characterised by Pliensbachian shallow-water carbonate bodies of the Lower Caloveto Fm fringing a Jurassic escarpment carved in the igneous and metamorphic basement. All the above is then onlapped by the early Jurassic units pertaining to the Longobucco Group (Innamorati & Santantonio, 2018). Despite the presence of the thrust plane, the original contacts between the basement and the Jurassic units are exceptionally well preserved and are only locally rotated. Remarkably, among the fir-woods of the Sila Range, isolated patches, few decimeters in thickness (Fig. 29) of the Paludi Fm have been mapped, representing the topographically and tectonically highest witnesses of a Cenozoic basin. Unfortunately, micropaleontological analyses were inconclusive here. Thin section analysis revealed the common presence of clay chips, bearing planktonic microforaminifera of contrasting ages: *Marginotruncana linneana*, (Turonian-Campanian) *Marginotruncana* cf. *marginata* (Santonian-Maastrichtian) and Palocene specimens pertaining to the genus *Subbotina*. Consequently, a reliable age assessment of this outcrop cannot be established. The discovery of this outcrop, however, is an indication that a basin, late Cretaceous in age, must have existed to the South of our main study area, which is dominated by basement outcrops, but was uplifted and eroded. Furthermore, the presence of the Paludi Fm in the middle of Sila, so distant from the areas where canonically this unit is known, allows us to assert that the extent of the Longobucco Basin itself, in the Eocene, was greater than previously thought. The presence of deeper water deposits with a planktonic microfauna implies that a severe uplift of the whole area must have occurred in the post compressional phase in order to erode it completely. Finally, the fact that the Formation rests on the basement directly suggests that much of the Mesozoic pre-compressive depositional system had already been eroded in the Eocene.

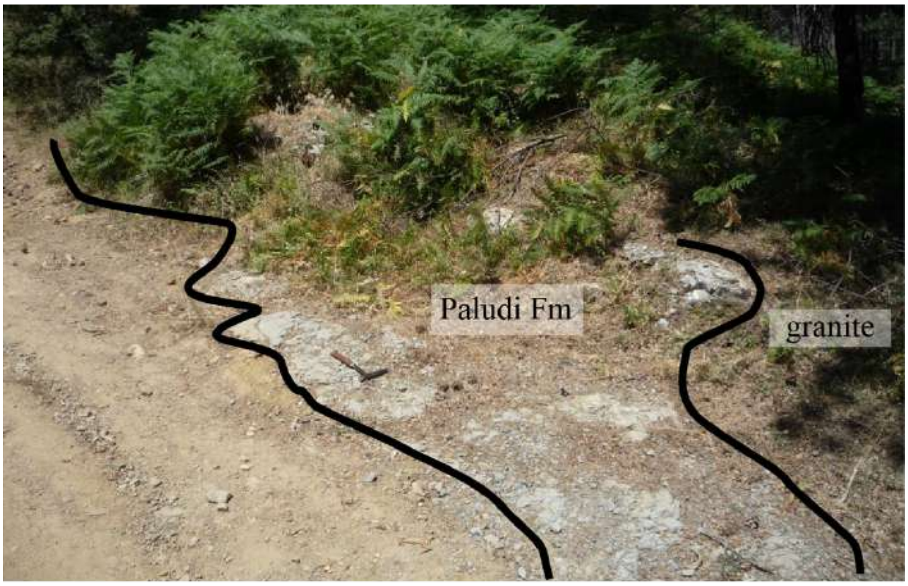

Figure 29: Small outcrop of the Paludi Fm resting unconformably on the igneous basement in the Sila Plateau.

The **Second Tectonic Unit** is made of a very tight, asymmetric footwall syncline with an overturned to sub-vertical south-western flank and an upright northeastern flank, gently dipping toward the SW. The syncline is characterised by formations belonging to the Longobucco Group onlapping the metamorphic basement and has the Pliensbachian-Toarcian Trionto Fm. at its core. At the meso-scale, the backlimb is characterised by intense folding, with S-shaped parasitic folds having axial planes coherently dipping toward the SW. In this Second Tectonic Unit the Paludi Fm. crops out only in the western sector of the basin, where it onlaps the western Jurassic paleomargin (Santantonio *et alii*, 2016), notably granites and/or Middle Jurassic units. A very steep thrust plane separates the Second and Third tectonic Units, placing the Hercynian basement above the Trionto Fm. In the third Tectonic Unit the Paludi Fm:

- crops out along the Colognati Valley in the western sector of the basin (see below);
- crops out in isolated and small outcrops localized on the crests of the hangingwall anticlines, onlapping the Middle Jurassic units in the central part of the basin (Serra Livera);
- is missing in the eastern sector of the basin.

A thrust plane dipping 30° toward the South juxtaposes the Middle Jurassic units in the central sector, and the basement in the eastern and western sector, above the Paludi Fm.

The Third Tectonic Unit is represented by the Longobucco Group unconformable on the Hercynian phyllites. Approaching the thrust plane, the Trionto Fm is overturned or subvertical and is characterised by parasitic S-shaped asymmetric tight folds (Fig. 30) with a south-western subvertical flank and an eastern upright flank mildly dipping toward the SW.

Figure 30: Parasitic folds in the overturned limb of a footwall syncline in the Trionto Fm near Ortiano.

In the third Tectonic Unit the Paludi Fm:

- crops out along the Colognati Valley in the western sector of the basin (see below);
- crops out in isolated and small outcrops localized on the crests of the hangingwall anticlines, onlapping the Middle Jurassic units in the central part of the basin (Serra Livera);
- is missing in the eastern sector of the basin.

A thrust plane dipping 30° toward the South juxtaposes the Middle Jurassic units in the central sector, and the basement in the eastern and western sector, above the Paludi Fm.

In the **Fourth Tectonic Unit** the Cenozoic arenites, in the footwall of the thrust, are intensely folded by metric scale parasitic folds only in the western and central sectors, forming a tight footwall syncline, while in the eastern sector this syncline is missing and only the upright flank crops out, where the Paludi Fm gently dips toward the South (Fig. 31).

In the eastern sector, the Paludi Fm rests unconformably against the Cozzo di Mastro Pasquale Jurassic structural high (Santantonio & Fabbi, 2020) or against the phyllites. This Jurassic high developed as an internally complex footwall block bordered by early Jurassic (latest Sinemurian and

Toarcian) normal faults, facing the basin toward the South, the North and the West. The contact between the Paludi Fm and the Mesozoic runs parallelly to the southern Jurassic escarpment of the Cozzo di Mastro Pasquale high.

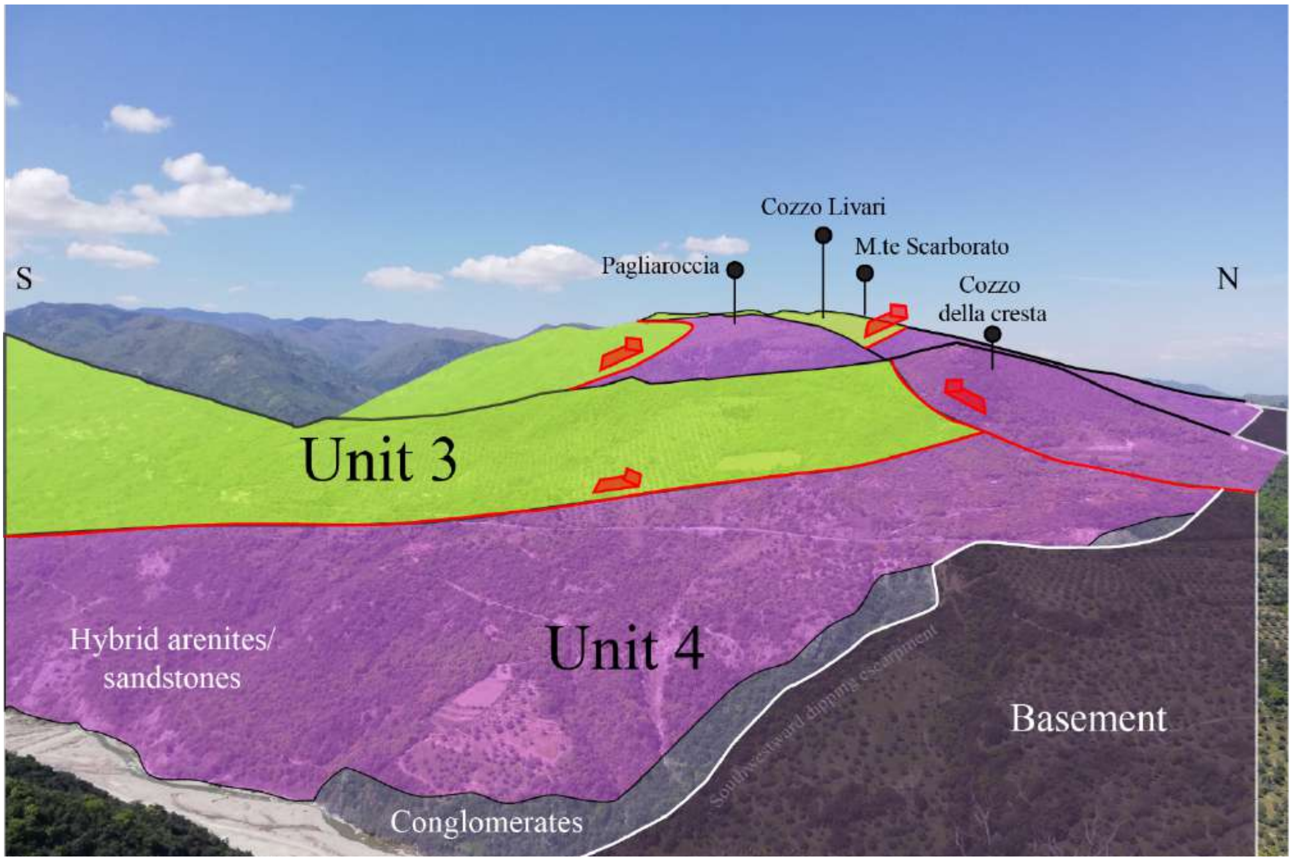

Figure 31: Structural scheme of the lowermost tectonic unit.

In the western sector, the Paludi Fm onlaps the Hercynian metamorphic basement only. A peculiar feature of this tectonic unit is the occurrence of the basal lithofacies of the Paludi Fm, represented by breccias and conglomerates.

CONTACT BETWEEN THE PALUDI FM AND THE SUBSTRATE

As mentioned above, the Paludi Fm crops out extensively in the lowermost tectonic unit and along the Colognati River, in the western sector of the basin.

Here the Paludi Fm onlaps the western Jurassic paleomargin of the basin mentioned above (Santantonio *et alii*, 2016). Both the Jurassic margin and the onlap of the Paludi Fm run parallel to the Colognati River from le locality Il Torno to Rossano for about 10 km. During the present work such margin has been mapped south-westward for an additional 4 km as far as San Pietro in Angaro. Along the southernmost part of the of the Area only sparse Jurassic and Paludi outcrops are preserved, as Quaternary erosion severely affected the area. Along this sector of the basin the Sant'Onofrio Sub

Group (Middle Jurassic – Lowermost Cretaceous) and isolated outcrops pertaining to the Caloveto Group onlap an escarpment carved in the granites (Fig. 32a). Beautifully preserved sedimentary dykes cut the intrusive substrate (Fig. 32b).

This area is structurally complex for three reasons:

i) the presence of a thrust plane juxtaposing the granites above the Meso/Cenozoic succession (both the Paludi Fm and the Mesozoic units);

ii) both the Paludi Fm and the Mesozoic units are buttressed against the Hercynian granites (Fig. 33);

iii) Mesozoic units underwent differential compaction against the Jurassic escarpment.

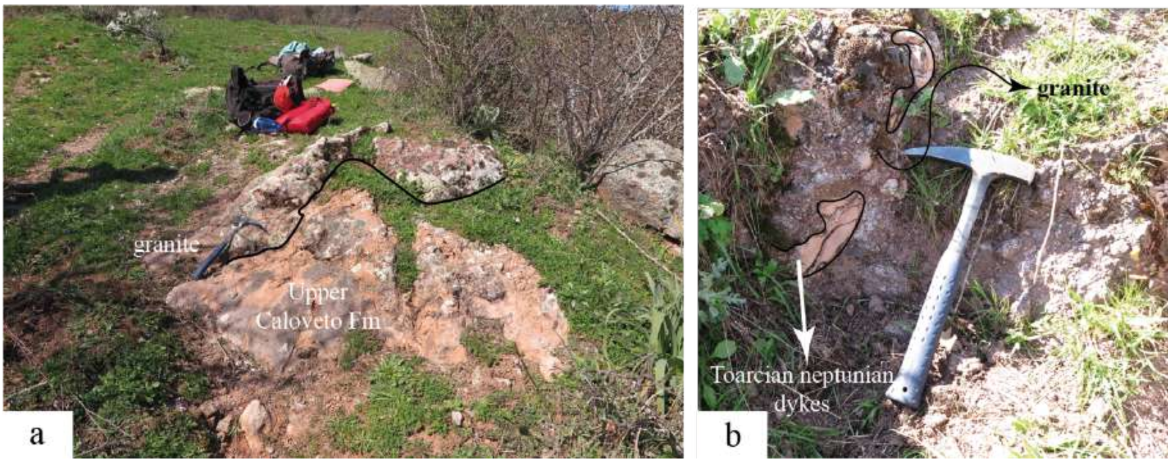

Figure 32: Contact between the Mesozoic succession and the basement at Cozzo del Principe (Colognati River): Upper Caloveto Fm resting unconformably on the granite; b) Upper Caloveto Fm dyke cutting the igneous basement.

Along the Colognati Valley the Paludi Fm onlaps the igneous basement and the Mesozoic sedimentary cover at different stratigraphic levels. The sedimentary units flooring the unconformity are: the Upper Caloveto Fm (Toarcian), Aalenian *Zoophycos* Marls, radiolarian cherts (Bajocian), and the Maiolica Fm (upper Tithonian-Hauterivian) (Fig. 34).

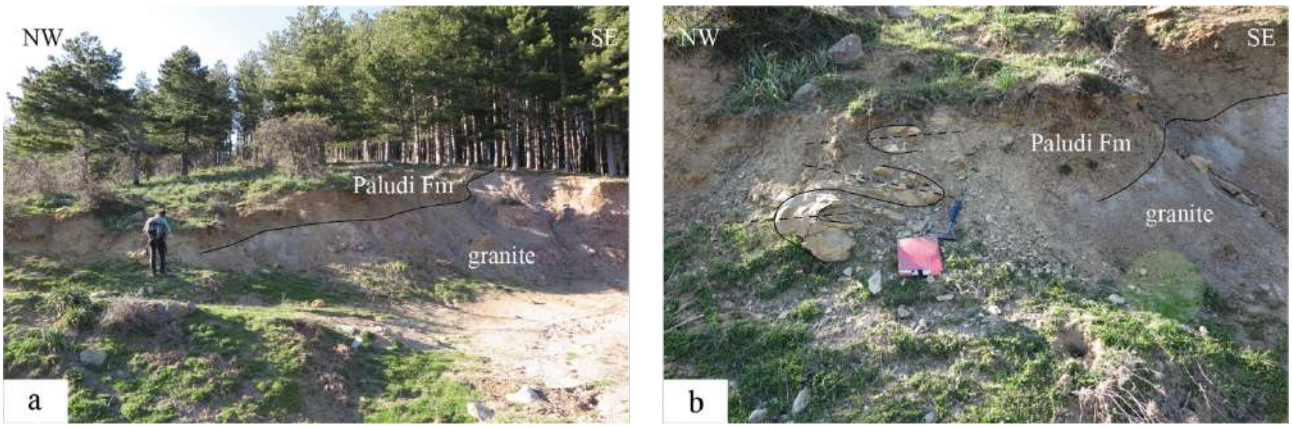

Figure 33: a) Stratigraphic contact between the Paludi Fm and the igneous basement (black line); b) buttressing of the Paludi Fm against the basement. The hybrid arenites are deformed by folds that show axial planes dipping toward the NE, but these geometries are not coherent with the expected vergence (toward the NE).

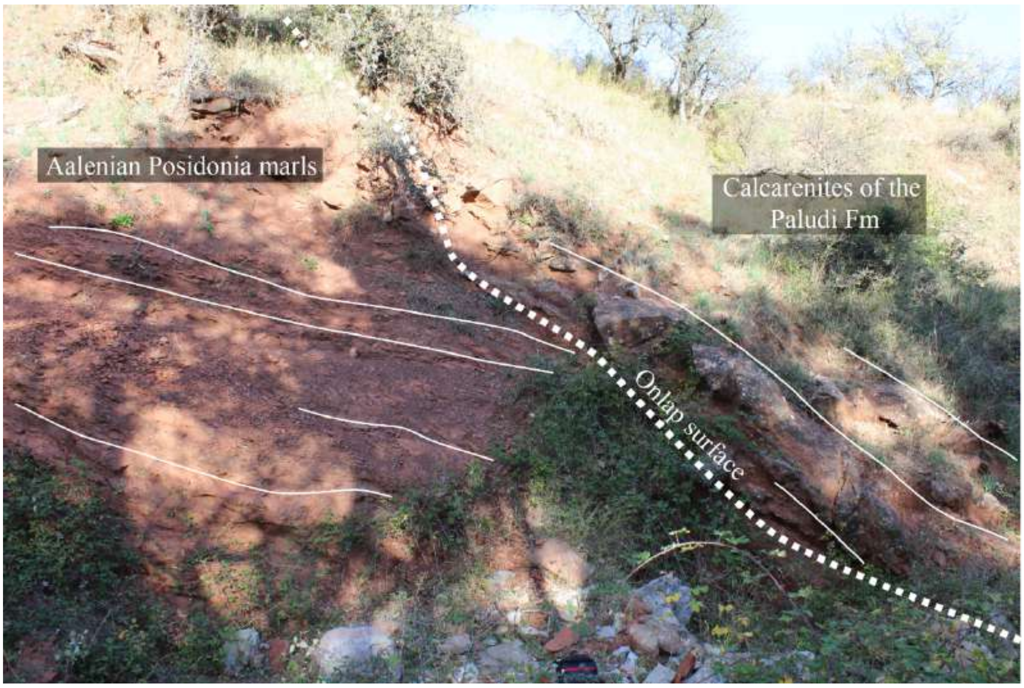

Figure 34: Stratigraphic contact between the third lithofacies and the Jurassic succession (Il Torno).

All along the unconformity surface, Larger Benthic Forams and microforaminifera were systematically sampled, and their assemblage confirmed the Ypresian-Lutetian age of the unit. In order to test the consistency of the various lithofacies among the different tectonic units, "Section 9" was measured in this area. Here the first (reddish marls) and third lithofacies (hybrid arenites) of the Paludi Fm are exposed, being the megaclastic deposits of the first lithofacies completely missing.

As highlighted in the tectonic units description, besides the outcrops of the Colognati Valley, in the second tectonic unit the Paludi Fm is exposed only at Serra Livera. This is the crest of the hangingwall anticline of the second tectonic unit. Such structure is characterised by the unconformable contact of the Trionto Fm above the Hercynian phyllites. This contact is marked by the presence of condensed micritic patchy deposits with a very limited areal extent, resting unconformably on the metamorphic basement. Such deposits have been interpreted as epiescarpment deposits (*sensu* Santantonio, 1993) (Fig. 35).

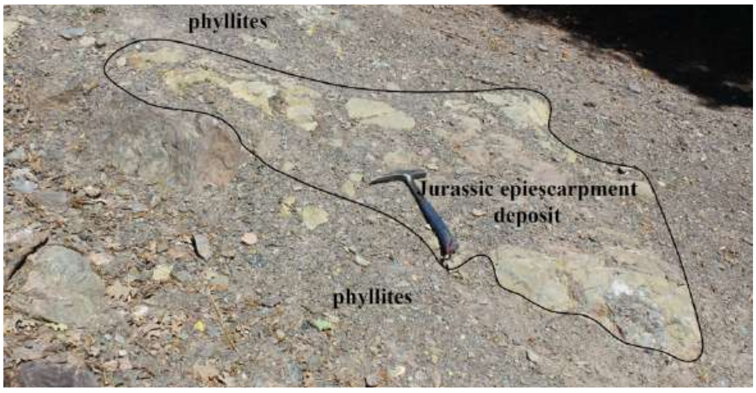

Figure 35: Patchy veneer of epi-escarpment pelagic deposits along the Coserie Jurassic escarpment, resting unconformably on the phyllites.

The southward continuation of this paleoescarpment has been mapped during their survey by M. Santantonio and S. Fabbi for about 5 km. Both the thin condensed deposits and the basement are covered by the Trionto Fm, that stratigraphically passes to the Gray Limestone with black chert. The top of Serra Livera is characterised by the passage to the overlying radiolarian cherts. Here the Paludi Fm unconformably covers the Jurassic succession with a low angle unconformity (Fig. 36).

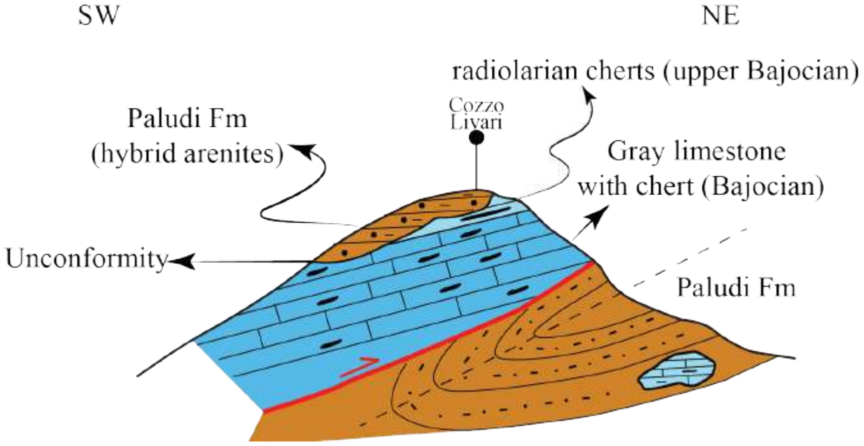

Figure 36: Geological section along Cozzo Livari. Note that in the footwall of the thrust fault radiolarian cherts megaclasts are embedded within the Paludi Fm.

In the lowermost tectonic unit, the Paludi Fm crops out along a narrow belt, 20 km long, and is severely shortened. East of the Trionto Valley, the thrust plane that separates the first from the second tectonic unit can be observed in two localities: along the Laurenzana Valley and in the eastern side of the Trionto River. In both these localities calcite veins have been sampled for U/Pb dating. Along the Laurenzana Valley (Fig. 37), where the structure cuts a marly interval, poor foliation and widely spaced S-C fabrics are developed. In the Trionto Valley, the thrust plane deforms the hybrid arenites, which are not deformed by S-C fabrics or by folding. Also, veining is very limited, which made the sampling for U/Pb dating difficult.

The thrust plane in this sector dips consistently about 30° toward the South, with the phyllites overriding the Paludi Fm. At the footwall of the thrust the Paludi Fm is not overturned, dipping regularly toward the South/South-West. Along the Laurenzana Valley the Paludi Fm is abruptly cut by an high angle (sub-vertical) fault, juxtaposing the basement to the Cenozoic unit.

West of the Trionto Valley the structural setting is more complex, as the footwall of the thrust plane is characterised by a very tight footwall syncline with an overturned backlimb. Both the backlimb and the forelimb display common parasitic folds. The axial planes of parasitic folds dip consistently toward the South/South-West. This has made it impossible to measure any long stratigraphic section in this sector of the basin. Another difference with the Trionto Area regards the nature of the juxtaposed units:

- In the eastern sector the basement constantly overthrusts the Paludi Fm;

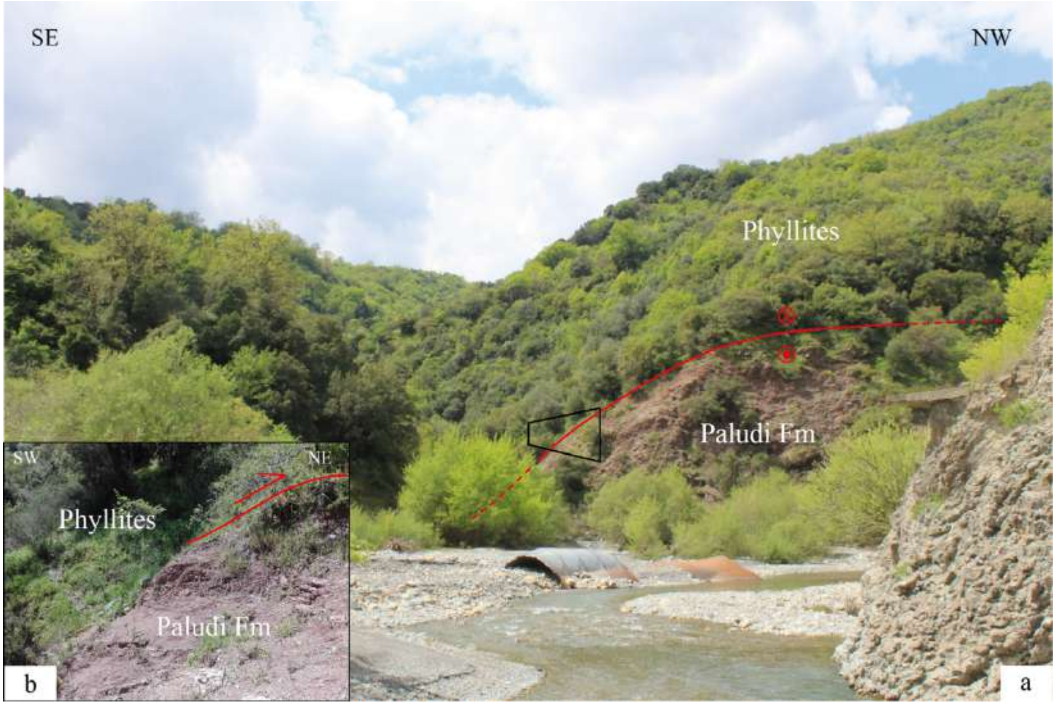

Figure 37: Thrust plane outcropping along the Laurenzana River.

- In the western sector of the basin the stratigraphic boundary between the radiolarian cherts and the Gray Limestone with chert overthrusts the hybrid arenites. In some localities (e.g., Vallone Sant'Elia) the tectonic contact is between the limestone and the Paludi Fm, while in other localities (Santa Maria ad Gruttam, Vallone Scarborato) it is between the radiolarian cherts and the Cenozoic siliclastics (Fig. 38).

Figure 38: Tectonic contact between the Upper Bajocian Grey Limestones with black chert and the Paludi Fm (Vallone Scarborato).

It is interesting to note that in both these localities the Paludi Fm, localised in the footwall block of the thrust plane, bears megaclasts (up to several tens of metres across) of the radiolarian cherts.

Concerning the basal contact, as we have already noted, the Paludi Fm onlaps:

- The Cozzo di Mastro Pasquale Jurassic structural high in the Laurenzana Valley;
- The Hercynian phyllites from the Trionto Valley to Paludi. West of the Colognati River, the contact with the basement is often covered by the Late orogenic siliciclastic deposits. These latter deposits dip gently toward the North. Especially between the Coserie River and Vallone Sant'Elia, the Paludi Fm forms a very narrow belt, with a thickness of few tens of metres, being overthrust by the Jurassic units while toward the North they are covered by the post-Tortonian deposits.

From the above the it emerges that at least along the Colognati Valley and at Cozzo di Mastro Pasquale, the Paludi Fm is unconformable on, and therefore re-utilizes, the Jurassic rift basin paleomargins. This makes the Mesozoic Longobucco Basin and the Cenozoic Paludi Basin as one welded basin (Fig. 39).

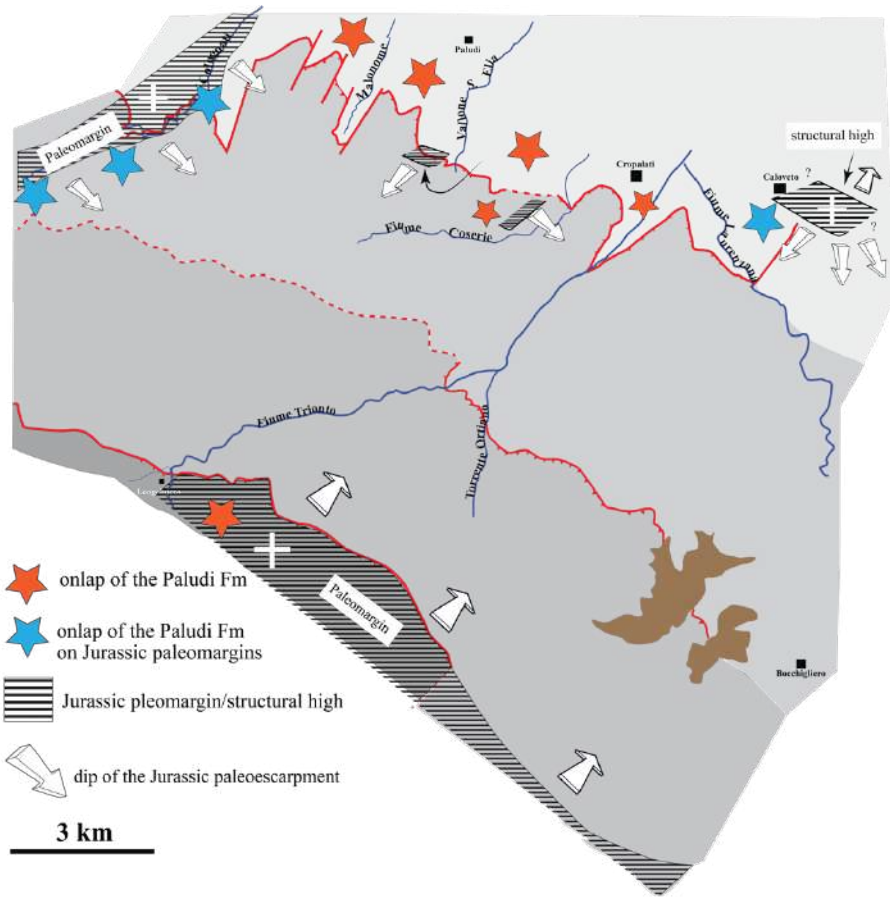

Figure 39: Relationships between the Paludi Fm and the Jurassic submarine paleotopography.

Stratigraphic sections and sedimentology of the Paludi Fm

The concept of "Turbidite" was developed between 1940 and 1950 thanks to the studies of Migliorini (1943) and Keunen & Migliorini (1950). In those times the concept of turbidite and turbidity current was in its infancy, and it was essentially put in relation with the development of a foredeep in a collisional context.

In 1962 Bouma proposed the first depositional model for this kind of deposits, describing the ideal internal architecture of a turbiditic interval. At the same time, studies addressing the emplacement mechanisms and hydrodynamic processes, linked with a turbiditic flow, flourished. These studies led to the development of a more complex scenario than previously imagined. In particular, it was demonstrated that the basal massive and erosive intervals of an ideal "Bouma sequence" were the product of a dense flow and not of a turbiditic one (Sanders, 1965; Hampton, 1972). This led to the development of the concept of bipartite flows (Sanders, 1965), in which a basal dense flow is overlain by a more diluted turbulent flow. The two flows and their respective deposits are clearly genetically linked. This is mirrored in the difficulties and controversies, found in the literature, in recognising the relative importance of the two processes in accumulating sandstones within deep-water basins (Shanmugam, 1996). This has resulted in a proliferation of new nomenclatures aimed at describing flow types and related products. For the sake of clarity, we will use the term "turbidity current" in the sense of Kuenen (1965), considering indeed both the basal dense flow and the upper turbulent flow. This choice, as suggested by Mutti *et alii* (2003), is dictated by the fact that deposits generated by these two distinct flow types are genetically linked. The term "dense flow" is here used to describe "highly concentrated mixtures of sediment and water" (Mutti *et alii*, 2003), while in the literature other synonyms are: "high density turbidity current", "debris flow", "sandy debris flow", "hyperconcentrated flow" ecc.

It then became also evident that the Bouma sequence was not sufficient to describe exhaustively turbiditic deposits. Consequently, since the '70s, a number of Authors proposed schemes of turbiditic facies associations in order to describe accurately a turbiditic deposit. To this day, as a number of classifications are used for the facies analysis of turbidites, reference will be made here to the one of Mutti *et alii* (2003), which describes the down-current evolution of a waning and depletive turbidity current (Fig. 40).

In the classification of these Authors, two are the fundamental characteristics to be considered: grain size and sedimentary structures.

Concerning the first category, Mutti *et alii* (2003) identify four grain-size populations characteristic of a specific flow type (dense flow vs turbiditic flow) and of each distinct sector of the depositional system.

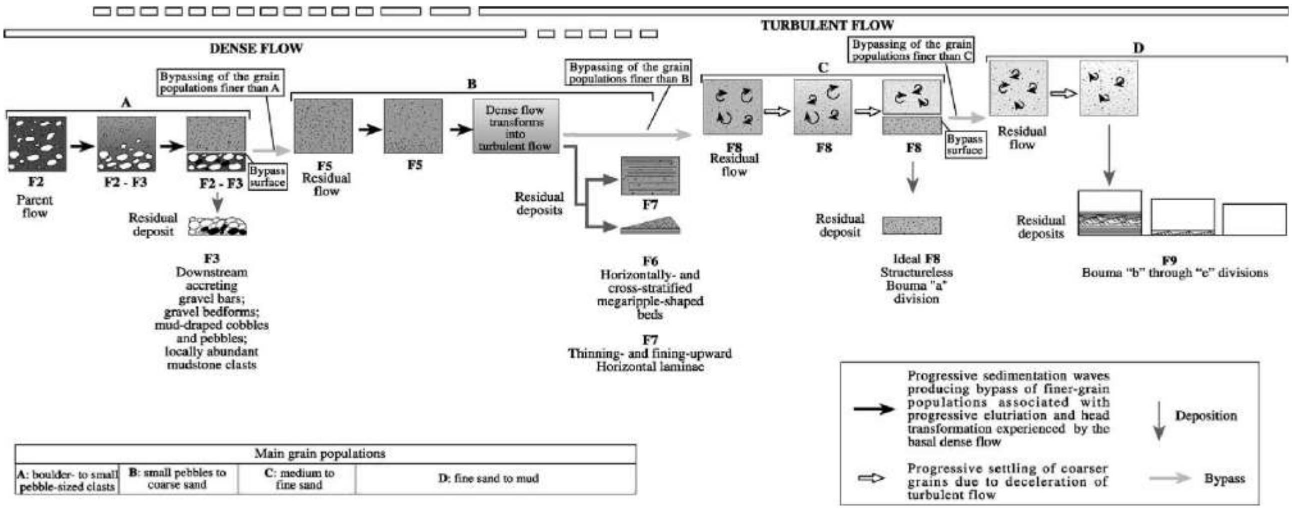

Figure 40: Ideal down-current facies organization of a waning and depletive bipartite turbidity current (after Mutti et alii*, 2003).*

Grain size populations are:

A: from boulder to small pebbles

B: from small pebbles to coarse sands

C: from middle sand to fine sand

D: from fine sand to mud

The coarser grain sizes (A and B) are transported and deposited by the basal dense flow. The dense flow is considered to be under conditions of excess pore pressure and the deposition of the coarser grain sizes is related to its gradual dilution and deceleration. In the scheme of Mutti and co-authors the basal dense flow is further divided in a “gravelly flow” and a “sandy flow”. This latter, after the freezing of the former, bypasses it being capable of greater runout distances. The gravelly flow is considered to be the “parent flow” and the other flows are all residual.

The turbulent flow, that initially moves more slowly than the basal one, bypasses the dense flow transporting the finer grain-size material. It must be noted that the C grain size population can be initially transported by the dense flow and then incorporated within the turbulent flow, while the fourth population (D) is instead typical of a fully turbulent flow.

The facies recognised by the Authors (Fig. 40) in both the dense flow and the turbulent flow are:

F2: matrix-rich conglomerates, disorganised and poorly sorted;

F3: clast-supported conglomerates;

Both these facies are the product of the parental basal gravelly flow, that is characterised by a high erosive power as highlighted by the presence of common rip-up clasts up to m-scale boulders. The F2 facies in particular is typical of hyper-concentrated flows.

F5: coarse sands arranged in massive, poorly sorted, thickly bedded intervals, sometimes graded. Dewatering structures indicate that deposition took place under overpressure conditions.

These deposits are both the products of the sandy dense flow bypassing the gravelly dense flow, or of the trailing edges of the gravelly dense flow.

F6: coarse sands characterised by megaripple stratification;

F7: crudely horizontally-laminated beds characterised by coarse to medium sand; Laminae are fining and thinning upward within individual beds.

Both these facies represent the bypass of the turbulent flow when the dense flow decelerates, which can take place with low or high rates. In the first case a passage from F5 to F7 is expected. In the second case the F5 facies is overlain by the F6 facies. In both cases the upper turbulent flow is not capable of fully suspending the coarser grain sizes (B) carried by the dense flow, so that they are deposited and reworked by the upper turbulent flow into tractive structures.

Since now onward, sedimentation is dominated by processes related to turbulent flows and is well described by the Bouma sequence. In particular:

F8: is a graded interval represented by fine sands and can be compared to the Bouma “a” division;

F9: are the “b” through “d” Bouma divisions represented by fine sand to silt and characterised by sedimentary structures related to traction and fallout processes, thus characterising the final stages of a turbidity current.

As proposed by Tinterri & Tagliaferri (2015), beds are then divided into groups or “bed types” in order to relate them to the dynamic conditions of the current.

An important aspect that was underestimated in the work of Mutti *et alii* (2003) is that of Mass Transport Deposits (MTDs). According to their definition (see Posamentier & Martinsen, 2011), MTDs include only those deposits where sediments are moved *en masse.* When several MTDs are present within the same lithostratigraphic unit we can refer to them as Mass Transport Complexes (MTCs). The main processes leading to MTDs are: creeping, sliding, slumping and debris flow. Rockfall deposits and turbidites are not to be considered MTDs. All the above-mentioned processes are often acting together as a “continuum”. In fact, one process can evolve into another or alternatively trigger another one. As already mentioned, the genetic link between debris flow and turbidites was understood since the ‘60s. However, besides debrites, turbiditic successions are known to embed chaotic intervals and structures related to the action of Mass Transport Processes, such as:

amalgamation surfaces, delamination structures, sand injections, boudinage, slumps and soft sediment deformations. Two are the main processes that lead to the development of these features:

i) Strong erosional power: during their downslope motion MTDs erode the substrate incorporating part of it in their bodies, but erosional features may develop also during the emplacement phase, due to deceleration. These features can be delamination structures, amalgamation surfaces, sand injections etc. These intervals are known as “slurry bed” or “hybrid bed” in the distal parts of depositional systems (see Fonnesu *et alii*, 2016), but are recognised also in the more proximal parts. As highlighted by Tinterri & Tagliaferri (2015) and Tinterri & Piazza (2019), these erosional features are very important in identifying a distinctive part of the tubiditic system and a specific dynamic condition of the flow, as they are indicative of the deceleration of the basal dense flow and consequently of the bypass of the upper turbiditic flow. The identification of such structures is fundamental also in assessing morphological confinement within a basin, as the deceleration can be produced by morphologic obstacles (morpho/structural highs at any scale). Due to the above, we will follow the methods introduced by Tinterri & Piazza (2019) and we will consider the intervals presenting impact structures as distinct with respect to those that do not have them (see chapter “Bed Type”).

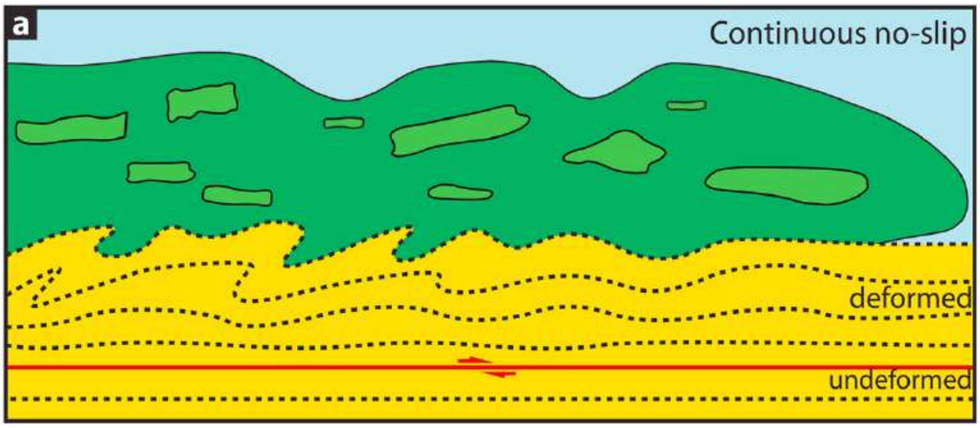

Figure 41: Deformation of the substrate due to transfer of shear stress (Sobiesiak et alii*, 2018).*

ii) Deformation of the substrate during downslope motion of MTDs. During their downslope motion, MTDs can transfer stress to the substrate, generating soft sediment deformation and chaotic intervals *s.l.* (Sobiesiak *et alii*, 2018) (Fig. 41).

The penetration of shear stress does not always occur, as liquefaction or hydroplaning processes can impede substrate deformation. The distinction between slump *s.s.* (and in general soft sediment deformation related to post depositional processes) and chaotic intervals can be sometimes tricky. As highlighted by Ogata *et alii* (2012), substrate deformation can range from progressive disruption of sedimentary structures, to dismembering of sandy interlayers with development of soft-sediment features, partial mixing between sandy and marly interlayers, and eventually complete stratal disruption.

Chaotic bodies and mass transport deposits (MTDs) have been described in this Thesis following Ogata *et alii* (2014) (Fig. 42), who recognised that deformations observed within plastic (still unlithified) clasts, related to gravity driven deposits, resemble those generated by tectonics. Consequently, these structures are described using the classification of Passchier & Trouw (2005), with the addition of the suffix "pseudo" (e.g., pseudo-SC structures).

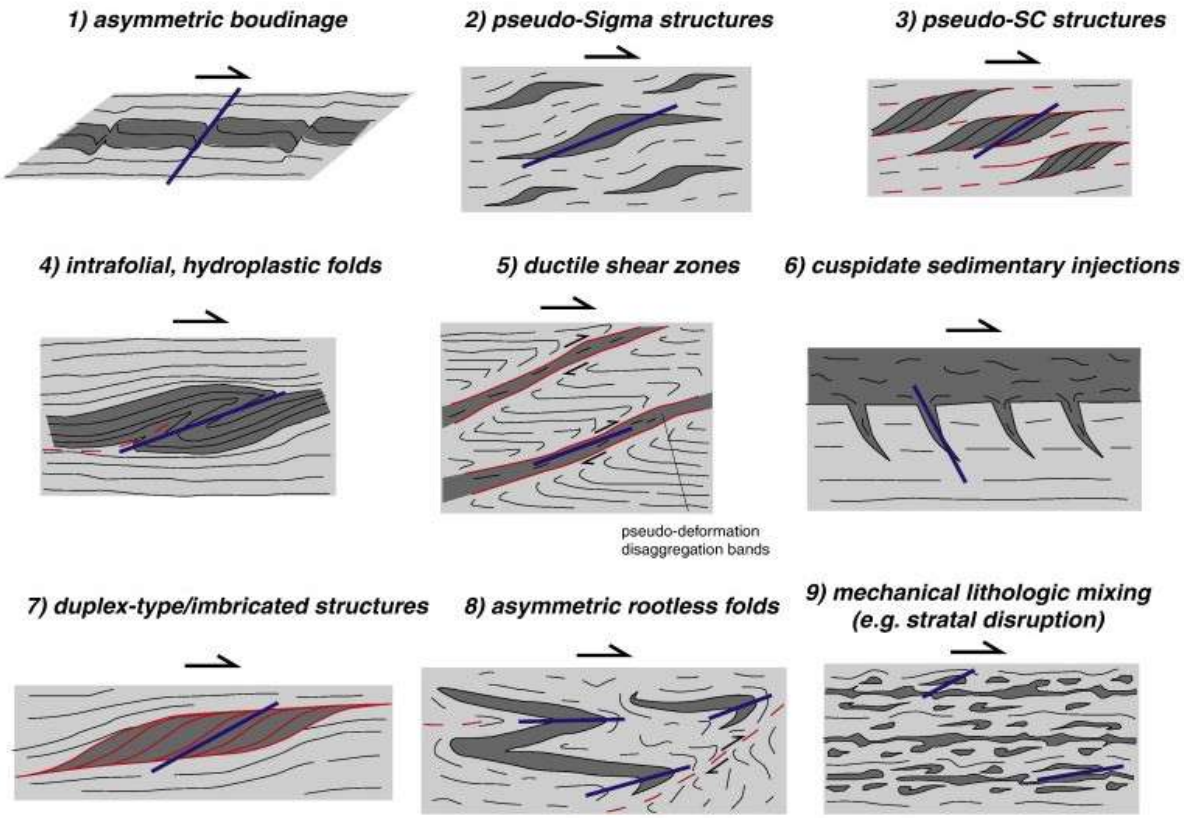

Figure 42: Nomenclature proposed by Ogata et alii *(2014) to describe ductile structures developed by MTDs.*

BED TYPES

The problem of existing models

We typically try to build schemes and models aimed at describing any kind of depositional system and consequently we try to trace a given studied outcrop back to an existing model. The use of universal models is somewhat reassuring, as they are predictive and can guide us to the understanding of basin architecture, facies distribution etc. As stated by Reading & Richards (1994), for thirty years the scientific community tried to classify deep-water turbiditic systems less successfully with respect to other depositional systems. Thus far, the concept of turbidite and turbiditic system still remains linked to that of unconfined elongate foredeeps, fed by deltaic systems through single or multiple point sources, where flows can spread radially generating the well-known "submarine-fan" geometry. Such systems are characterised by discrete, standard elements such as lobes, canyons, channels etc. This is clearly due to the fact that most of the studied fossil examples of turbiditic systems are represented by foredeep deposits. As we will discuss later, our Paludi basin is a far cry from a "canonical" turbiditic basin.

Taking into account all the above, we elected to use the model proposed by Mutti *et alii* (2003), as it addresses all the processes potentially effective in a deep-water basin. In doing so, however, we were fully aware that we are dealing with an "unconventional" system, so the facies association could in some way depart from the model.

Reference to facies analysis studies of confined basins (see Tinterri & Piazza, 2019 for an example) was essential, as we found analogies with the Paludi basin. In these basins the seafloor topography severely affects flows, hindering their downstream path and thus preventing the development of the "ideal" lobe- and fan-shaped depositional bodies. Clearly also the facies associations are different from those predicted by canonical models. In particular, confinement promotes deceleration of flows, hydraulic jump and bypass of the turbulent part of the flows, until the morphological obstacle is blanketed. The result of such processes is the presence and local dominance of massive and coarse-grained intervals (F5), impact structures, and coarse-grained tractive structures (F6 and F7 of Mutti *et alii*, 2003).

In papers describing confined basins, Bed Types are grouped similarly, so we decided to use as a reference scheme the one proposed by Tinterri *et alii* (2017), as it represents the most similar field example to the Paludi basin. The following description of Bed Types is ordered by decreasing grain size, and attempts to frame each Type within the facies classification of Mutti *et alii* (2003) (Fig. 43).

Type 1 Beds

This category can be further subdivided in three sub-categories.

Type 1a: These thick (> 5 metres) beds form poorly sorted, matrix-poor breccias with clasts ranging in size from small pebbles to large boulders (F3 facies). These beds are the product of the basal dense flow, and in particular of the gravelly dense flow.

Type 1b: these beds are made of very thickly bedded, matrix-rich conglomerates, where clasts are represented by basement/Mesozoic sedimentary cover rocks and remarkably by plastically deformed elements pertaining to the Lithofacies 2 and 3 of the Paludi Fm. (F2 facies). Type 1b beds can be either capped by very thickly bedded, poorly sorted, polymictic matrix-poor breccias (F3) or by highly deformed megablocks (bed-stacks) pertaining to the substrate.

The F2 facies is the product of the body of an hyperconcentrated flow, characterised by a strong erosional power ("bulldozer effect"), that deeply erodes the substrate. The largest blocks cannot be incorporated within the flows, so they are transported at its front, which is testified by the presence of the oversized megablocks at the top of the F2 facies. Similarly, the F3 facies is sustained by the hyperconcentrated flow and transported.

Type 2 Beds

These beds can be further divided in two sub-categories:

Type 2a: Thickly bedded breccias (F3) capped by pebbly sandstones/very coarse/ coarse sands arranged in massive and structureless beds (F5). Bimodal beds, characterised by abrupt jumps in grain size (F3 capped by F5) are interpreted as the product of the bypass of a flow, and they can be produced by the bypass of a sandy dense flow (residual) over the gravelly dense flow (parental).

Type 2b: pebbly sandstones/very coarse sands arranged in massive and structureless beds that can be thin (exposure along the SP177) or very thick, or coarse sands arranged in structureless beds. Both these facies are the product of a sandy dense flow and are associated respectively with the leading edges and the body and tail of such flows.

Type 3 Beds

Coarse to very coarse, structureless sands (F5) overlain by tractive structures that can be represented either by horizontally laminated beds with laminae that can fine and thin upward (F7), or by cross-stratified beds (F6). F6 and F7 can be capped by mud-drape scours indicating the complete by-pass of the sediment transported by the turbulent flow.

Type 4 Beds

Coarse structureless sands, capped by massive medium sands that pass upward to laminated fine sands.

Type 5 Beds

As suggested by Tinterri & Piazza (2019) beds characterised by impact-related structures must be analysed separately because they are indicative of the deceleration phase of the flow. Type 5 beds are therefore referred to the previously described bed types wherever they exhibit impact and delamination structures. As suggested by Tinterri & Piazza (2019), these beds will be divided in sub-groups indicating the impacting bed: for example, Type 4-1a are those beds referred to the Type 1a that display interaction with the substrate, and so on.

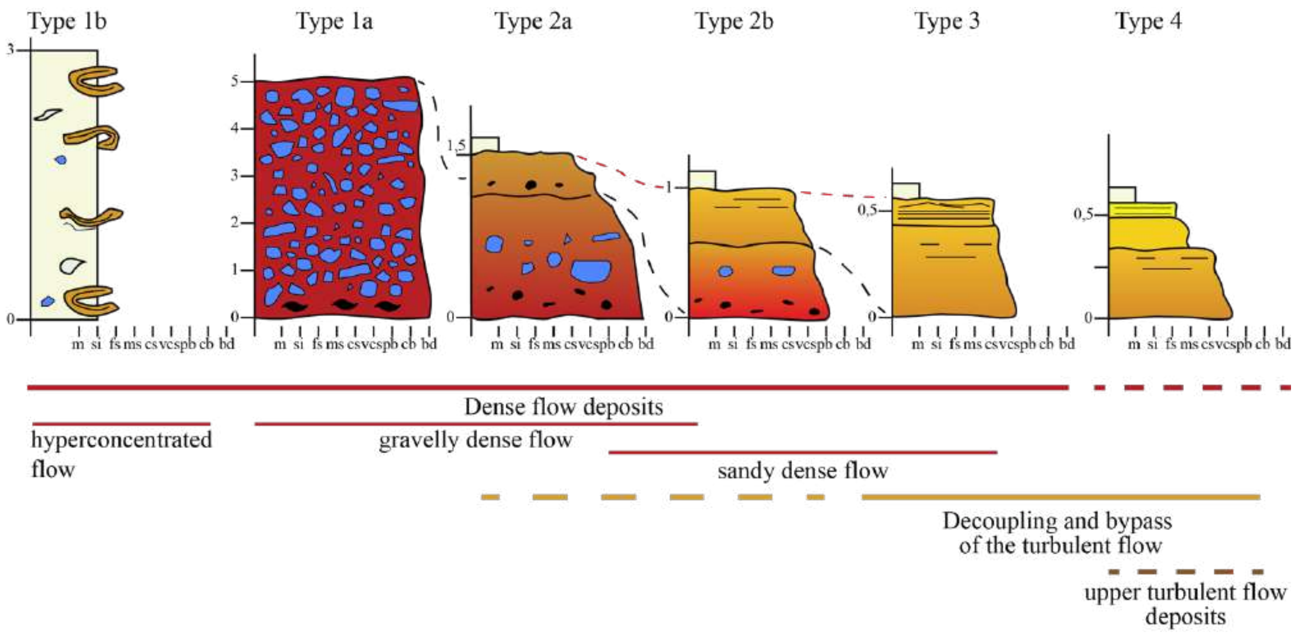

Figure 43: Summary of Bed Types and related processes. The Bed Types are ordered in a down-current direction.

STRATIGRAPHIC SECTIONS

In this chapter, the sedimentological descriptions of individual stratigraphic sections will be framed in the Lithofacies subdivision (1 to 3) of the Paludi Fm. proposed for the present Thesis (Fig. 44).

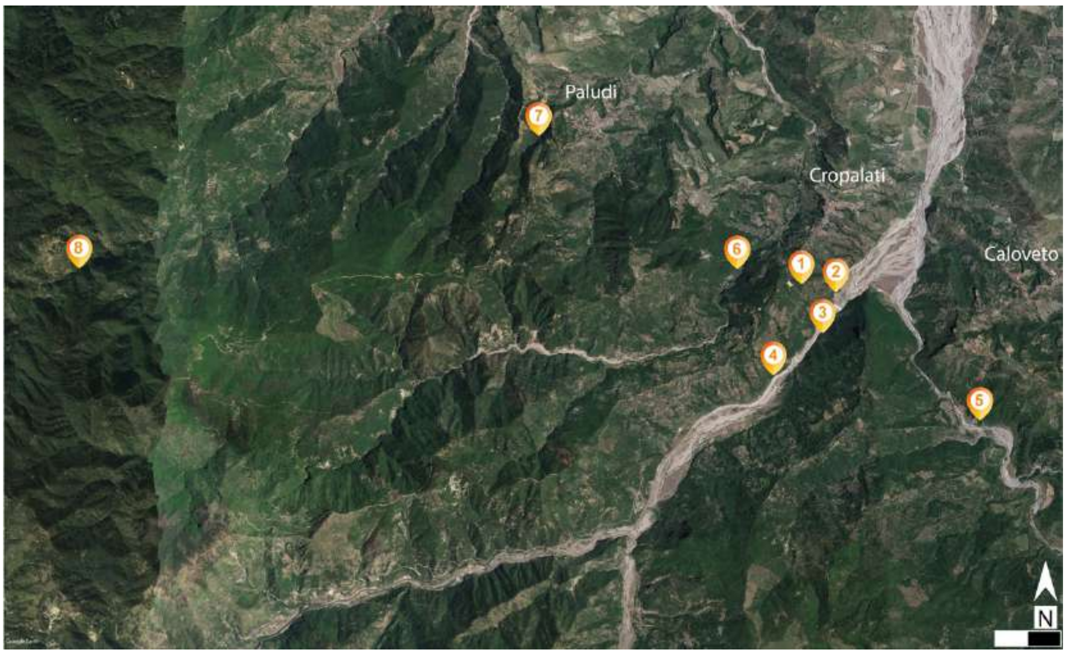

Figure 44: Location of the measured sections.

Section 1

This section has been measured along the SP 177 road connecting the villages of Cropalati and Destro, between the Km 58 and 59, for a total thickness of 28.41 m. It comprises deposits ascribed to Lithofacies 1 and 3 of the Paludi Fm., as defined in this Thesis. It is located at around 400 m of altitude and it is parallel to the Trionto Valley section. Bedding attitude is 180/40 and the section runs parallel to the bed-dip (about N-S).

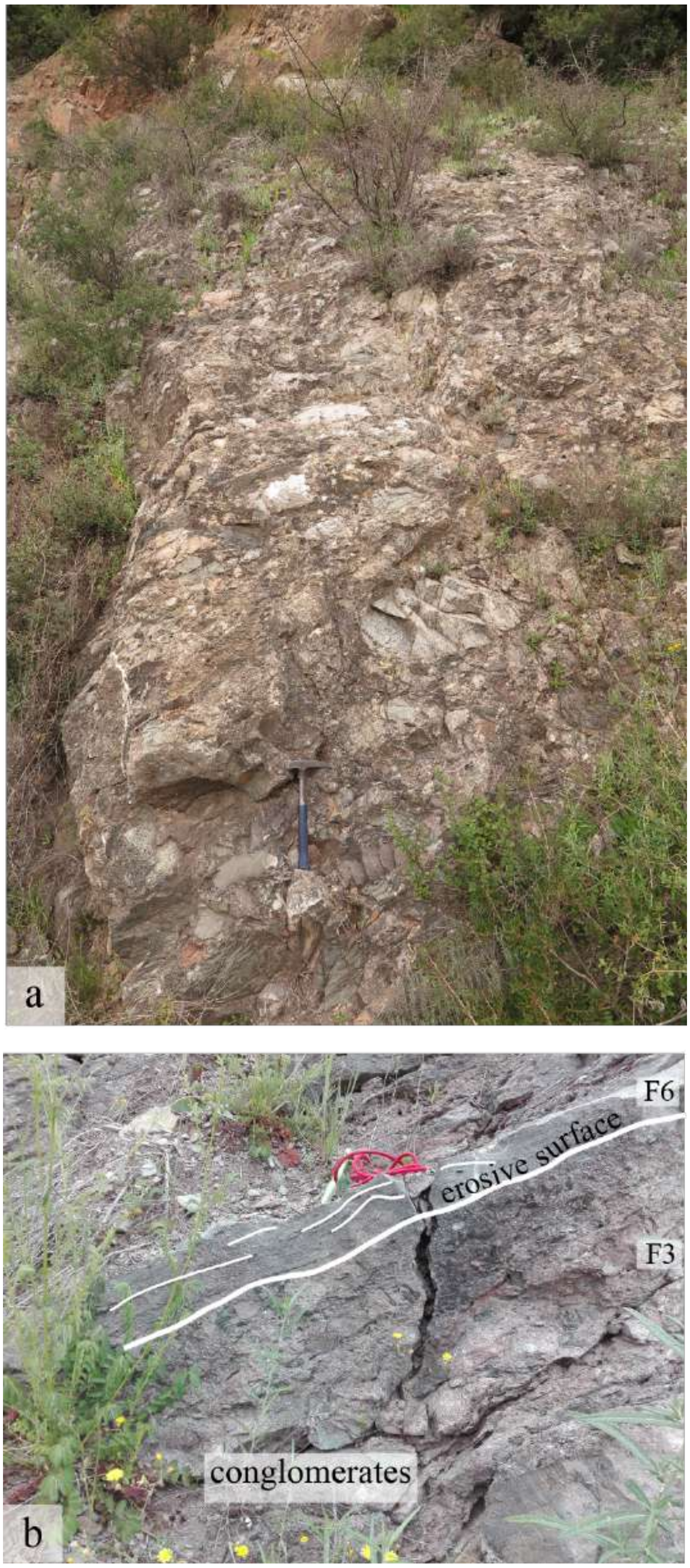

Figure 45: a) general appearance of the F3 facies, represented by poorly sorted conglomerates; b) bimodal bed characterised by a sand rich conglomerate (F3) passing upward, throught a sharp erosive surface, to a coarse sand characterised by wavy laminae (F6).

At Km 59 the contact between the basement and the arenaceous sandstones is exposed.

Overlying the poorly exposed calcarenite interval, from bottom to top we observe:

- 9.80 m of poorly sorted (small pebbles to large boulders, with longer axis up to 1m), massive conglomerate (Fig. 45a) s. Clasts are made of abundant angular phyllites, sub-rounded granite and quartz, and less common Lower Caloveto Fm. limestone. Clasts are densely packed in a coarse sand matrix. This bed is capped by a very thin, laterally discontinuous drape made of reddish silty marls;

- 7 m: very coarse sands/pebbly sands massive and structureless

- 1.15 m: poorly sorted (small pebbles to large boulders, with longer axis up to 30 cm), massive conglomerates in a silty matrix. Beds are draped by thin silty marls.

- 1.77 m: two markedly bimodal beds (1.15 m and 0.62 m thick, respectively), with a conglomeratic base (1 m and 58 cm respectively) and an arenaceous top (15 cm and 4 cm), the contact between the two being sharp and erosional. The conglomerates are moderately sorted, with clasts ranging from coarse to small pebbles with sporadic large cobbles, pertaining essentially to the basement. The tops are characterised by coarse and massive sands with scattered fine pebble-sized clasts pertaining to the basement (Fig. 45b). The top of the second interval is characterised by wavy laminae. In the hybrid arenites Larger Benthic Forams (LBF) have been detected. Since now onward essentially all the arenaceous beds contains such fossils.

- 60 cm: moderately sorted massive very coarse pebbly sands with clasts ranging from coarse to small pebbles pertaining to the basement;

- 20 cm: reddish marls, intercalated with very thinly bedded (1 to 2 cm) sand- rich conglomerates (clasts up to 2 cm);

- 20 cm and 40 cm: two structureless and ungraded beds of yellowish very coarse sand/pebbly sands, with phyllite clasts up to 1 cm across;

- 60 cm: chaotic interval made of reddish/greyish marls, with broken and plastically deformed arenitic beds up to 20 cm thick. Marls are deformed due to differential compaction, sand injections are present;

- 45 cm: very thinly bedded massive and structureless coarse sands intercalated with thin discontinuous reddish marls. Only one of these arenitic bed has a crudely laminated top;

- 2.14 m: chaotic interval, reddish/grey marls with sand injections and plastically deformed arenaceous centimetric beds;

- 18 cm: very coarse pebbly sands;

- 2 cm and 6 cm: coarse structureless massive sands, the second bed shows an erosive base;

- 60 cm: fine pebbles evolving upward to very coarse sands;

- 1 m: centimetric alternations of very coarse sands and grey/reddish marls;

- 40 cm: massive and structureless coarse sands;
- 1. 80 m: centimetric alternations of very coarse sands and grey/reddish marls;
- 53 cm: massive and structureless coarse sands.

In this section the passage between the first and third lithofacies can be observed. This is not a sharp passage as we will see below. This section exhibits a general fining and thinning upward trend. Beds are initially massive, with clasts up to boulder-size, while toward the top they decrease both in thickness (thin to very thin) and grain size (coarse sand). This general trend can also be recognised at a smaller scale, in particular two fining upwards trends can be recognised.

The first cycle is represented by 9.80 meters of conglomerates followed by the about 7 meters thick interval represented by very coarse sands/granules with scattered pebbles. This interval is represented by Type 1a and Type 2b beds respectively. These beds represent the product of gravelly dense flow followed by sandy dense flow. The basal conglomeratic stack shows at its top a mud-draped scour, indicating by-pass.

The second fining upward trend is represented by a second conglomeratic interval (the two bimodal beds) followed by the arenites. The first 10 metres and the bimodal beds represent two major pulses of catastrophic clastic sedimentation, due to collapse of basin margins. As we will see later, both the clastic intervals are herein under-represented with respect to the sections measured downcurrent.

Bimodal beds (Fig. 45b) are a peculiar example of Type 2a beds indicating a jump in flux dynamics, being their base representative of a gravelly dense flow, while their top indicates the by-pass of the sandy dense flow. The sandy deposit displays wavy laminae (F6), suggesting severe decrease in the velocity of the flow. In this example the MTDs, represented in this setting by the conglomeratic deposits, produce a confinement to the flow due to their convex-upward shape, forming morphological obstacles capable of dissipating the energy of the flow.

Hereafter only bed type 2b has been detected, as coarse sands prevail with respect to very coarse sands/pebbly sandstones in the upper part of the section. Therefore, the intermediate part of the section is represented by deposits of the leading edge of a sandy dense flow, while the uppermost part of the section is the product of the body and trailing edge. This interpretation can be corroborated by the predominance of coarser grain sizes (pebbly sandstones) in a down current position with respect to this section (see section Trionto). Consequently, we can assume that the depositional system was shifting toward the south, as the depocentral areas were gradually being infilled (Fig. 46).

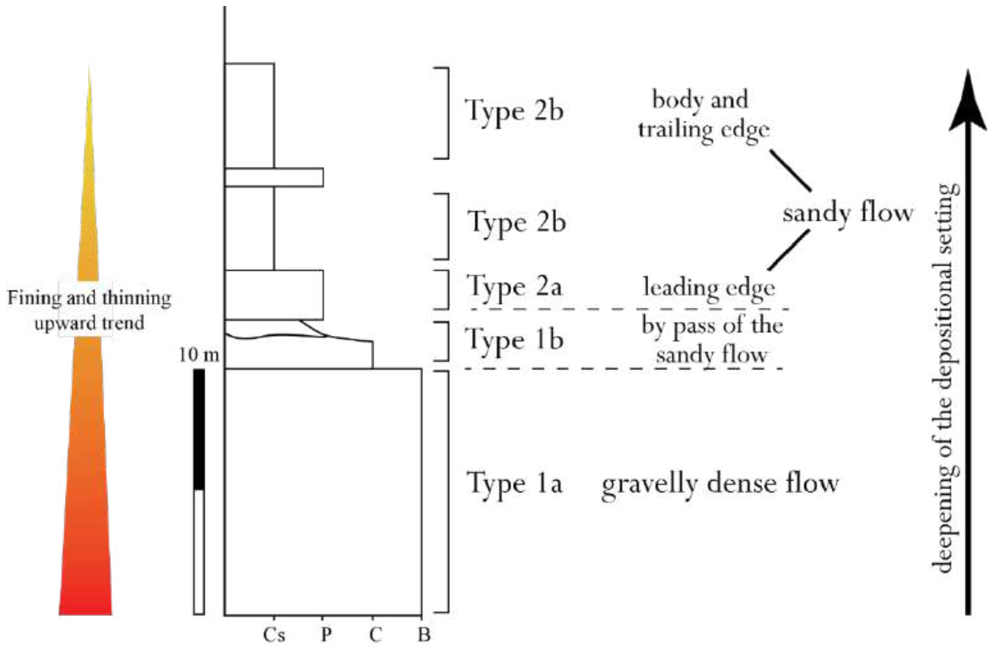

Figure 46: Scheme representing the trend recognised in the SP177 section. Beds ascribable to gravelly dense flow at the base are followed by deposits produced by the leading head and the body/trailing edge of a sandy dense flow.

In this section impact related structures are essentially missing, while chaotic intervals characterised by plastically deformed very thin coarse sand beds in a marly matrix are present. They can be interpreted as slumps or as post depositional deformation produced by the action of down-streaming flows. Post depositional fluidization/liquefaction processes (liquidization *sensu* Ogata *et alii*, 2012) can also be invoked to explain the genesis of these intervals. As such deposits are under over-pressure conditions, fluid escape could also occur in a post depositional phase, deforming discrete intervals.

Section 2

- 3,60; 5,80; 2,20 m; 4,35 m: clast supported, poorly sorted breccia, with angular and randomly oriented clasts pertaining to the basement and to the Caloveto Group and Sant'Onofrio sub-group, sub-rounded clasts of Qz, granite and Lower Caloveto Fm. limestone. Clasts range in size from few cm's to 50 cm. Matrix is virtually absent, being represented essentially by granules (Fig. 47);
- 27 m: vegetation covers this interval, making it difficult to measure precisely the thickness of each bed. From a qualitative analysis breccias appear as those previous described;

Figure 47: General appearance of the matrix- poor breccias outcropping along the Trionto River.

- 4, 75 m: basal 4 m matrix supported breccia, the matrix being represented by very coarse sands. Grain size ranges from small pebbles to small cobbles. Reverse grading is probably due to buoyant force acting during the down-stream flow. 11 cm of coarse sands; this interval is laterally discontinuous, being eroded by the overlying breccia bed (64 cm-thick). The basal and upper breccias are amalgamated (Fig. 48);
- 4,00 m; 5,00 m; 0,80 cm; 5,00 m: breccias resembling those previously described, locally with a coarse sandy matrix. Clasts are up to 1,80 m across;

Figure 48: Two amalgamated breccias, and a deeply eroded sandy interbed.

- 1,00 m: two amalgamated beds of very coarse pebbly sandstones, plastically deformed chips of reddish marls, abundant angular clasts of basement rock (Fig. 49);

- 8,00m: clast supported breccia, clast size ranges from few cm's to few dm's; a sandy matrix is present. Clasts pertain essentially to the metamorphic basement;
- 30 m: covered;
- 2,40 m: tripartite bed represented by two breccia intervals (basal 2 m and upper 30 cm) that are partly amalgamated. A thin (10 cm) and laterally eroded interval of coarse sand is present between the two.
- 1,00 m: very coarse pebbly sandstones with scattered clasts (~3 cm).
- 2,00 m; 1,60 m: breccias with a scarce coarse-sand matrix. Both beds exhibit a coarsening upward trend related to decrease of the sandy matrix; Sparse oversized clasts are up to 40 cm across.
- 2,00 m; 3,60 m: structureless breccias with a coarse sandy matrix. Clasts are up to 50 cm across;
- 21 m: poorly exposed massive clast-supported conglomerates;

Figure 49: Sandy interval with a plastically deformed clast; the blue arrow indicates a coarsening upward trend related to freezing of the flow.

The measured section indicates the basal part of the lithofacies 2 is essentially represented by F3 facies (Type 1b beds), testifying the predominance of gravelly dense flows. Conglomerates, especially at their base, are clast-supported, the matrix being virtually missing. The emplacement mechanism can be ascribed essentially to rock-fall and grain flow. Bed Type 5-1b occur sporadically. They are characterised by amalgamated breccia beds that deeply erode arenaceous interbeds. This testifies that the absence of arenaceous intervals in this section can be either due to the predominance of megaclastic sedimentation or to scouring of finer-grained levels.

Section 3

This section is located downstream with respect to the SP-177 one, and stratigraphically below the Trionto section. It is located in the eastern side of the Trionto River. On the opposite side of the river (western side) a steep cliff made of conglomerates dips toward the measured section (Fig. 50).

Figure 50: a) cliff made of breccias of the second lithofacies facing the measured section; b) view of the measured section.

The top of the cliff represents the stratigraphically lower part of the measured section being buried under the river plain. The measurement of a section along the cliff has been impossible due to obvious logistic difficulties, but some observations can be made: this interval is very thickly bedded (> 1 m) and is represented by clast supported, polymictic, poorly sorted conglomerates with pebble- to boulder- size clasts. At the top of the cliff the matrix drastically increases being represented by coarse sands. Clasts are compositionally similar to those found at the base but are loosely scattered in the sandy matrix.

- 24 cm: matrix-rich conglomerates, characterised by a coarse sandy matrix with chaotically scattered rounded clasts pertaining to the basement. Average clast size is ~5 cm, but clasts up to 30 cm across can be sporadically observed (Fig. 51);
- 90 cm: massive coarse sands/granules with millimetric clasts pertaining to the basement, sporadically clasts up to 3-5 cm can be observed;
- 1,08 and 1,40 m: matrix-poor conglomerates with clasts ranging in size from 5 to 8 cm, some clasts can be up to 40 cm. The matrix is a coarse sand. Bases are erosional;
- 32 cm: clast- supported conglomerates with densely packed and chaotically arranged clasts of phyllites only. Clasts are typically 5 cm across. Matrix is essentially missing, while centimetric deformed chips of reddish marls are locally observed;

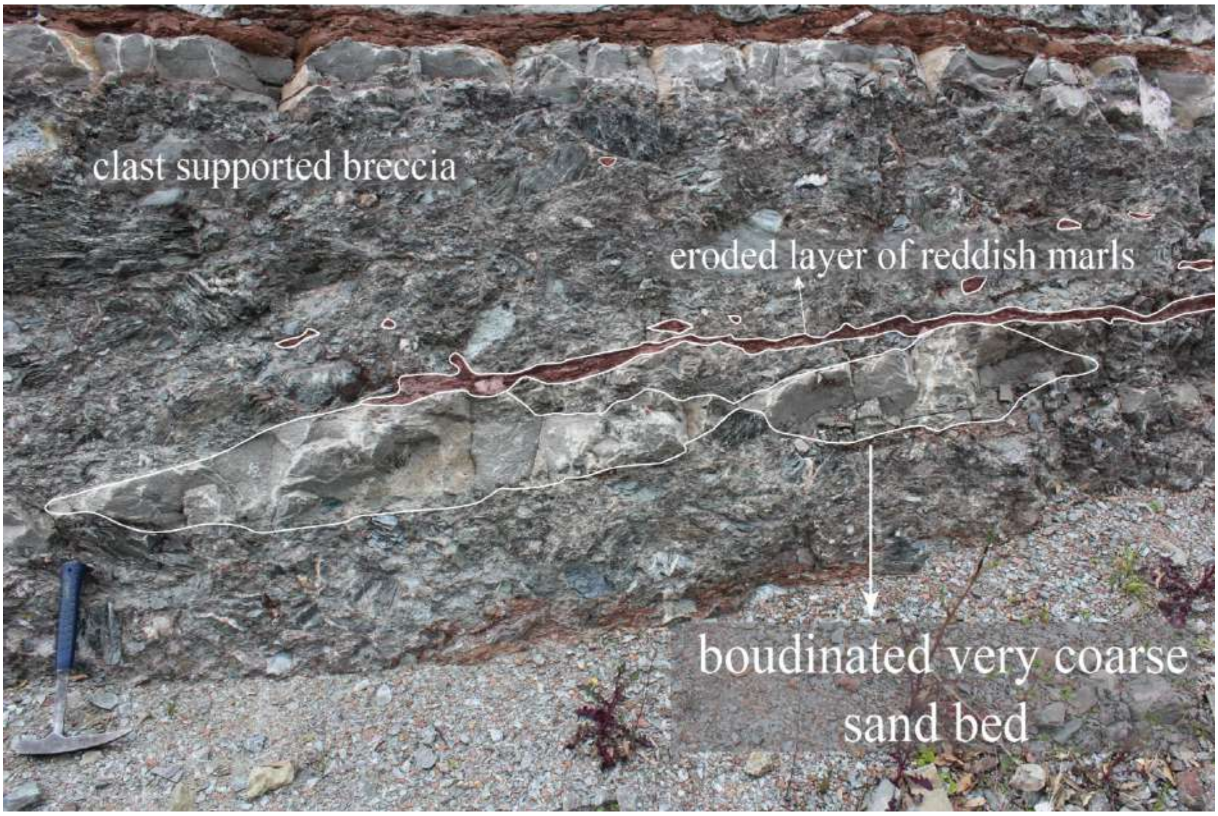

Figure 51: general appearence of the studied section. Thickly bedded clast-supported breccias are alternated with boudinated hybrid arenites and deeply eroded marly interlayers.

- 7 cm: reddish marls that laterally thin-out due to basal erosion by an overlying clastic level;
- 30 cm: clast- supported conglomerates with densely packed and chaotically arranged clasts of phyllites only. Clasts are typically 5 cm across. Matrix is essentially missing, while centimetric deformed chips of reddish marls are locally observed;
- 2,5 cm: reddish marls that laterally thin-out due to basal erosion by the overlying clastic level;
- 45 cm: clast- supported conglomerates with densely packed and chaotically arranged clasts of phyllites only. Clasts are typically 5 cm across, but boulder sized clasts up to 45 cm are also seen. Matrix is essentially missing, while centimetric deformed chips of reddish marls are locally observed;
- 2 cm: reddish marls that laterally thin-out due to basal erosion by the overlying clastic level;
- 6 cm: very coarse sands that are boudinated and laterally eroded;
- 6 cm: reddish marls that laterally thin-out due to basal erosion by the overlying clastic level;
- 18 cm: clast-supported conglomerate showing a coarsening upward trend;

- 28 cm: regular alternation of thinly bedded coarse structureless sands and thinly bedded reddish marls;
- 25 cm: clast-supported conglomerates, clasts are 2-3 cm across, made of basement rock; sparse oversized clasts are up to 50 cm across. Clay chips are plastically deformed reddish marls;
- 2 cm: reddish marls that laterally thin-out due to basal erosion by the overlying clastic level;
- 7 cm: clast-supported conglomerates with an erosive base;
- 23 cm: regular alternation of centimetric, structureless coarse sands and reddish marls; Larger benthic forams are present within the sandy beds.
- 20 cm: coarse structureless sands;
- 5 cm: matrix-poor conglomerates, clasts are about 3 cm. The base is erosive;
- 6 cm: alternation of centimetric beds of coarse sands and reddish marls;
- 15 cm: matrix-poor conglomerates. Clasts are 2-3 cm across and are represented by phyllites;
- 14 cm: alternating thinly bedded reddish marls and thinly laminated sand;
- 15 cm: alternating very thinly bedded reddish marls and coarse sands;
- 35 cm: tripartite bed characterised by a conglomeratic base (9 cm) with angular pebble-sized clasts pertaining to the basement. This interval has an erosive base characterised by injections, and is overlain by a 13 cm thick structureless interval of coarse sands. This interval is eroded at its top by a 13 cm thick conglomerate with densely packed angular pebble-sized clasts of phyllite;
- 20 cm: reddish marls with discontinous silty levels;
- 14 cm: coarse sands with loosely scattered angular pebbles pertaining to the basement;
- 16 cm: alternating very thinly bedded marls and coarse sands;
- 18 cm: bimodal bed characterised by a conglomeratic base (11 cm) and an arenaceous top (7 cm). Conglomerates are moderately sorted, the avarage grain size corresponds to coarse pebbles, with sparse large phyllite cobbles. Sands are structureless.
- 24 cm: coarse structureless sands characterised at their base by scattered angular, coarse phyllite pebbles;
- 3 cm: reddish marls laterally eroded;
- 5 cm: matrix poor conglomerates;
- 9 cm: reddish marls laterally eroded;

- 20 cm: matrix-rich conglomerates: the matrix is a coarse sand, and clasts are coarse angular phyllite cobbles;
- 45 cm: regular alternations of moderately bedded coarse sands and moderately bedded red marls;
- 80: clast supported conglomerates;
- 10 cm: reddish marls laterally eroded;
- 1 m: This interval consists of two amalgamated conglomeratic beds that are moderately sorted, their avarage grain size corresponding to coarse pebbles, with sparse large phyllite cobbles. Between them an arenaceous bed and a marly bed are deeply eroded;
- 70 cm: this interval is characterised by a conglomeratic top that is poorly sorted, with sparse boulder-sized phyllite clasts. The base of this bed deeply erodes the underlying alternation of marls and coarse sands. Marls injects the overlying conglomerates with flame structures, and thinly bedded hybrid arenites are deformed (Fig. 52a-b);
- 144 cm: alternation of moderately bedded very coarse/coarse sands, moderately bedded clast-supported conglomerates and thinly bedded reddish marls. As usual marls are deeply eroded;
- 208 cm: very thickly bedded (between 60 and 110 cm) clast-supported conglomerates characterised by cobble-sized phillite clasts with scattered boulder-sized elements. Reddish marls are present as mud-chips;
- 130 cm: cover;
- 52 cm: clast supported conglomerates characterised by cobble-sized phillite clasts with scattered boulder-sized elements. Reddish marls are present as mud-chips;
- 64 cm: regular alternations of medium bedded very coarse massive sands and thinly bedded laterally eroded reddish marls;
- 8 cm: clast supported conglomerates bearing mud chips, clasts are represented by coarse pebbles;
- 139 cm: regular alternations of moderately bedded very coarse sands and thinly bedded grayisch marls. Sporadically very coarse pebbly sands are present. The top of the sandy bed is often characterised by parallel lamination.

In this section the passage between the second and third lithofacies can be observed. This passage is similar to that observed in the Section 1, as the first conglomeratic interval (here represented in the cliff in the western side of the river) is followed by an interval characterised by very coarse pebbly sands. They can be interpreted as Type 1a and Type 2a beds, representative of a

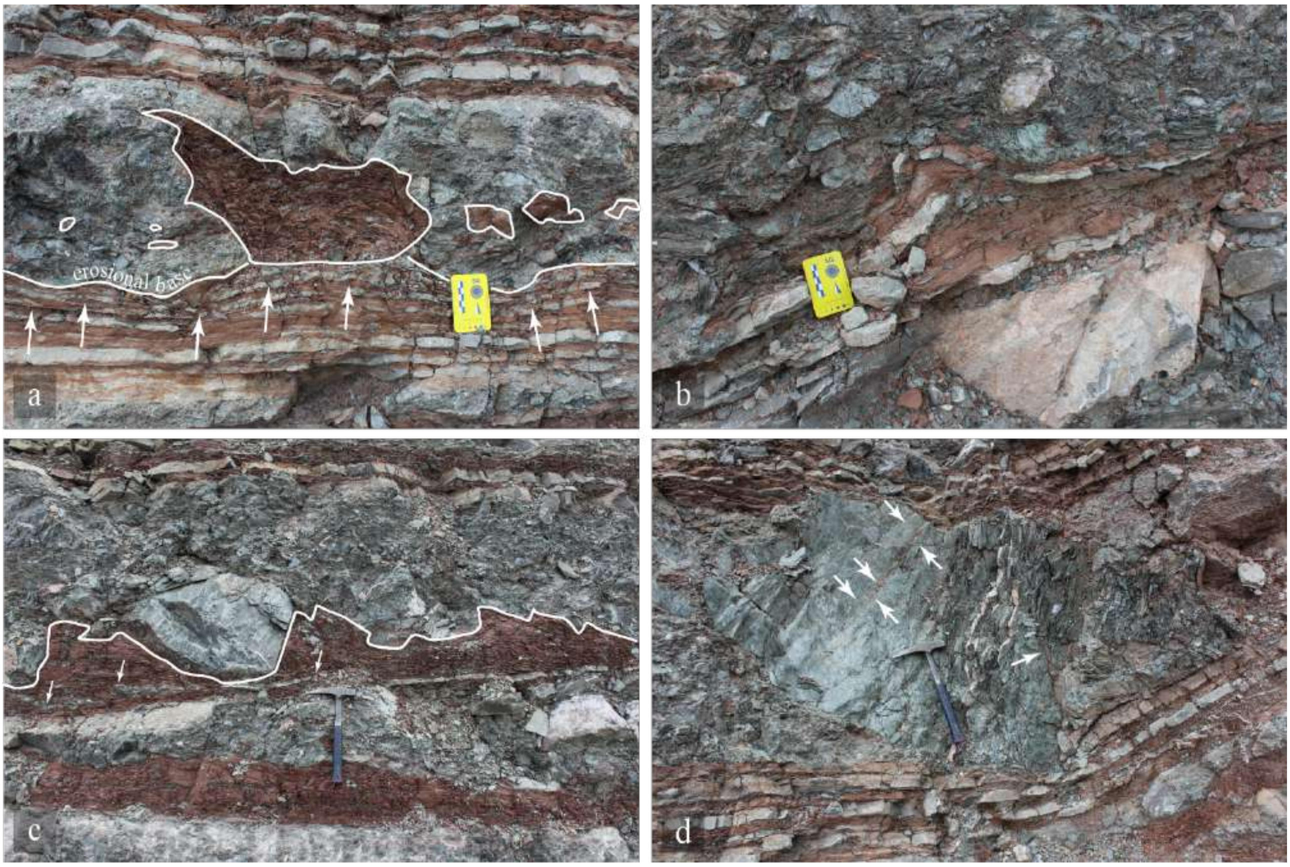

Figure 52: a) plastically deformed clasts within a clastic bed. Note how the substrate is deformed and locally boudinated o boudinaged? by the breccias (white arrows); b) asymmetric deformation within the substrate related to the emplacement of a clastic interval; c) flame structures related to overpressure, white arrows indicate sheared and deformed gray arenaceous layers; d) huge basement-derived blocks deforming the underlying substrate. White arrows highlight the presence of fractures filled with the red marly matrix, indicating fluid overpressure and hydrofracturing.

gravelly and sandy dense flow, respectively. A second conglomeratic interval is represented in the central part of this section by very peculiar conglomerates entirely made of phyllite clasts. Here the most common features are amalgamation surfaces and impact-related structures. The amalgamation of coarse clastic beds is spectacularly represented by the “wavy” appearance and local thin-out of reddish marl or sandstone interbeds, due to the erosive power of the flows overlying them. Reddish marl clasts are often incorporated within the flows and plastically deformed. The very thinly bedded, very coarse sands are typically structureless and can evolve upwards to parallel and cross lamination.

This interval is represented by Bed Type 5 – 1a, indicating strong interaction with the submarine topography, deceleration of the flows and erosion of the seafloor, and by Bed Types 2a, 3b and 3c, indicating the presence of sandy dense flow, bypassing the gravelly dense flows, that are decelerating (more or less rapidly) and are bypassed by turbulent flows.

Based on the peculiar composition of this interval, we can tentatively correlate it with the second clastic pulse of the Section 1. The thickness of these two Key beds is completely different as this section represents the *locus* of accumulation of clastics, being located at the base of the escarpment carved in the phyllites, and downcurrent with respect to Section 1. Conversely Section 1 represents a proximal area, along the escarpment, where only the trailing edges of flows are partially preserved.

The presence of flame structures and fractures filled with matrix testifies fluid overpressure conditions of the matrix to the point of causing hydrofracturing (Fig. 52c-d)

The final part of the section represents the third lithofacies, where major catastrophic clastic pulses are essentially missing.

Section 4

This section has been measured in a downstream position with respect to the Section 1.

- 7.20 m: thickly to very thickly bedded conglomerates. Clasts are essentially small pebbles/small cobbles, densely packed in a sandy matrix. Oversized clasts (up to 25 cm across) pertain to the basement or to the Mesozoic succession;

Figure 53: a) typical aspect of massive, structureless thickly to very thickly bedded very coarse sands; b) medium bedded basal F5 interval displaying at its top crude parallel lamination (F7), partially eroded marly interval, overlain by very coarse sand displaying a basal part characterised by crude and thinning upward laminae (F7) and an upper part characterised by trough cross stratification; (F6); c) impact related structures in a pebbly sandstone; d) example of amalgamation surface in a Type 2a bed (pebbly sandstones); note how the marly rip-up clasts display soft sediment deformation (pseudo-sigma structures)indicating a top to the South (left of photograph) sense of shear.

- 83 cm: alternations of thinly bedded reddish marls and thinly to medium bedded coarse sands, in massive and structureless beds with erosive bases. Marls display lateral thickness variations, being eroded by the calcarenites, resulting in amalgamated beds. In the arenaceous intervals LBF have been detected: such microfossils characterise from now onward all the arenitic beds.

- 80 cm: reddish marls;

- 21 cm and 9 cm: coarse structureless and massive sands with erosive bases;

- 1.23 m: alternations of thinly bedded reddish marls and thinly to medium bedded structureless and massive coarse sands;

- 36 cm: microconglomerates, with small pebbles scattered in an abundant sandy matrix; Basal sand injections and delamination structures are present;

- 1.05 m: chaotic interval, made of plastically deformed coarse sands in a reddish marly matrix;

- 1.30 m: alternations of medium bedded marls and coarse structureless sands;

- 1.93 m: thickly bedded coarse structureless sands/very coarse pebbly sands. Marl chips mark amalgamation surfaces;

- 47 cm: alternations of medium bedded coarse structureless sand and greenish marls;

- 70 cm: conglomerates/ pebbly coarse sands, with phyllite clasts up to 10 cm across, floating in a marly/silty reddish matrix;

- 1.20 m: medium to thickly bedded coarse sands, in structureless beds; discontinuous marly interlayers indicate the calcarenites are commonly amalgamated (Fig. 53a);

- 32 cm: conglomerates with angular phyllite clasts up to 20 cm across, scattered within a sandy matrix;

- 30 cm: medium bedded coarse sand, structureless;

- 31 cm: greyish marls;

- 3.0 m: alternations of thinly bedded greyish marls and medium to thickly bedded coarse sands/pebbly coarse sands, arranged in structureless beds with erosive bases. The marls display lateral thickness variations, being eroded by calcarenitic beds that are often amalgamated; along amalgamation surfaces, marl rip-up clasts show plastic deformation features like pseudo-sigma structures and boudinage. Some calcarenites show crude parallel lamination at their top or wavy laminae (Fig. 53b-d);

- 22 and 21 cm: two amalgamated conglomeratic beds; the amalgamation surface is marked by plastically deformed marl chips; conglomerates are made of small angular cobbles of phyllite in a sandy matrix;

- 46 cm: alternations of thinly bedded greyish marls and medium to thickly bedded coarse sands, arranged in structureless beds with erosive bases; marls are often eroded along amalgamation surfaces.

The section is entirely made of Lithofacies 3.

This section exhibits a general fining and thinning upward trend. The basal interval (7 m) is essentially characterised by Type 1a beds, that are the product of gravelly dense flows.

Upwards the section is monotonous and records essentially the deceleration of the leading heads of sandy dense flows, being the Type 4-2a beds predominant. The most common features are amalgamation surfaces, highlighted by the presence of plastically deformed rip-up clasts indicating substrate delamination. Locally sand injections are indicative of excess pore-pressure in the parent flow. The analysis of pseudo-sigma structures indicates that the flows were directed toward the South. In the last 4 metres a further fining trend is detected through the dominance of Type 2b and Type 4-2b beds. Intervals characterised by even parallel lamination (F7) and trough cross lamination (F6) indicate the reworking of F5 intervals by traction carpets of a bypassing turbulent flow. All the above indicates a zone of deceleration of sandy dense flows with associated decoupling of upper turbulent flows capable of producing tractive structures (F7 and F6). F7 beds predominate, indicating a moderate decrease in velocity.

This section is located in a downcurrent position with respect to the Section 1, as the third lithofacies is represented by coarser grain size (from coarse/very coarse sands up to granules) being the product of the leading edges of gravelly/sandy danse flows. In this section the Lithofacies 2 is not recorded being located in a lower stratigraphic position.

Section 5

Along the Laurenzana River a beautifully exposed section of the Paludi Fm is found. The river flows toward the north-west at an altitude of 236 m a.s.l. and the valley cuts the formation transversally with respect to its dip direction, so it is possible to observe both the vertical organization of the facies and their lateral variations. Contrary to the other localities, where the conglomeratic lithofacies is monotonous, here different sub-lithofacies have been mapped, suggesting the action of an array of processes.

The section ends after 50 metres as the river takes a direction parallel the strike of beds, and observations are made impossible by the very steep wooded slopes.

The contact with the underlying lithofacies is unclear, being covered by recent scree, but bed attitudes suggest a high angle contact. This is the only section where conglomerates rest on the marly lithofacies, instead of the basement.

- 2 m: massive grey silty marls, rich in sand-sized grains derived from the metamorphic basement;
- 2.3 m: massive grey silty marls with rare blocks, decimetres to 1 meter across;
- 3.60 m: the clastic component increases, the abovementioned grey matrix still prevails, but clasts are very abundant and range in size from some millimetres to 2 metres; clasts are made of angular phyllites, subrounded granites, sub-rounded quartz, sub-rounded Lower Caloveto limestone, angular radiolarian cherts and, remarkably, a metre-size clast of bedded calcarenites rich in macro-foraminifera; calcarenite clasts are plastically deformed and boudinaged, and the marly matrix is deformed below and around them due to differential compaction. In this interval reddish marly clasts, plastically deformed, form "pseudo-sigma" structures (Fig. 54a-b). These reddish marls pertain to the reddish marls of the second lithofacies of the Paludi Fm;
- 5,60 m: the grey silty matrix decreases gradually toward the top, clasts are still subordinate and are made of the same lithologies as in the previous interval; phyllite clasts can be up to 2 meters.

The deposits change abruptly upward, through a complex erosive surface. This surface is the site of intense deformation of a discontinuous marly interval, representing a beautiful field example of basal shear surface, where "no-slip flow" interaction can be described (*sensu* Sobiesiak *et alii*, 2018) (Fig. 54c-d).

- 16 m: the fine silty matrix disappears, and the interval is represented by very thickly bedded (two beds of 6 and 10 metres respectively), poorly sorted, clast-supported conglomerates (Fig. 54c); the matrix is a very coarse sand; the deposit is heterometric and poorly sorted, with clasts ranging in size from few millimetres to 40 cm; clasts can be angular (phyllites, radiolarian cherts, calcarenites, rare encrinites, Lower Caloveto Fm.) or sub rounded (Lower Caloveto limestone, granites, quartz); reddish marls of the second lithofacies are plastically deformed forming "pseudo-sigma" structures; in some clasts, fitted contacts or injection of the sandy matrix can be detected. This clast-supported interval is clearly wedge-shaped, reaching its maximum thickness toward the southwest and pinching out toward the northeast. The conglomerates can be followed along the river parallel to their strike direction for 1.3 km.

At this point the river plain cuts the outcrop and the measurement has been divided into two sub sections at opposite sides of the valley: the first section continues along the north-eastern side, while the second continues in the south-western side.

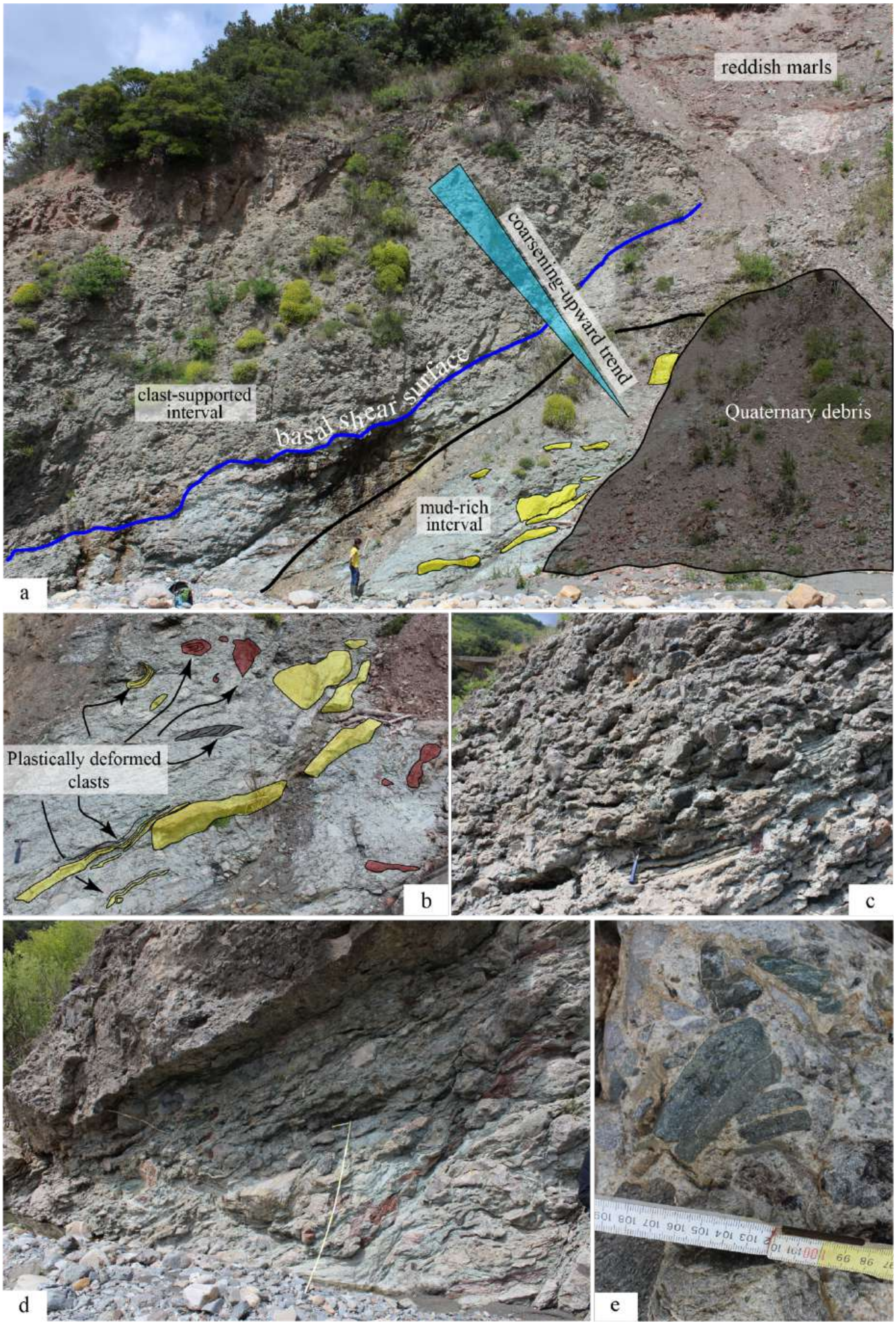

Figure 54: a) view of the basal interval of the measured section; b) plastically deformed clasts of hybrid arenites (yellow) and reddish marls (red); c) appearance of the clast supported interval; d) close up on the basal shear surface, affected by strong deformation (pseudo-sigma and pseudo-sc structures); e) hydrofracturing.

Considering the local width of 120 m of the river bed, and a dip of about 20° of the conglomerates, the missing thickness of the section would be of about 41 metres. It must be noted that the exposed intervals are lens-shaped, therefore the minimum and maximum thickness for each will be indicated.

In the north-eastern side, the succession continues upward with:

- 11 m: conglomerates, partially covered; At 6 meters above the base of this interval, a megaclast made of a 1.80 m-thick stack of white cherty limestone beds (Maiolica Fm) can be followed laterally for about 50 m. Above and below this huge clast, conglomerates are comparable to the 16 m interval described above. In the last two meters of this interval, the reddish marl matrix drastically increases as the clasts are made of sub-rounded elements of the Lower Caloveto Fm.

In the south-western side of the river the section continues with:

- 0-40 m: lensoid matrix-rich breccias. The matrix is a dark clay displaying intense deformation in the form of shear planes. Clasts are moderately sorted and made of cm- to dm-sized elements of radiolarian chert, chaotically scattered within the matrix, while flat elements are not aligned. Up-section this lithofacies displays a chaotic lens-shaped megaclast made of disrupted strata of a greyish burrowed limestone, interpreted here as a block of the upper part of the Maiolica Fm (Fig. 55).

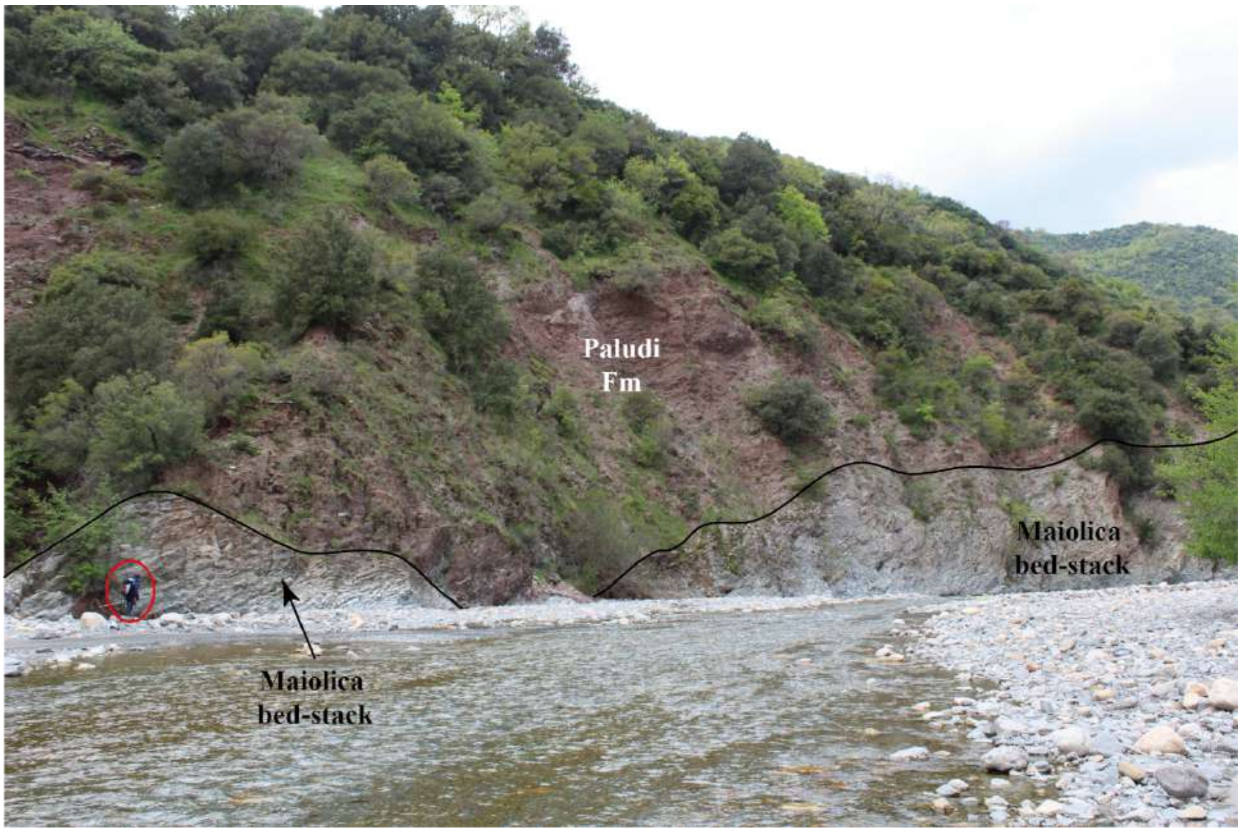

Figure 55: Huge Maiolica megablocks embedded in the reddish marls. Geologist for scale in the red circle.

This block has a maximum stratigraphic thickness of 6 metres. Two adjacent megablocks are also present, for a total lateral length of about 500 meters, draped by reddish marls. The beds composing the megablocks are recognisable, but are locally broken, boudinaged and slumped, being the original bedding completely hidden. In the basal part of the blocks complete stratal disruption, with the development of a block-in-matrix fabric, is observable. Clayey interbeds are present, so wherever stratal disruption is severe, the mud intrudes the broken elements, deforming them plastically.

The two Maiolica bed-stacks are covered by the marls of lithofacies 2. Locally the reddish marls and the Maiolica-like strata are concordant, so at first sight they could be mistaken for one stratigraphic succession (Fig. 56).

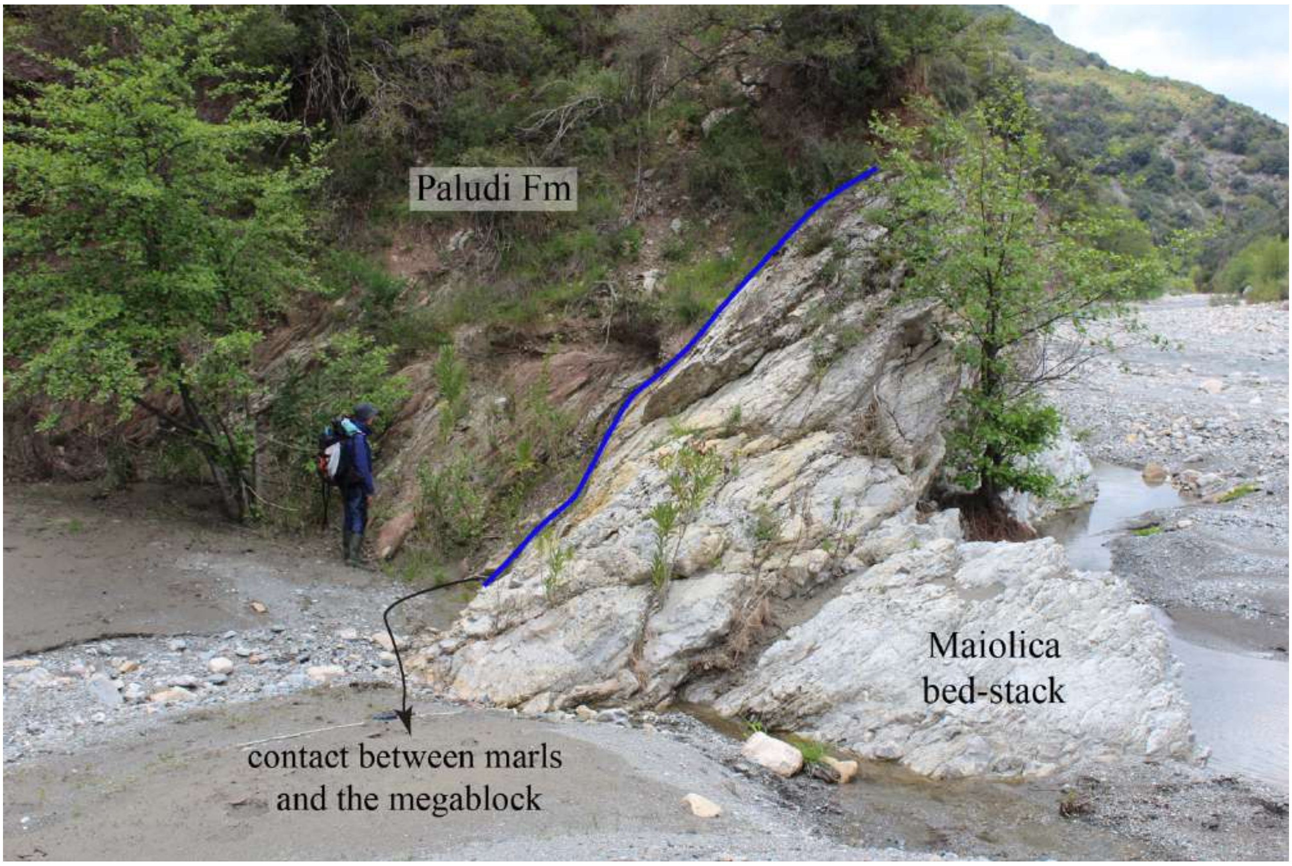

Figure 56: Concordant contact between marls of the Paludi Fm and the Maiolica bed-stack.

- 13.9 m: reddish marls, with scattered clasts in the basal part of the interval. Up-section thinly bedded coarse sands are interbedded within the marls.

The stratigraphic section cannot be followed along the western flank of the Valley because a thrust plane cuts the marls, juxtaposing Hercynian phyllites on the lithofacies 2.

In the opposite side of the valley, it can be seen that sedimentation continues upwards with the turbidites of the third lithofacies. A detailed measurement of that unit has proven impossible.

The Laurenzana outcrop displays a peculiar vertical organization with respect to the other measured sections pertaining to Lithofacies 2. To summarize, we can divide the measured section into two intervals: the first is characterised by a general coarsening upward trend (gradual increase in clastics and decrease in marly matrix) and by a basal matrix-rich conglomerate (that can be further divided into two distinct intervals that we will name for simplicity U1 and U2) followed by a matrix-poor breccia (that we will name U3) (Fig. 57).

An important distinction between the two basal marl-rich intervals (U1 and U2) concerns the typology of the clasts embedded in the matrix and their sedimentological features. In the U1 interval clasts pertain only to the basement/Mesozoic cover, are loosely scattered within the matrix, and are few. In the U2 interval, beside basement/Mesozoic cover- derived elements, clasts pertaining to the lithofacies 1 and 3 are common, displaying soft sediment deformation. Deformation can be more or less marked within hybrid arenites blocks, that can in fact preserve their stratal geometry, or can be completely folded. Below and above the deformed blocks, the encasing beds are deformed due to loading (underlying beds) and differential compaction (lateral and overlying beds). Clasts of red marl are strongly deformed.

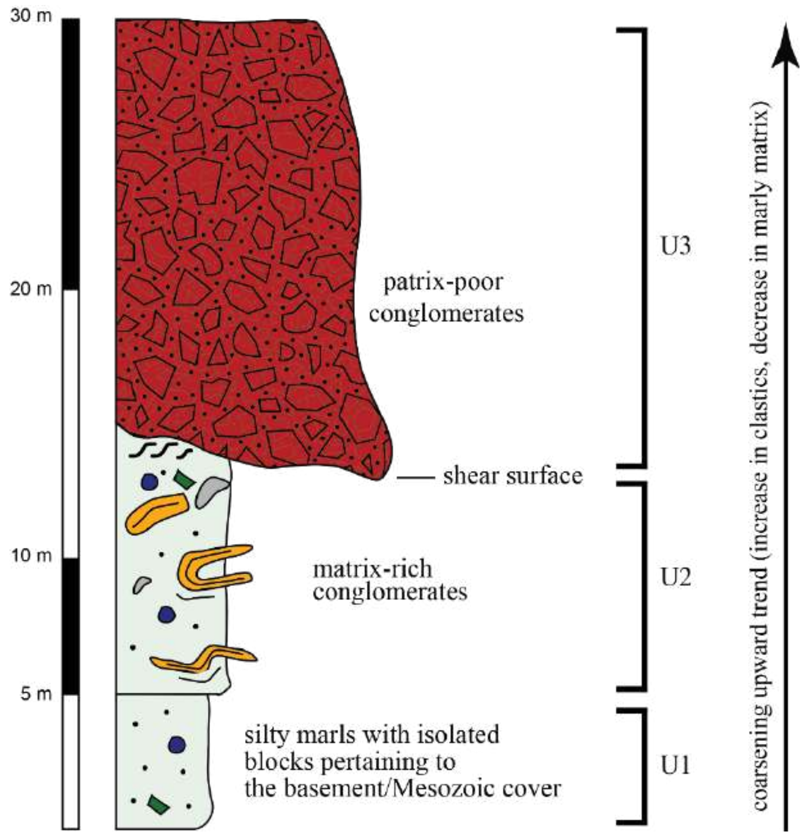

Figure 57: Summary of the main characteristics of the basal clastic interval.

The two intervals are consequently different from a genetic point of view. The U1 interval can be interpreted as the product of hemipelagic sedimentation punctuated by episodic rockfalls sourced from an uplifted sector of the basin.

The U2 interval is conversely the product of Mass Transport Processes, as testified by the presence of unconsolidated deposits that are essentially penecontemporaneous with the megaclastics (hybrid arenites and reddish marls), scraped off from the substrate and plastically deformed, indicating that these clasts must have been subjected to shear stress related to movement within a flow.

The U3 interval is a breccia, characterised by a scarce sandy matrix with densely packed elements ranging in size from small pebbles to boulders, and can be ascribed to a F3 facies. As a whole, the U2 and U3 intervals are referred to Type 1c beds. The two are the product of an hyper-concentrated flow and of a dense gravelly flow, respectively. We cannot exclude that the two are to be considered co-genetic and the deposit is the product of a bipartite flow. The upper F3 interval is characterised by a beautifully preserved basal shear surface, characterised by plastically deformed and sheared greyish marls. This surface indicates a strong interaction with the underlying deposits and consequently a deceleration of the gravelly dense flow. The presence of hydrofracturing in the U3 interval indicates overpressure conditions.

The second interval, that is unfortunately partially covered by the river plain, is characterised by a basal matrix-rich breccia, overlain by huge Mesozoic cover-derived megablocks. The matrix of the basal interval is a dark clay, strongly deformed by ductile shear planes. The overriding megablocks display a variety of soft sediment deformation and in general stratal disruption. At the very base of the Maiolica bed-stacks, the stratal disruption is complete, and the development of a block-in-matrix fabric can be observed. Above this interval a folded interval can be observed. Upsection, bedding is characterised by asymmetric boudinage. Also this interval can be considered as a Type 1b bed devoid of the upper F3 division and characterised by the basal matrix-rich unit (F2) produced by an hyperconcentrated flow and overlain by huge megablocks.

Mass transport deposits are commonly characterised by a vertical organization similar to the one displayed in this outcrop: basal matrix-rich interval overlain by a clast- supported interval (Pini *et alii*, 2012; Ogata *et alii*, 2014). The basal hyperconcentrated flow has a strong erosional power in particular at its front, producing a “bulldozer effect” on the substrate. This results in the incorporation of huge parts of the seafloor within the flow, with a higher concentration in the leading edges (Tinterri & Piazza, 2019). Furthermore, due to grain-to-grain interaction and to buoyant forces, clasts concentrate in the upper portion of the flow. This leads to the development of the so called blocky-flows (Ogata *et alii*, 2014; Mutti *et alii*, 2006) at the front of F2 intervals.

We can therefore explain the decrease in marly matrix and the increase in clasts in the U1 interval in the light of the abovementioned processes: "bulldozer effect" at the front of the flow, plus action of buoyant forces. In any case, even the uppermost part of the U1 interval cannot be considered as a true blocky flow deposit, like those described by Ogata *et alii*, 2014 in the Paleogene Friuli Basin. A similar consideration can be applied to the upper interval of the section: a basal matrix-rich interval (Fig. 57a) is followed by a block-dominated interval, and only the latter can be considered a genuine blocky flow deposit.

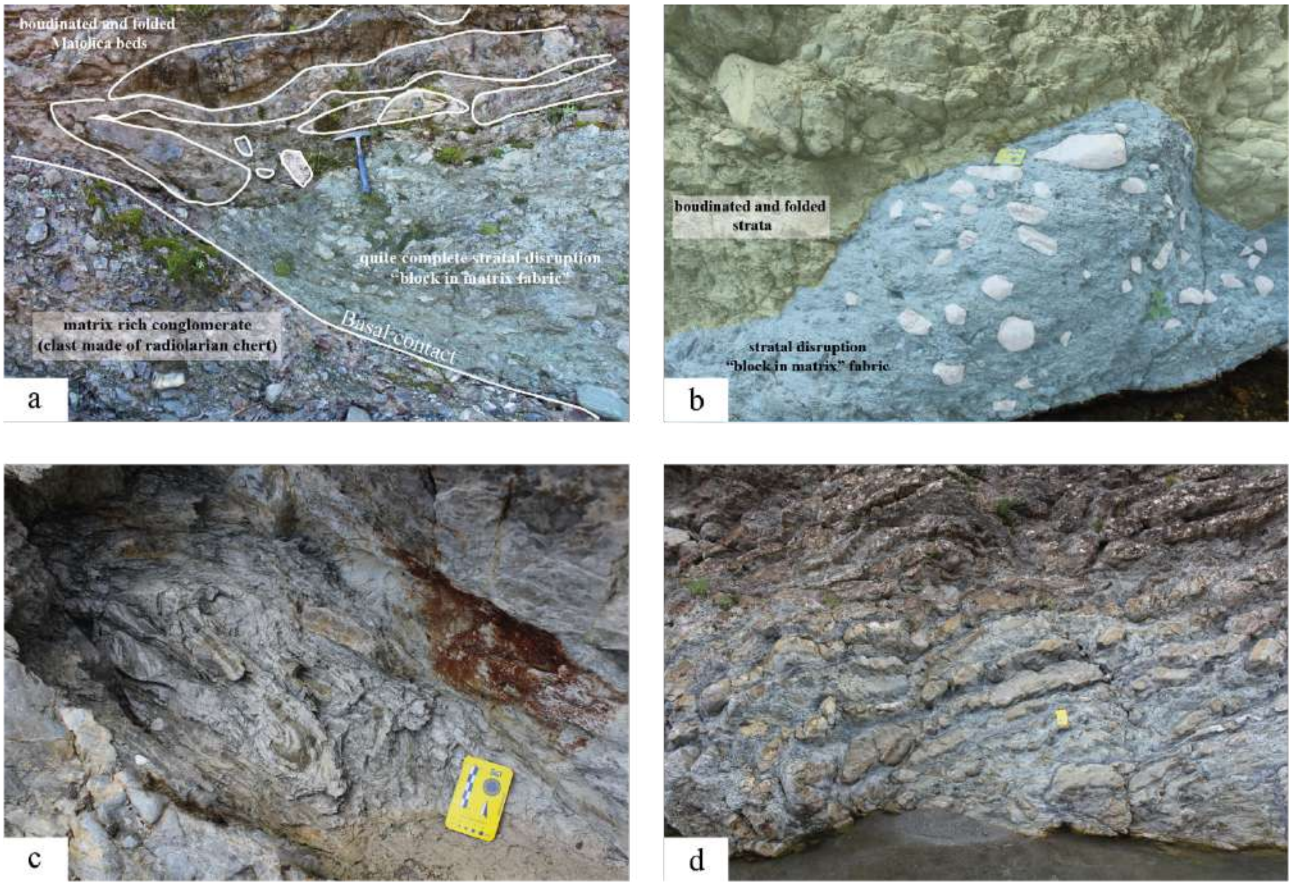

Figure 58: a) basal contact between the Maiolica megablock and the basal matrix-rich interval; b) basal part of the megablock displaying complete stratal disruption; c) folded strata; d) boudinaged interval.

Section 6

In Vallone Sant'Elia the transition between the conglomeratic lithofacies and the hybrid arenites is exposed. Here a thrust fault juxtaposes the Middle Jurassic Calcari Grigi con selce on the Paludi Fm, while a minor basal splay of the main thrust further displaces the conglomerates. The presence of parasitic folds in the backlimb of the footwall syncline makes it difficult to measure a thick section within the Paludi Fm.

Bedding attitude is toward the south and the contact with the basement is well exposed, highlighting pinch out geometries of the conglomerates, whose thickness decreases toward the North and toward higher elevations. Minor angular unconformities are also detectable between the arenites and the conglomerates (Fig. 59).

- 25 cm: poorly sorted breccias with clasts pertaining to the igneous and metamorphic basement. Clasts are essentially angular (apart from some rounded Qz and granite clasts) and the flat ones are randomly oriented. Clasts are up to 3 cm and are scattered in a coarse sandy matrix;
- 35 cm: poorly sorted breccias with clasts up to 6 cm. The matrix is a coarse sand, with sparse clasts of igneous and metamorphic basement rock. Phyllite clasts are always angular, while Qz and granite clasts are commonly subrounded;
- 54 cm: Very poorly sorted breccias with phyllite clasts up to 30 cm across. Clasts are subangular, granite and Qz clasts are typically subrounded. Matrix is abundant and is represented by very coarse sands;
- 34 cm: bimodal bed represented by (i) basal interval (30 cm) of very coarse sands with rare flat clasts up to 2 cm pertaining to the metamorphic basement; (ii) upper interval (4 cm) of grayish marls;
- 12 cm: structureless fine gravel with an erosive base;
- 7 cm: grayish marls;
- 15 cm: very coarse structureless sands with erosive base, their top (2 cm) is represented by laminated coarse sands;
- 7 cm: grayish marls;
- 43 cm: alternations of thinly bedded very coarse sands with erosive bases and thinly bedded grayish marls (locally Larger Benthic Forams are glauconitised);
- 14 cm: very coarse structureless sands, with laminated top (6 cm);
- 16 cm: fine gravel;
- 1 cm: greyish marls, laterally discontinuous;
- 30 cm: structureless fine gravel (LBF glauconitised);
- 28 cm: the basal part is structureless very coarse sand, the upper part (2 cm) shows crude parallel lamination;
- 8 cm: very coarse structureless sands with erosive base;
- 20 cm: covered;
- 17 cm: the basal part (14 cm) is a very coarse structureless sand, overlain by laminated very coarse sands (3 cm);

- 7 cm: very coarse sands with erosive base;
- 30 cm: very coarse structureless sands (uppermost 7 cm are laminated); this bed is draped by a thin and discontinuous layer of greyish marl;
- 24 cm: alternations of thinly bedded very coarse sands and greyish marls laterally discontinuous;
- 50 cm: coarse gravel, densely packed with Jurassic clasts;
- 10 cm: greyish marls;
- 12 cm: two beds of very coarse structureless sands with erosive bases;
- 34: structureless coarse gravel, with sparse basement cobbles.

In the Vallone Sant'Elia section, the passage between the first and third lithofacies can be observed. The lower 114 cm are essentially represented by matrix-rich breccias (Type 1a beds) that are the product of gravelly dense flows. In this section the clastic interval is characterised by pinch-out geometries indicating that the clastic wedge abutted the basement. The rest of the section is represented by monotonous alternations of very coarse sands/pebbly sandstones and marls referrable to Type 2a beds that are essentially the product of sandy dense flows.

The absence of medium- to fine-grained sands indicates that these grains sizes were transported downcurrent.

Figure 59: Stratigraphic contact between the phyllites, the second and third lithofacies. Note that the conglomerates pinch out toward the north.

Section 7

This is the westernmost measured section. Unfortunately, the section is extremely discontinuous due to modern erosion along the slopes of the river valley. At the same time, this section shows peculiar characteristics with respect to the previously described sections. For this reason, despite the composite nature of the stratigraphic log, I have decided to describe the most important characteristics, even if robust quantitative data are missing.

Three short sections were measured. I will now proceed with a qualitative description of the whole area, but the reader can refer to the Table for the measured intervals.

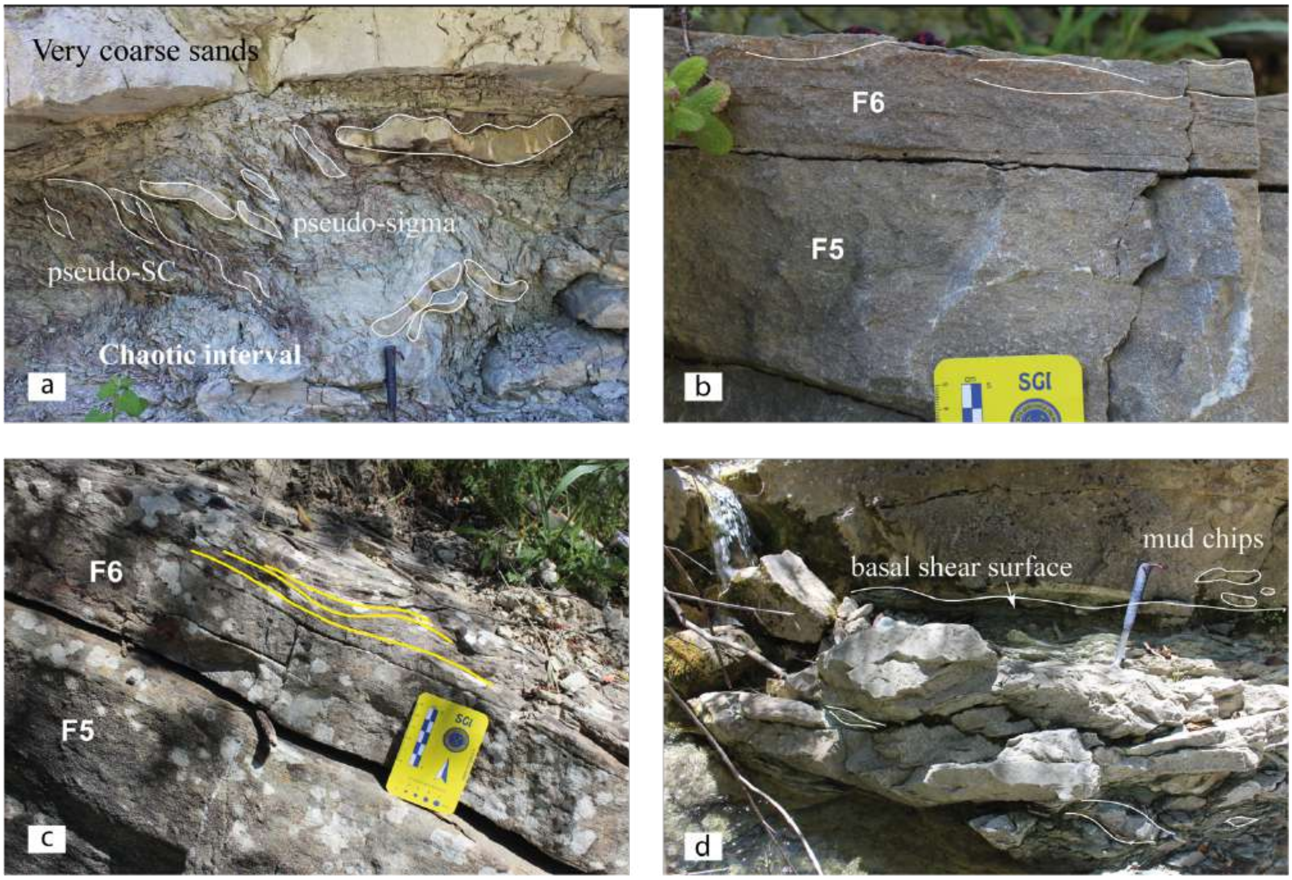

Figure 60: a) chaotic interval at the base of a very coarse sandy bed (F5). The arenaceous interlayer are deformed by pseudo-sigma structures while the marly portion is intensively foliated and deformed by pseudo-SC structures; b) and c) coarse sands characterized by wavy laminae (F6) overlie coarse massive sands (F5), pointing to the by-pass of the upper turbulent flow; d) chaotic interval at the base of a F5 interval characterized by mud-chips.

In this area the basal megaclastic unit (second lithofacies) is missing (or buried) and the hybrid arenites directly onlap the Hercynian phyllites.

In this section the grain-size population changes slightly as very coarse sands/pebbly sandstones and breccias, arranged in very thick/thick beds, are essentially missing. Here coarse sands in moderately/thinly bedded intervals predominate (F5), bearing occasional pebble-size clasts. We are thus dealing with sandy dense flows that pinch-out down current. Such beds are often overlain by laminated intervals (both parallel laminae and wavy laminae) ascribable to the F7 and F6 facies (Fig. 59b-c). This testifies the by-pass of turbulent flows. Locally some medium-grained thinly bedded structureless sands directly overly the F5 facies (F8). This jump in grain size indicates the by-pass of turbulent flows.

A further characteristic of this section is the occurrence, at the base of the coarser F5 beds, of "chaotic" intervals, characterised by sand injections, deformed mud-clasts (pseudo-sigma structures), and contorted and deformed sandstones clasts (Fig. 60a-d). At the contact with the overlying coarse sandy beds, spectacularly preserved shear surfaces within the marls can be observed.

All the above indicates strong interaction with the substrate at the head of the flows due to their decrease in velocity. The substrate was thus sheared and deformed and at the same time the upper part of the bipartite flow by-passed the dense flow generating the F7 and F8 facies.

Section 8

- 26.40 m: reddish marls, highly deformed by scaly fabric (Fig. 61);
- 7 cm: very coarse structureless sands with basement clasts up to 4 cm across; cm-sized plastically deformed clasts of reddish marls are also present;
- 9 m: reddish marls;

Figura 61: General appearence of the first lithofacies.

- 60 cm: alternating very thinly bedded reddish marls and thinly bedded very coarse structureless sands with erosive bases;
- 20 cm: two amalgamated beds (8 and 12 cm respectively) of very coarse structureless sands, the amalgamation surface is identifiable through the presence of plastically deformed reddish marls;
- 330 cm: reddish marls;

- 106 cm: alternation of medium/thickly bedded coarse structureless sands with erosive bases and thinly bedded reddish marls, with rare sand injections (Fig. 62), clay chips are common;

Figure 62: Sand injection.

- 7 and 10 cm: microconglomerates with clasts pertaining essentially to the basement; the two beds are separated by a reddish marl bed (maximum thickness 7 cm) that is locally completely eroded;
- 25 cm: alternation of thinly bedded very coarse/coarse sand and reddish marls;
- 80 cm: reddish marls;
- 11 cm: massive coarse sands;
- 42 cm: reddish marls;
- 12 cm: microconglomerates with clasts up to 2 cm, with scarce matrix;
- 15 cm: coarse structureless sands;
- 1 m: reddish marls;
- 11 cm: massive coarse sands;
- 42 cm: reddish marls;
- 11 cm: massive coarse sands;
- 42 cm: reddish marls;
- 12 cm: microconglomerates with clasts up to 2 cm, with scarce matrix;
- 15 cm: coarse structureless sands with crude parallel lamination (Fig. 63);
- 1 m: reddish marls;
- 3 cm: microconglomerates with clasts up to 2 cm, with scarce matrix;
- 1 m: reddish marls;
- 6 m: covered;
- 1. 47 m: alternation of medium bedded reddish marls and thickly bedded coarse structureless sands;

- 88 cm: three amalgamated beds of coarse sands; amalgamation surfaces can be recognised thanks to local preservation of otherwise deeply eroded thin marl layers. In the intermediate bed pillow structures deform both the sandstones and the marls (Fig. 64);
- 56 cm: two amalgamated beds of very coarse structureless sands, with angular basement clasts (< 5 cm). Plastically deformed chips of reddish marls occur along the amalgamation surface;
- 30 cm: graded coarse sands, with paralle-laminated top;
- 3 cm: reddish marls;
- 55 cm: three beds (10, 22, 23 cm) of very coarse structureless sands, with parallel-laminated top;

Figure 63: Very coarse sands with crude parallel lamination (F7 bed).

- 87 cm: alternations of very thickly bedded coarse structureless sands and very thinly bedded reddish marls;
- 2,67 m: alternations of medium/thickly bedded structureless coarse sands, medium/thickly bedded microconglomerates and medium bedded reddish marls. The topmost part of the interval (within the coarse sands) is characterised by crude lamination;
- 8,26 m: regular alternations of medium/thickly bedded very coarse pebbly sands/microconglomerates, medium bedded very coarse sands and thinly bedded reddish marls, clay chips are common;

Figure 64: a) Amalgamated bed made of very coarse sands, note the deformed chips of reddish marls; b) post-depositional structures, marking fluid escape due to fluid overpressure.

- Gully;
- 1.73: regular alternations of medium/thickly bedded very coarse pebbly sands/microconglomerates, medium bedded very coarse sands and thinly bedded reddish marls;
- 6 m poorly exposed marls;
- 100 cm: reddish marls;
- 4,69 m: regular alternations of thinly bedded marls, noderately/thinly bedded coarse sands and sporadic thickly bedded conglomerates;
- 60 cm: breccias, with abundant plastically deformed clasts of reddish marls. The matrix is represented by very coarse sands; clasts are prevalently basement-derived elements, while clasts of the sedimentary cover are rare;

Figure 65: Conglomerates (F3 facies) with plastically deformed clasts of the reddish marls (arrowd) are overlain by coarse sands with wavy laminae (F6) indicating by-pass of the upper turbulent flow.

- 30 cm: reddish marls;
- 17 cm: coarse/very coarse sands;
- 30 cm: very coarse sands/ pebbly sandstones with abundant plastically deformed clasts of reddish marl up to 4 cm. Sparse centimetric angular clasts of the basement are also observed. They are localized in the upper part of the interval, which is probably due to buoyant force related to contrast in density and to the sudden freezing of the flow;
- 12 cm: coarse sands showing cross-bedding (through cross lamination) capped by 17 cm: reddish/greenish marls (Fig. 65);
- 33 cm: pebbly coarse sands;

- 48 cm: marls;
- 300 cm: clast-supported breccia, with scarce marly matrix. Plastically deformed reddish marls are up to 20 cm thick. Clasts up to 10 cm across represent both the basement (angular to sub angular) and its Mesozoic cover (sub-rounded) (Fig. 66);

Figure 66: Clast-supported breccia.

- 24 cm: breccias with clasts (up to cobble grade) belonging to the basement and its Mesozoic cover, embedded in a coarse sandy matrix;
- 72 cm: structureless very coarse pebbly sands;
- 48 cm: alternations of medium bedded coarse sands and thinly bedded marls;
- 53 cm: breccias with coarse sandy matrix;
- 100 cm: alternation of medium bedded coarse sands and thinly bedded marls;
- 30 cm: breccias with clasts up to 3 cm across in a scarce sandy matrix;

This section is located in another tectonic unit with respect to the others. Here the passage between the lithofacies 1 and 2 can be observed

This section is characterised by alternations of:

- Moderately bedded, disorganized breccias made of pebble-size clasts made of reddish marl and of basement rock (less common Mesozoic clasts) in an abundant sandy/marly matrix, these deposits can be interpreted as F2 facies, and are the product of hyperconcentrated flows;

- Very thickly bedded breccias, with angular clasts up to 5 cm across in a sandy matrix. F3 facies related to gravelly dense flow. It must be remarked that such beds are rare in this section;
- Very coarse pebbly sandstones/very coarse sandstones that can be interpreted as F5 facies and are the product of sandy dense flows. These beds can also be amalgamated, displaying impact structures, mudstone clasts and locally also fluid escape structures. These beds record a strong interaction with the substrate due to the deceleration of the sandy dense flow (Fig. 67);
- Locally the F5 facies can be overlain by F7 and F6 facies, recording the by-pass of the upper turbulent flow;

All the above indicates that the main sedimentation processes were related to the action of dense flows, both gravelly (F2 and F3 facies) and sandy (F5). The action of turbulent flows can only be detected through the presence of sporadic F6 and F7 facies. The background sedimentation is essentially represented by hemipelagites that were largely eroded by the clastics but are still present in a higher percentage with respect to the other sections.

Concerning the dense flow-related deposits, they do not form extensive wedge-shaped deposits (as for example in the Trionto valley), but are rather intercalated within the reddish marls. Furthermore, the clastic deposits are not clast-supported as for the Trionto River, but are matrix-

Figure 67: Very coarse pebbly sandstones, with clast pertaining to the Jurassic succession (arrowed in blu) and clast scraped-off from the substrate. Note that one of these marly elements present a fracture filled by the coarse sandy matrix (white arrow) indicating hydrofracturing. The marly interlayer is eroded.

supported, indicating different emplacement mechanisms (rock-fall vs debris flow) for these two deposits. This suggests that the Torno area was a distal zone with respect to the Trionto area. These deposits are thus the product of the leading edges of the flows, enriched in finer-grained material and thus capable of higher run-out distances. This distal section highlights that, at least in the basal part of the Paludi basin, the system was largely not characterized by turbulent flows.

Another important consideration concerns the abundance of marly deposits (reddish marls). This testifies that the background sedimentation in the basin was essentially hemipelagic. These deposits where deeply eroded in the more proximal areas, as testified by the Section 3, where the clastic deposits erode the marly interbeds that are locally completely cannibalized, and breccia beds are consequently amalgamated. In this distal section, in contrast, the erosional power of the dense flows was lesser, and the marly deposits are generally preserved as a result.

In the light of the above, we propose that during the initial stage of the formation of the megaclastic interval, dense flows were prevailing, with very few exceptions (represented by the F6, F7 and mud-draped scours of the distal section) and the turbidity flows (if present) where sedimented in other sectors of the basin that are today lost. Only in a successive stage, when the parental flows where enriched in finer grain size populations sourced from the shelfal areas, turbulent flows where able to fully develop as testified by the deposits of the Vallone Scarborato area.

SEDIMENTOLOGICAL DISCUSSION

Essentially all the measured sections of the central and eastern part of the basin are characterised by a basal megaclastic interval overlain by hybrid arenites.

For what concerns the megaclastic interval, it is well represented in the Trionto sections, thanks to the river cut that allows a detailed study in a downcurrent direction. Here, the megaclastic interval can be further divided in three major clastic intervals (named Key-Beds) interspersed by minor arenitic intervals. The three recognised Key-beds from bottom to top are:

1) It has been measured in the Trionto Section, its maximum thickness in the accumulation area is about 130 m. It is clearly wedge-shaped as it pinches toward the north and the bedding attitude changes, with dip angles decreasing toward the south. Unfortunately, due to vegetation, the vertical thickness variations cannot be measured. The interval is very monotonous and only Bed Type 1b or Bed Type 5-1b are present, being the product of gravelly dense flows. Finer grain sizes are uncommon, and when present they are laterally discontinuous and eroded by the conglomeratic beds. Such finer grain sizes where

characterising a sector of the basin localised down-current (see Section 4). This body indicates a phase of sedimentation characterised by repeated catastrophic collapses of basin margins.

2) It is well represented in Sections 1 and 3. In the former section it is characterised by breccias with an abundant coarse sandy matrix, while in the latter section the basal part of the Key-bed is represented by matrix-poor breccias with an increase of coarse sandy matrix up-section. The thickness increases down-current (toward the south). Also, this Key-bed is essentially the product of gravelly dense flows. These flows are characterised essentially by a horizontal grain size segregation, well represented in the area, while in the depocentral areas (Section 3) coarser grain sizes (up to boulder-size clasts) are dominant over pebbles and very coarse sands, which are instead characteristic of the proximal areas (Section 1). In this Section the by-pass of sandy dense flows can be also identified, thanks to the presence of the F5 facies, which locally abruptly overlies the top of some F3 intervals.

3) It is very well represented in Section 3 while in Section 1 it is represented only by two beds. This Key-Bed has a wedge-shape as well, with maximum thickness measured in the southern section. In Section 3 the basal interval is represented essentially by F3 facies (Bed Type 1b) characterised by up to boulder-size clasts being the product of gravelly-dense flows. Up-section, F3 and F5 facies alternations can be appreciated , evidencing the by-pass of sandy dense flows. The most important feature of this outcrop is the pervasive plastic deformation, generated by the emplacement of clastic intervals, evidencing a strong interaction with the substrate. Despite the general wedge-shape of the Key-bed is well defined by its pinch-out toward the North, the local variation on the dip-attitude from the South to the South-East suggests a local lobate shape. In the SP 177 section this interval is represented by finer grain sizes (pebble-size clasts in a coarse sandy matrix), indicating horizontal grain size segregation. In this section the presence of F6/F7 facies, capping the F3 facies through an erosive surface, indicates the by-pass of both the sandy dense flows and the co-genetic turbulent flows. The sedimentary structures in the coarse sands (F7 and F6 facies) are the product of a decrease in velocity of the flows that is here generated by the presence of the clastic body acting as an obstacle, producing small hydraulic jumps. These phenomena are documented also in other basins where MTDs are present (see Tinterri & Piazza, 2019).

The composition of the clasts of the three Key Beds is indicative of the progressive unroofing processes, at least in the Trionto area. In the basal part of the first Key bed, clasts pertaining to the

Caloveto Group, Sant'Onofrio Subgroup and the basement are present. The relative abundance of the clasts pertaining to the Mesozoic sedimentary cover decreases in the middle Key Bed. In the third Key Bed they are completely absent. The very thin Mesozoic sedimentary cover was therefore progressively dismantled by the older collapses, so that the sole available source for the last key bed was the exhumed basement. A similar situation, although in a rift basin setting, was also recognised by Santantonio *et alii* (2016) for the middle Pliensbachian clastic wedge in the Cozzo del Morto area, near the village of Paludi, where the basal intervals are characterised by both Mesozoic- and Basement-derived clasts, while the younger beds only bear granite clasts.

As already highlighted, the Key-bed cannot be traced continuously across the whole basin. Despite this, the pinch-out geometries of the megaclastic interval are evident everywhere, highlighting a general wedge-shape of this interval with a lateral continuity of at least 6 km. The megaclastic interval can be therefore described as a continuous wedge-shaped belt deposited at the toe of a long escarpment. This geometry is not typical of a submarine fan. Among the proposed model to describe deep-water turbiditic systems (Reading & Richards, 1994), the one that best fits the Longobucco basin architecture is that of a linear-source gravel-rich slope apron, which is unfortunately the less well studied and known both in fossil and in modern examples.

As highlighted in the definition of apron proposed by Stow *et alii* (1983), these bodies are "*multiple or linear sediment sourced deposits fed directly across a non-channelised or "straight" gullied body*". Consequently, the typical punctual source invoked for the "submarine fan" is essentially missing.

The megaclastics in the accumulation area are essentially the product of gravelly dense flows (rock-fall and related rock avalanching in the case of matrix poor breccias and debris-flows in the case of matrix-supported breccias) and are essentially the product of the parent flow itself. Up-section the products of the sandy dense flows are also observable.

A special mention must be made for the Laurenzana section, where peculiar characteristics of the deposits can be observed. Here the interval can be divided in two parts, both characterised by bipartite flows. The basal part of each interval is a matrix-supported deposit characterised by plastically deformed elements rafted from the substrate (hybrid arenites and marls). This deposit can be referred to F2 facies of Mutti *et alii* (2003) or to an hyperconcentrated flow. Their presence suggests a different original textural composition of the parental flow (different amount and type of sediment incorporated within the flow). Essentially the F2 facies is the result of the presence of greater amounts of fine mud with respect to the F3 facies. This finer component can also be incorporated within the flows during their down-stream due to the erosion of a muddy substrate. These deposits are typically found at the base of clast supported intervals and in their frontal part, and effectively sustain the entire flow due to dispersive forces related to fluid excess pressure that reduces both the basal and internal friction, enhancing slide mobility. These flows also have an impressive erosional power, as highlighted by the presence of huge megablocks pertaining to the substrate. These flows account for complete stratal disruption of thin layered sequences, with the progressive development of the so-called block in matrix fabric. A similar facies association has been found in the Trionto River, where the arenites are severely deformed and dismembered, developing soft-sediment deformation structures, and huge blocks of the substrate are deformed by pseudo-SC structures.

Probably the Laurenzana deposits correspond to the Key-bed 2, but this tentative correlation is highly speculative.

With reference to the hybrid arenites, previously interpreted in the literature as turbiditic deposits related to the dismantling of a rising orogen, they can be reinterpreted in the light of the facies associations herein presented. The measured sections are characterised by F5 facies and in minor amount by F3 facies deposits, and are thus the product of sandy dense flows and gravelly dense flows respectively.

Some sections are characterised by chaotic intervals, impact structures, amalgamation surfaces, highlighting a strong interaction with the seafloor and flow. These structures are here interpreted as the product of hydraulic jumps related to deceleration of the flows due to the abrupt decrease in the gradient at the base of the escarpment bordering the basin toward the north. Hydraulic jumps favoured flow expansion with the consequent decrease in transport capacity. Only in few cases we have reported F7 facies that testify the by-pass of fully turbulent flows, but the related deposits are today missing in outcrop. An exception is the Vallone Scarborato section, where F8 facies deposits are present. We can thus consider this section as having been localized in a more distal area with respect to the others. This could be further proved by the absence of the basal megaclastic interval in the whole western sector of the basin. The absence of this basal part can be explained by invoking the erosion of this part of the system in this sector of the basin, and the preservation of the distal parts of the Paludi basin, elsewhere buried or tectonically elided.

As highlighted earlier it must be remarked that what we are describing here is a very peculiar system that can be related, at least for the basal megaclastic interval, to a gravel-rich apron of Reading & Richards (1994) (Fig. 68) (the upper arenaceous interval will be briefly discussed below).

As already mentioned, such systems are poorly studied, but comparable systems are those studied by Surlyk (1989) in East Greenland, North Sea and Scotland. These basins are hangingwall blocks of normal faults, their formation being linked to a middle/late Jurassic extensional tectonic phase that affected the North Atlantic realm. Surlyk highlighted that these systems, especially in their embryonic stages, are characterised by wedge-shaped breccias, both at the toe of slope and along the

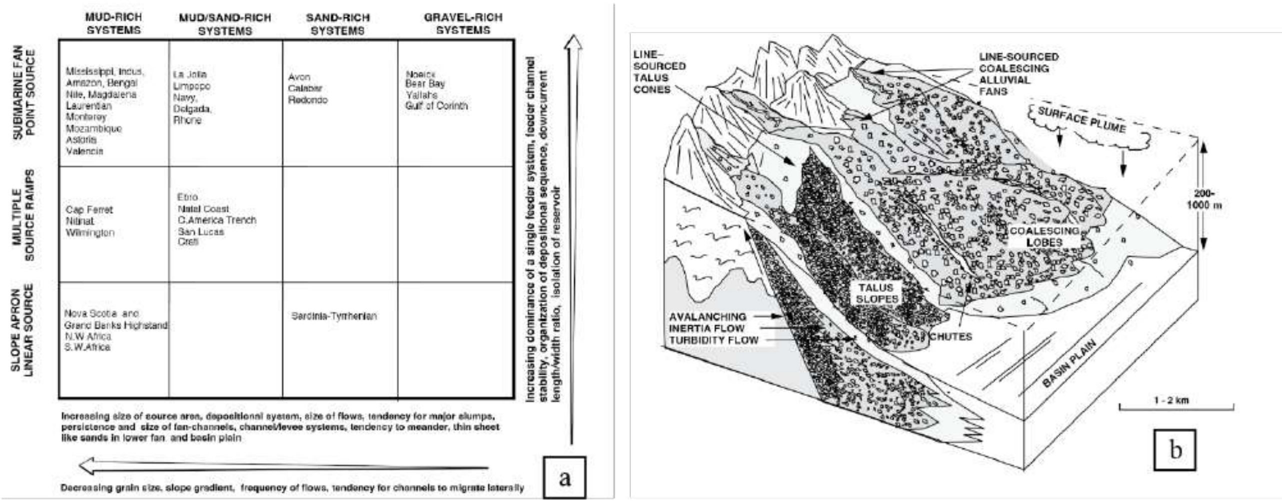

Figure 68: a) classification of Reading & Richards (1994) based on grain-size and source type; b) gravel-rich apron.

basin axis. Such submarine bodies were not connected with subaerial fan deltas during most of their evolution, and if at all present these deltas must have been steep and energetic. If fault activity persisted, the deposits were characterised by marked aggradation and only minor progradation. Real channels were essentially absent and eventually only short-lived chutes or scours were present. Being the triggering mechanisms linked especially to tectonic pulses along fault escarpments, we are dealing with systems fed by a linear and continuous source and not by a punctual one. These pulses are short-lived and catastrophic, which influences the rate of sediment supply. The main emplacement mechanisms are thus mass transport processes, eventually evolving into turbiditic flows. Being a fan-delta not well developed, clasts are mainly sourced from the escarpment itself, they are obviously coarse as mud supply is very limited, and this consequently impacts on the overall efficiency of the system. This last consideration is essential also for the interpretation of the sedimentary processes that are behind the arenaceous part of the Paludi Formation. We have already stressed that also this lithofacies is essentially the product of dense flows, and the upper turbulent part is essentially missing. Only locally have some by-pass related structures been detected, and the finer grain size with the sedimentary structures typical of turbiditic currents are virtually missing. As highlighted in the Torno sectionnumero, even the more distal sections are devoid of such deposits. Probably turbiditic flows were, at least in an initial stage, not able to form, being the finer grain size missing or very limited. As highlighted by Mutti *et alii* (2003), high-efficiency turbidity currents can develop only if the parental flow is rich in fine grain sizes. The Authors stress that in young, tectonically active extensional basins with uplifting margins, sedimentation is essentially driven by catastrophic collapses of the footwall and as such is mostly characterized by coarse grain sizes (Fig. 69). High-efficiency turbidity currents may develop later, as a forced (tectonically driven) lowstand is generated, thanks to increased elevation of drainage basins and buckling of flood generated flows. This uplift may drive sediment failure along the shelfal and slope areas, increasing sediment concentration within the flows. All the above can generate highly efficient turbidity currents, which are missing wherever fine grain sizes are not added to the parental flows.

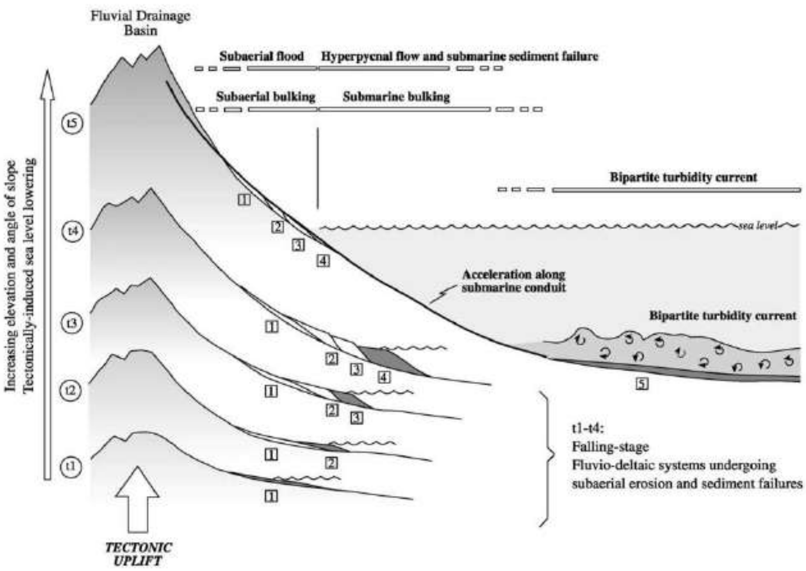

Figure 69: Due to tectonic uplift and gradual enrichment in fine material, high-efficiency turbidity systems can develop only in a tardive stage of the evolution of the basin (from: Mutti et alii, 2003).

Furthermore, Reading & Richards (1994) stress that in those systems where a punctual source is missing (both in multiple-sourced and linear-sourced systems), the sediment inflow is portioned and its penetration into the basin is thus limited.

In the case of the Paludi basin the source area of the sand-sized grains would be a mixed shelf, as testified by the presence of Larger Benthic Forams, sub-rounded Qz and granite clasts. The geometry, age, and position of this shelf are unknown as this part of the depositional system is completely lost, due to quaternary(?) erosion. It is important to note that grains pertaining to high-grade metamorphic rocks, typical of the Alpine chain edifice outcropping in Calabria are totally missing. This information, paired with the emplacement mechanisms that indicate very short run-out distances, allow to infer a local provenance for the siliciclastics.

The arenaceous part of the succession is hard to classify in the reference scheme of Reading & Richards (1994) as the measured sections do not allow us to describe the lateral geometry of this part of the Paludi Formation. We can safely place our system in the sandy sector of their model, and being these deposits partly intercalated with the megaclastics we can confidently infer a linear or multiple sources. Our system is probably something that is straddling the “sand-rich ramp” and the “sand-rich apron”. Both these systems are backed by a narrow fault-controlled shelf that generates low efficiency systems with very limited downslope extent.

Similarly, we can tentatively relate our system to the “mixed turbidite systems” (Mutti *et alii*, 2003) that are: “*relatively small and generally sand-rich depositional systems sharing several* *characteristics with basinal turbidites, but differing from these by showing a more immature facies development (cf. 'poorly-efficient turbidite systems')*". In the original meaning of Mutti and coauthors, these deposits are associated with deltaic deposits, that are here apparently missing. These depositional systems are related to faulting and tectonic uplift along basin margins, which generated a steepening of the depositional profile enhancing erosion and sedimentation.

GENERAL CONSIDERATIONS

From the measured sections the following considerations can be made (see Fig. 70):

i) A megaclastic interval is present along a more than 6 km long belt oriented ca EW;

ii) This megaclastic interval is represented by many major pulses (at least three) that are sedimentologically similar (matrix-poor breccias evolving upwards to matrix-rich breccias), apart from the Laurenzana outcrop where the clastics show peculiar characteristics;

iii) The basal clastic interval rests on an escarpment carved in the Hercynian phyllites;

iv) All the clastic intervals show pinch-out geometries, and their hickness decreases to 0 m at higher elevations (toward the North). This describes a wedge-shaped geometry for the clastics;

v) The clastic interval dips consistently toward the South, with dip angles decreasing toward the top of the interval. At least in the Trionto Area, a lobate shape of some deposits can be invoked considering the change in the bedding attitude (dip toward the East) in a cross-current direction;

vi) Progressive unroofing processes can be evidenced through the analysis of the composition of the clasts in the Key-beds. Clasts of the Mesozoic cover are common in the lower part of the Paludi Fm. and gradually disappear upsection, eventually being replaced entirely by basement clasts. This demonstrates that the thin sedimentary cover represented by the Caloveto Group was totally dismantled through repeated catastrophic collapses of submarine escarpment margins;

vii) The hybrid arenites are essentially represented by the F5 and less commonly F6 and F7 facies of Mutti *et alii* (2003) and are thus the product of sandy dense flows. These facies are proximal to the source as also confirmed by the mapping of the unconformity surfaces;

viii) Sourcing was local, as the clasts contained in the Paludi Fm pertain to the unmetamorphosed Mesozoic cover or to the basement;

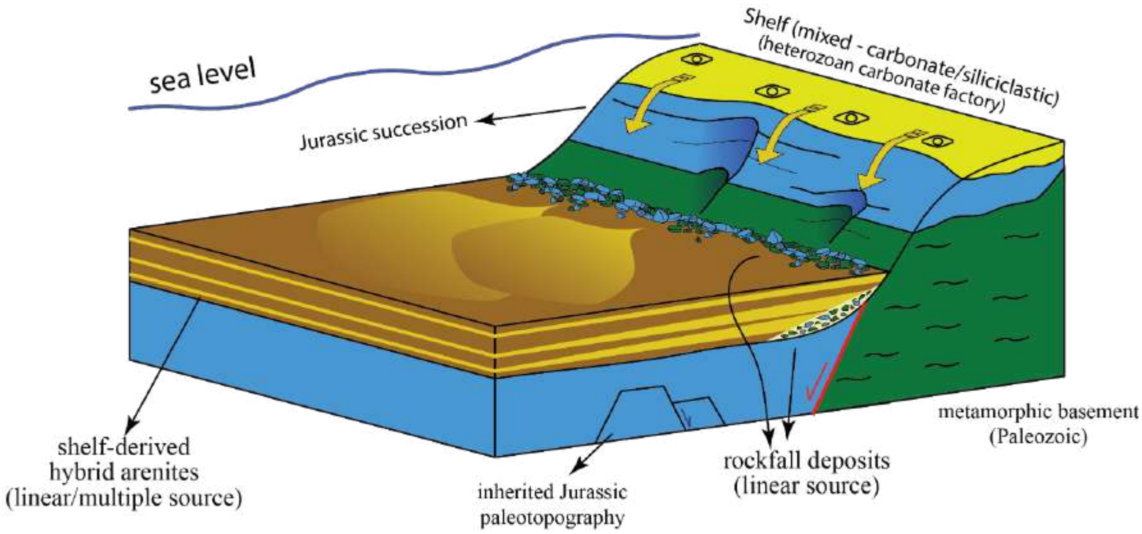

Figure 70: Cartoon depicting multiple-source sedimentation in the fault-controlled Paludi Fm basin.

ix) In the sections located in the South, the presence of structures referrable to interaction with the substrate are common, indicating halting/freezing of the sandy dense flows;

x) Turbiditic flows, probably, at least in the basal part of the Formation were not able to form, being the system immature and the parental flows not sufficiently rich in fine grain sizes;

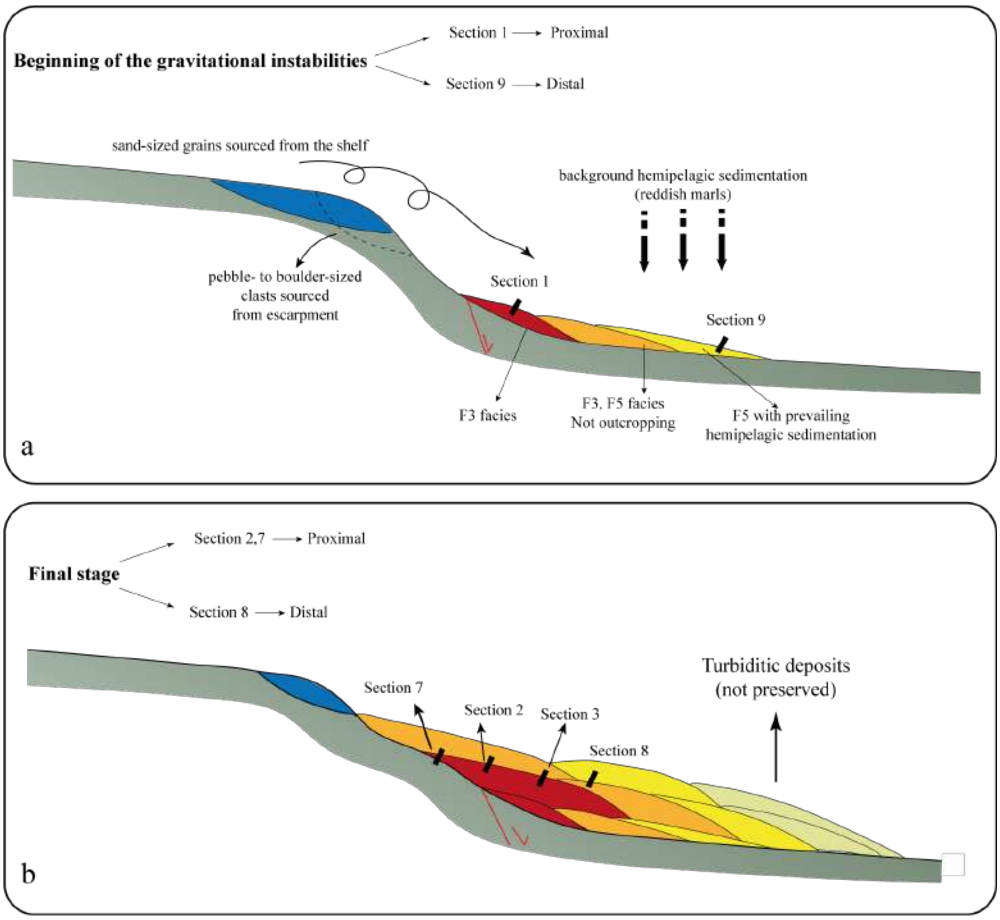

Figure 71 – Depositional architecture of the Paludi Fm.

xi) A mixed siliciclastic-carbonate shelf had to back the depositional system, feeding the sandy dense flows, as highlighted by the presence of LBF and rounded clasts of granite and polycrystalline quartz. This shelf is unpreserved;

xii) The triggering mechanisms for such flows where repeated seismic shocks along fault-block escarpments.

In Fig. 71 the depositional architecture of the Paludi Fm is resumed.

Basin Modelling

One of the blind spots of the tectonic-stratigraphic history of the Longobucco Basin concerns the timing of the compressive stage, since a huge stratigraphic gap exists between pre- and post-thrusting units (see “Geological setting”).

In order to fix this issue, and with the more general goal of reconstructing the burial and exhumation history of the Basin, X-ray diffraction analyses of clay minerals, organic matter optical analysis and apatite fission tracks have been performed at the Alba Laboratory of the University of Roma Tre and in collaboration with Prof. L. Aldega (Sapienza). The collected samples are shown in figure 72 with respect to their stratigraphic position. Data are shown in the attached tables.

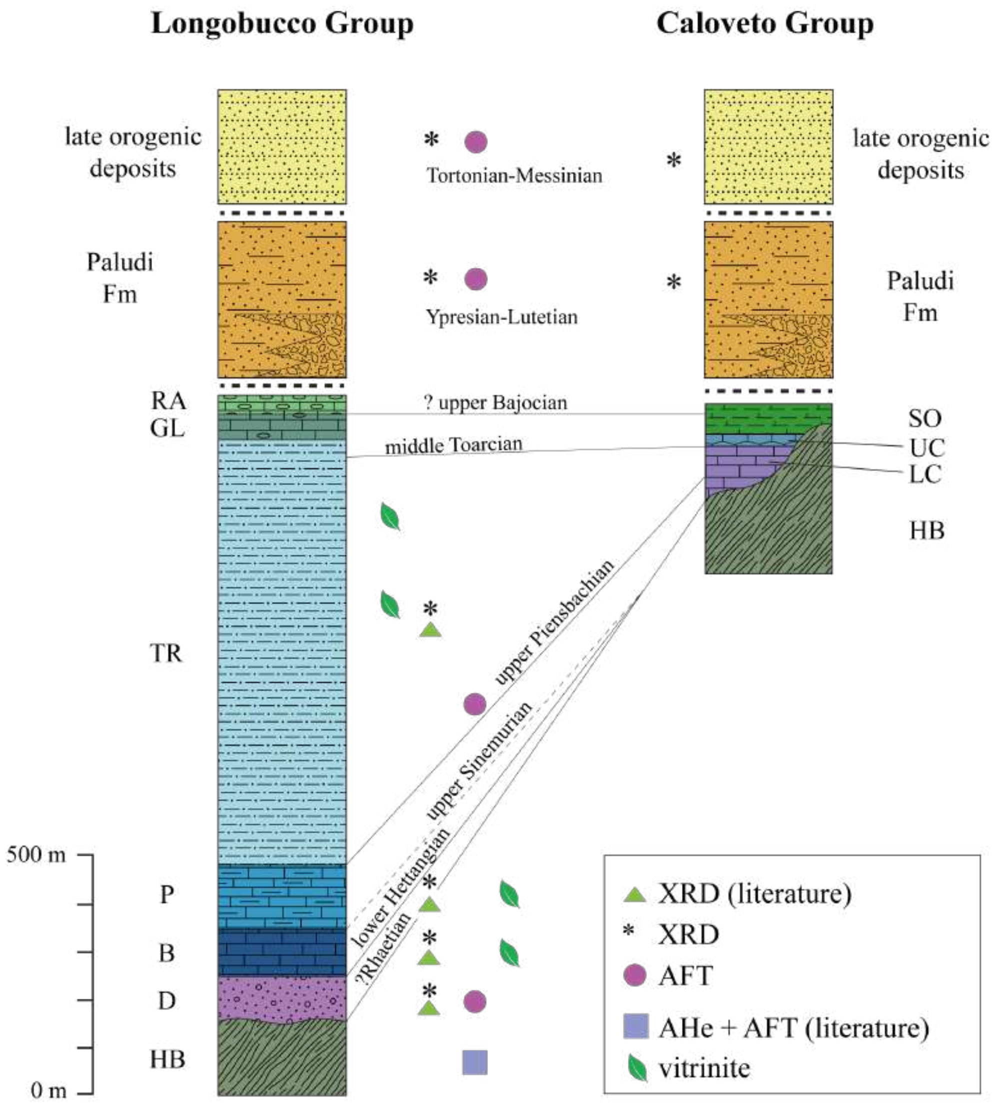

Figure 72: Samples and their stratigraphic position. HB: Hercynian basement; D: Torrente Duno Fm; B: Bocchigliero Fm; P: Fosso Petrone Fm; Tr: Trionto Fm.; GL: Grey Limestone with black chert; RA: Radiolarian Chert; LC: Lower Caloveto Fm; UC: Upper Caloveto Fm; SO: Sant’Onofrio Subgroup.

PALEOTHERMAL CONSTRAINTS

Paleothermal constraints are provided by the X-ray analyses of mixed-layered clay minerals and by the organic matter optical analysis.

Reactions in clay minerals are irreversible under normal burial conditions, so that uplifted and exhumed sedimentary succession generally retain indexes and fabrics indicative of thermal maturity and maximum burial. Clay minerals are mainly sensitive to temperature, and the use of mixed layers illite-smectite (I-S) and the transformation sequence dismectite-random-ordered mixed layers I-S (R0)-ordered mixed layers I-S (R1 and R3)-illite-di-octahedral K-mica (muscovite) as indicator of maximum paleotemperature conditions is generally accepted. In fact, changes in the composition of mixed layering, layer expandability, and I-S ordering are empirically related to temperature changes due to burial (Hoffman & Hower, 1979; Aldega et *alii*, 2014).

The second technique allows quantifying the percentage of reflected light of organic macerals (VR%). The internal organization of vitrinite-huminite macerals increases as a function of burial and this is mirrored in a higher efficiency in reflecting light. Vitrinite reflectance measures can be derived from organic-rich stratigraphic intervals. The whole Mesozoic succession contains organic matter-rich layers with plant remains, being the depositional system backed by a continental shelf sourcing resedimented material (see geological setting). Sampling of the organic-rich intervals has been performed in the second and third tectonic units from three distinct lithostratigraphic units belonging to the Longobucco Group: the Bocchigliero, Petrone and Trionto formations. The Cenozoic units (both the Paludi Fm and the late orogenic siliciclastic units) are completely devoid of organic-rich levels, consequently vitrinite reflectance analyses have not been performed for this time interval.

Sampling of clay rich intervals was carried out in the Longobucco Group, in the Paludi Fm and in the late-orogenic siliciclastic deposits, which enabled us to draw comparisons among different lithostratigraphic units.

The obtained values of clay mineral assemblages for the Mesozoic succession and the Paludi Fm are unexpectedly comparable, suggesting that both the Mesozoic succession and the Paludi Fm experienced similar levels of thermal maturity in deep diagenetic conditions. Conversely late orogenic siliciclastic deposits show low levels of thermal maturity in early diagenetic conditions. Consequently, both the Longobucco Group and the Paludi Fm underwent a deep burial that could have been caused by either tectonic or sedimentary loading. This loading must therefore have occurred in post-Paludi/pre-Serravallian times, that is predating the late-orogenic succession.

EXHUMATION

Fission tracks are linear damage zones produced on the crystalline lattice of a mineral (apatite or zircon) by particles released by spontaneous decay of ^{238}U. Such fissions are not stable, their length decreasing due to a process called "annealing". At temperature higher than 120°C the tracks are completely annealed, while between 60-120°C this process is only partial (PAZ → Partial annealing zone).

Consequently, the statistical measurement of the length and density of fission tracks provides information about the exhumation history of the mineral.

The results show that the Paludi Fm was exhumed at about 22 Ma, and the Mesozoic lithostratigraphic units between 15 and 13 Ma. The exhumation ages have been used to constrain the timing of exhumation (see below).

U/PB DATING ON CALCITE VEINS

As already mentioned in the Geological Setting section, multiple tectonic phases have been recognised in the Basin. These phases are more or less precisely constrained thanks to stratigraphic evidence.

We tried to better constrain the age of the main tectonic events through U/Pb dating on calcite veins. Analyses were performed in collaboration with the "Petrology, Geochronology & Structure" laboratory of the Santa Barbara University (California).

1 – Toarcian extensional phase: this phase corresponds to the second pulse of extensional tectonics that took place in the rift basin in the Early Jurassic. As an effect of this phase, the shallow water carbonate bodies, fringing the exhumed basement, were drowned. In the Cozzo di Mastro Pasquale Area, one of such faults is beautifully preserved. The area has been mapped in detail by Santantonio & Fabbi (2020), who reconstructed the stratigraphic relationships between pre- syn- and post-rift deposits and the complex architecture of the depositional system cut by the Toarcian fault (Fig. 73). The sampled structure displaces the Lower Caloveto Fm, forming a stepped surface, and is sealed by the Aalenian *Zoophycos* marls. Slickenfibers are present, displaying a pitch of 90°. The Fault attitude is 205·51, 208·62, 182·63.

2 – Post-orogenic phase(s). Across the Basin, a number of normal faults has been mapped, displaying in some cases also a horizontal component. The geometrical relationships of such structures with the succession are well exposed and these structures can be considered as post-Tortonian since they displace the late orogenic siliciclastic deposits and can be consequently considered as post-orogenic. These faults are especially well exposed in the northernmost sector of the Basin.

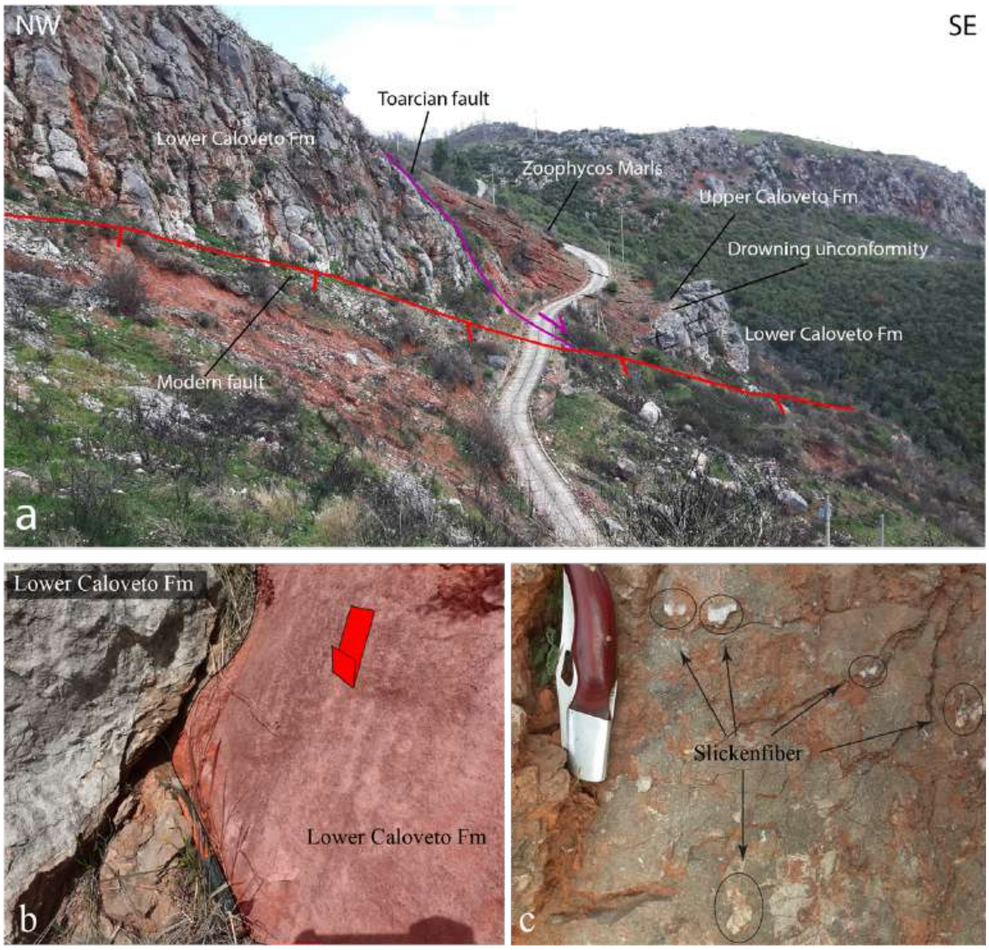

Figure 73: a) Structural scenario of the Cozzo di Mastro Pasquale area (from Santantonio & Fabbi, 2020); b) Fault plane displacing the shallow water limestones of the Lower Caloveto Fm; c) The plane displays slickenfibers sampled for U/Pb dating.

In the Caloveto area, one of these structures has been sampled. The fault plane shows a lateral continuity of about three kilometres, and it has been sampled in two distinct localities at the outcropping tips of the structure. In the first locality, near Contrada Dema, the fault attitude is 225·48 and it juxtaposes the marly member of the late orogenic deposits to the sandy member. The fault plane is characterised by slickenfibers with a pitch of 0°. In the footwall block, several secondary planes are arranged forming overlapping left stepping jogs, having a subvertical attitude and striking N330. Such planes develop extensional pull-apart structures filled with calcite (Fig. 74).

Toward the South-East, at Cozzo di Mastro Pasquale, near the abovementioned Toarcian faults, the same fault plane juxtaposes the sandy member of the late orogenic deposits on the Lower Caloveto Fm. Here, both the main plane and some Riedel planes in the footwall block, bearing slicken-fibers, have been sampled (pitch 90°).

3 – Undetermined extensional phase. In the southern part of the basin, several extensional faults can be mapped (Fig. 75). In this part of the Basin, unravelling time-constraints for the activity of such structures is difficult for two main reasons:

- The upper Miocene succession is eroded;
- They only cut the Mesozoic succession/Paludi Fm;

Figure 74: Calcite infill of an extensional jog at Contrada Dema.

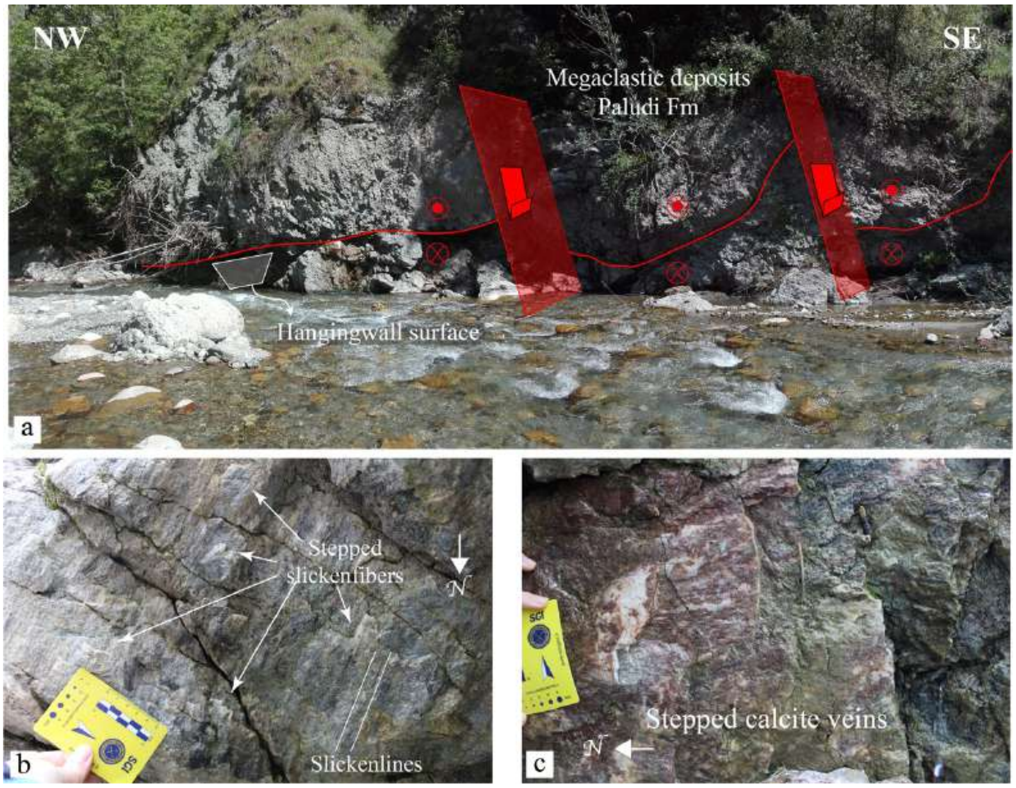

Figure 75: a) one of the sampled extensional faults pertaining to the unknown extensional phase. The sampled structures dip gently toward the north/northeast and is cut by high angle transversal normal faults; b) and c) the exposed hangingwall displays stepped slickenfibers (indicating a movement toward the northwest) and calcite veins.

We therefore attempted to assess the absolute age of these structures in order to establish if they are pre-or post-orogenic normal faults.

Unfortunately, the results of these analyses are not satisfactory, as the error brackets are huge. Attempts can be made to reduce this error, as two distinct clusters are seen in the graphics, which could highlight a secondary and younger opening of the system. This overprinting probably generated the abovementioned errors.

The most reliable data point to a late Miocene activity, as one sample gives an age of 8Ma.

This age corresponds essentially to the Tortonian/Messinian boundary. In the Rossano Basin this is a turning point in the evolution of the sedimentary cycle. In particular, two cycles can be recognised: i) transgressive cycle characterised by a deepening trend and represented by deltaic to deep marine sediments (?Serravallian-Tortonian); ii) coarsening upward trend related to a major pulse in clastic sedimentation as testified by the "Molassa di Castiglione" made up of basal conglomerates and breccias. The contact between the two cycles is an unconformity, corresponding to an erosional surface (Barone *et alii*, 2008).

This time constraint, provided by U/Pb dating, fits well with other data derived from stratigraphy. We used this age in the modelling to constrain a second pulse of erosion. It must be remarked that the Serravallian-Tortonian succession in the Longobucco area is laterally continuous and no evidence of displacement has ever been found. The only evidence of deformation is a mild tilting, generating a monocline gently dipping toward the north. This tectonic phase, consequently, affects the Longobucco area by generating uplift and re-opening of the veins systems, but not any major displacement.

PREVIOUS STUDIES

The first data collected in the Calabrian Arc were obtained by Thomson (1992), during his PhD thesis. The Author analysed AFT and ZFT from the basement of the Calabrian Arc. His data came from the Calabride Complex (also from the unit that underwent Alpine metamorphism) but no suitable samples were collected from the Liguride Complex. The results of his work served to assess that part of the High-Grade metamorphic unit (HGM) and part of the Hercynian granites were deeply buried at temperatures higher than 250 °C, as they do not retain ZFT. Part of the Hercynian granites were, at the same time, also buried at temperatures between 120°C and 250°C as they retain ZFT only, and no AFT. The Low-grade metamorphic units, with their sedimentary covers, only experienced temperatures lower than 120°C, thus retaining both AFT and ZFT. Thomson in his papers (Thomson, 1994a, b) demonstrates that the HGM and the Hercynian granites underwent very rapid exhumation between 35 and 15 Ma. The Author interpreted this process as the product of collapse of the orogenic wedge due to extensional detachments (see also Rossetti *et alii*, 2001). On the contrary, the Low-Grade metamorphic units (LGM), with their sedimentary covers, did not experience such a rapid exhumation and erosion, being these Units already above the PAZ. During his PhD thesis, Thomson tested the Fission Tracks method also on the Stilo Capo d'Orlando Fm (SCOF), being the age of this Unit coeval with the timing of increased denudation detected though FT analysis on the basement. The results of this work demonstrate that the provenance of the Stilo Capo d'Orlando Fm was from local basement units only, and that at the beginning of deposition of the SCOF, 2 km of paleo-relief existed in the basement of the Calabrian Arc, and 5 km of vertical section had been eroded from the Hercynian basement since the Oligocene.

In the Longobucco Basin, Perri et *alii* (2008) studied clay mineral assemblages of the Mesozoic Longobucco Group. Their data, based in particular on mixed-layer I/S, show a lithostatic/tectonic loading of 4–6 km for the Longobucco Group.

Two studies of the same research group carried on in the Longobucco Basin are those by Vignaroli *et alii* (2012) and Olivetti *et alii* (2017), centered respectively on AFT and U(Th)/He on Apatite. Vignaroli and coauthors identified a main exhumation phase between 18 and 13 Ma, and correlate it with a thrusting event. The Authors highlighted that this compressive event was coeval with the extensional phase active in the Coastal Chain. These data were consequently used to prove that, at least in northern Calabria, compression, which had started in the Cretaceous/Paleocene (66 Ma), progressed continuously so that it reached the eastern sector (facing the Ionian Sea) some 30 million years later (see Rossetti *et alii*, 2001). Some considerations on propagation rates of a thrust front are herein mandatory. The linear distance between Catena Costiera and Sila Greca is about 60 km. If we hypothesize the onset of compression at 66 Ma, the rate of thrust migration would be 1.4 mm/yr. This rate is clearly unreasonably low, especially when compared with the rates calculated for other orogens (Himalaya: 25 mm/yr, Bolivia 15 mm/yr DeCelles & Gilles, 2001; between 15 and 25 mm/yr in the Central Apennines).

Olivetti and co-authors identified an exhumation event taking place over the same time span as that proposed by Vignaroli *et alii* (2012). In their discussion paragraph, Olivetti *et alii* (2017) state that such event was very rapid. Problem is, the Olivetti and coauthors' ages for AHe are older than those of Vignaroli *et alii* (2012), which is a paradox as the AHe ages should always be younger than AFT ages, since they record exhumation from a lower temperature (AHe – 60°C vs AFT 110°C). Consequently, in my opinion, exhumation rates for this event are objectively poorly constrained. In addition, this is not in agreement with data from Thomson (1994a), who demonstrated that the LGM with its sedimentary cover was already above the PAZ at 30 Ma, showing the lowest exhumation rates among the different types of Calabrian basement ("Alpine" *vs* "HGM and LGM).

Another result presented in Vignaroli *et alii* (2012) and Olivetti *et alii* (2017), achieved thanks to numerical modelling, is the computation of slow erosion rates (0,1 mm/yr) in the 18 Ma (age of thrusting) to 1 Ma (Pleistocene) time window, followed by a dramatic increase up to 1 mm/yr. The modelling performed by these Authors showed also the relief, at the end of the exhumation phase, was between 2.4 to 3 times that observed today. This is in agreement with erosional data from Thomson (1994a).

Olivetti *et alii* (2017) also investigated the western and southern Sila Range, concluding that the onset of abrupt uplift was not detectable in such areas. These data clearly disagree with those of Thomson (1994a), who highlights in the HGM units of western Calabria a major, fast exhumation phase between 30 and 15 Ma.

When we compare our AFT data with those of Vignaroli *et alii* (2012), it is apparent that our exhumation ages are older, at least for the Paludi Fm. Consequently, the thrusting phase must be considered older than the one proposed by Vignaroli *et alii* (2012). Despite the differences in exhumation ages, the interpretation offered by Vignaroli is not conservative, as an exhumation age of 18Ma implies that the thrusting event had to be older.

Two points arise from the above discussion:

- Across the whole Calabrian Arc, the basement underwent exhumation between 30 and 15 Ma
- Both the basement and the sedimentary cover of the Longobucco group experienced a load of at least 4km

The interpretation of the causes that led to exhumation are indeed opposing. Thomson (1994a) considers this as the product of extensional tectonics, while for Vignaroli *et alii* (2012), for the Longobucco area only, it is held as the product of compression. Furthermore, Thomson's data are in contrast with those of Olivetti *et alii* (2017), as the former proposes more rapid exhumation rates for the LGM.

BASIN MODELLING

The burial history of the sedimentary succession, from Early Jurassic to Holocene, was modeled using the software Basin Mod-1D. The results of modelling are shown in figures 76 and 77 and the Thermal maturity curve calibrated against illite content in mixed layers I–S and vitrinite in figure 78. Figure 76 shows the burial history since Rhaetian, while figure 77 focuses on the Cenozoic phase. General assumptions are: 1) rock decompaction applied to clastic units only; 2) thermal evolution is mainly influenced by sediment thickness rather than by water depth; 3) thrusting is considered instantaneous when compared with the duration of stratigraphic successions;

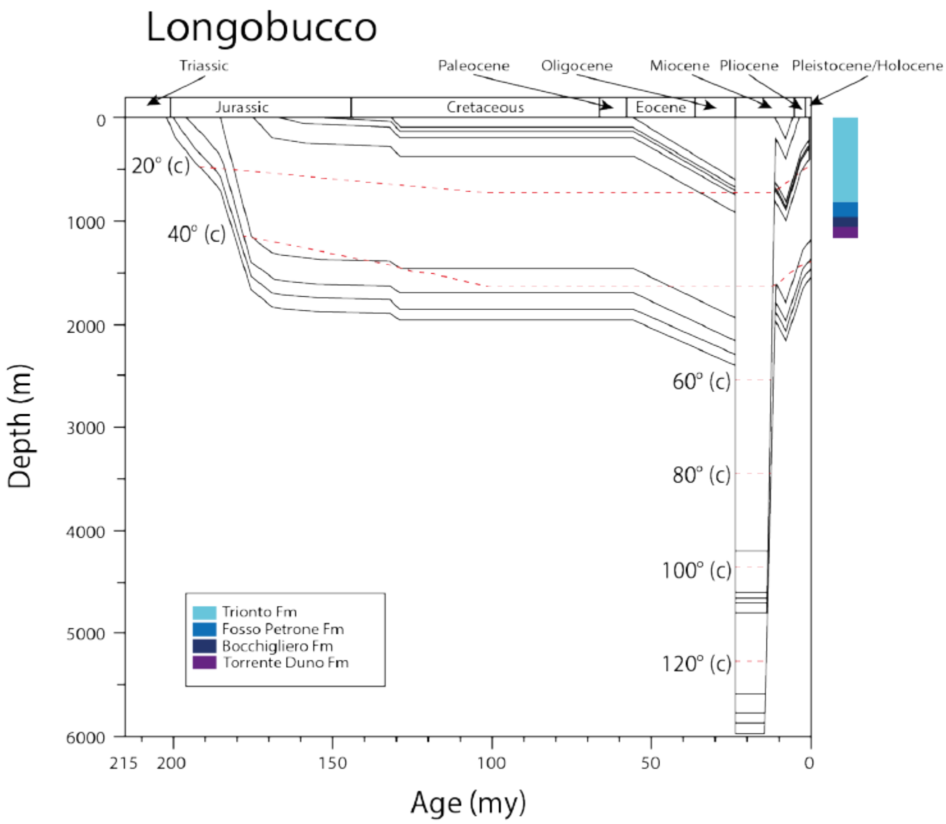

Figure 76: Burial history since the Triassic/Jurassic boundary.

4) a variable geothermal gradient was adopted as a function of the tectonic setting (45°C/Km during rifting; 22°C/km during compression) The thickness of lithostratigraphic units was obtained from field data by Santantonio & Fabbi (2020), Innamorati & Santantonio (2018), Santantonio *et alii* (2016) and Young *et alii* (1986). An issue that arose during modelling concerns the hypothetic thickness of the missing Mesozoic to early Paleogene deposits, corresponding to the timespan of the hiatus (~80 Ma) existing between the top of the Longobucco Group and the base of the Paludi Fm. The following considerations can be made:

- If we consider a canonical post-Hauterivian (=youngest age detected in the Longobucco Group) Tethyan succession, we can infer a total missing thickness of ~350 m. Considering that typically the Longobucco units are rich in resedimented material, we could stretch the thickness up to 500 m
- The plastic olistoliths discovered in the Paludi Fm confirm that the missing units were similar to those of the UMS basin (Fucoidi marls, Scaglia Bianca).

Summarizing, most of the missing sedimentary succession has been eroded parallel to deposition of the Paludi fm in a subsiding basin. It is therefore impossible for the software to process basin subsidence and, at the same time, erosion, that typically is associated with uplift. We therefore uploaded this missing thickness in the modelling.

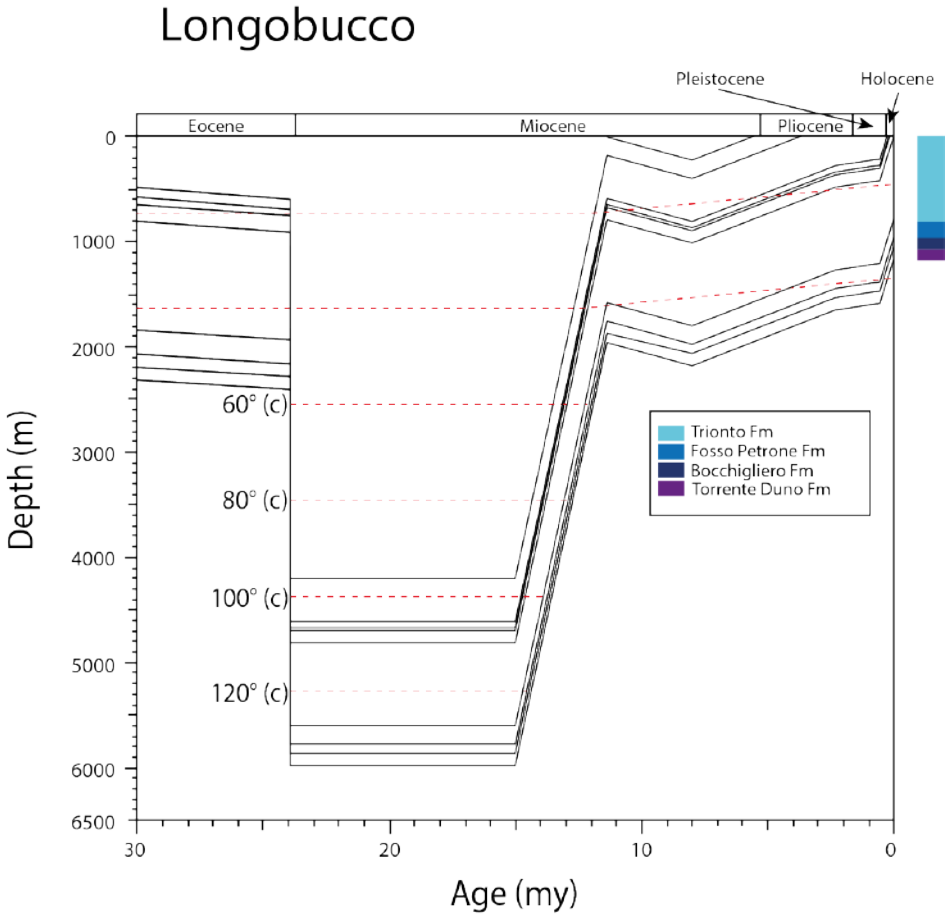

Figure 77: Close-up on the Cenozoic burial history.

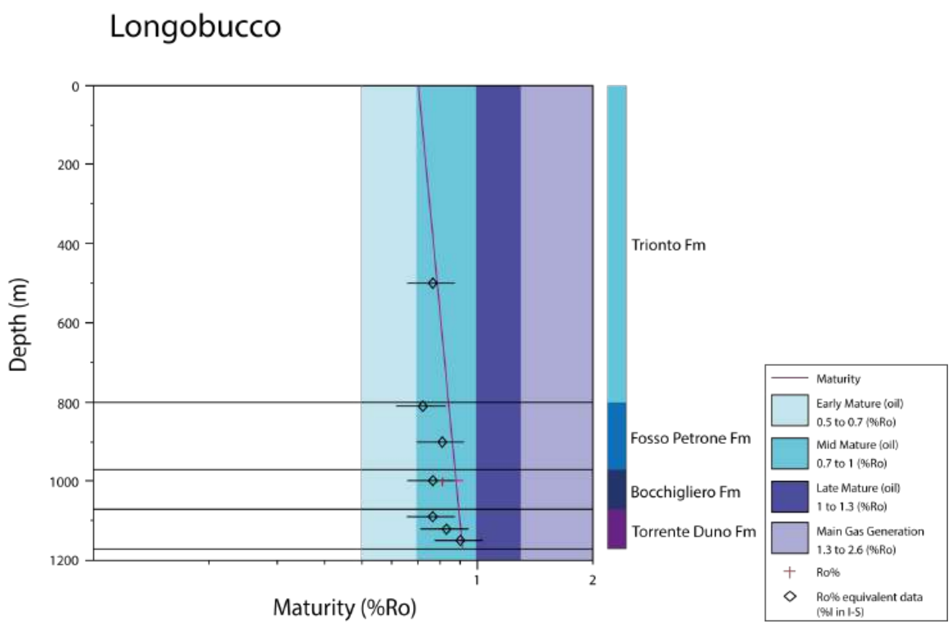

Figure 78: Thermal maturity curve calibrated against illite content in mixed layers I–S and vitrinite.

In any event, as it will be clarified below, additional loading from the missing Mesozoic/Paleocene succession can be considered as negligible to achieve the very high levels of thermal maturity indicated by paleothermal indicators and apatite fission track data.

The most challenging issue was the age of thrusting. This age must be comprised between 37 Ma (early Priabonian: youngest age of the Paludi Fm) and 22 Ma (exhumation of the Paludi Fm). In accordance with data from other areas of the CPA, we place the thrusting phase in the late Oligocene, at 24 Ma.

We performed four distinct models in four areas of the basin, differing for stratigraphy (Longobucco vs Caloveto) and tectonic position (Tectonic Unit 2, 3 and 4). The models are consistent with each other and can be summarized as follows:

The Jurassic history was already well established (Santantonio *et alii* 2016; Innamorati & Santantonio, 2018; Santantonio & Fabbi, 2020), and the burial history essentially confirms it, evidencing an abrupt deepening of the basin in the middle Pliensbachian, being the base of the Longobucco Group (Torrente Duno Fm) buried at >2000m. After the thrusting phase the Paludi Fm was buried at a maximum depth of 4800 m, the Torrente Duno Fm (oldest lithostratigraphic unit) was buried at maximum depths of 6000 m under a 4200 m thick tectonic overburden. The modelling here highlights that the high maturity shown by vitrinite and XRD are the products of tectonic load only. The tectonic slice as of today is completely eroded, with no remnants left anywhere. This is consistent with data obtained from the Stilo-Capo d'Orlando Fm, that highlight the presence a of "Lost Cover" (Thomson, 1994b) eroded and not outcropping anymore, but detectable within the clasts of the SCOF.

Since the Aquitanian until the Serravallian, the nappe pile was exhumed at a constant rate of 0,31 mm/yr and the >4000 m nappe stack was eroded along with parts of the Mesozoic/Cenozoic units. This rate is a little bit higher than that proposed by Olivetti *et alii* (2017) (0,1 mm/yr), but, as already proposed by these authors, it confirms slow and steady erosion that cannot be attributed to tectonic denudation, such as extensional detachment, as proposed by Thomson (1994a) and Rossetti *et alii* (2001) for western Calabria.

An additional consideration is now needed. The whole CPT, south of the Catanzaro Through, is characterised by the presence of the SCOF, that is apparently missing in the northern sector of the Arc. In southern Calabria, Oligo/Miocene times correspond, therefore, to a period of erosion paired with subsidence of the SCOF Basin. In the Longobucco Basin, such phase of subsidence seems to be completely missing, being conversely a time of uplift only. Detrital mode analyses of Upper Miocene units in the Longobucco area reveal the presence of clasts pertaining to the Basement, the Longobucco Cover and Paludi Fm, and did not detect the presence of other types of clasts (Barone *et alii* 2008). Another observation can be made considering the geology of the whole CPT. From South to North the hiatus that elapses between Oligocene and Miocene increases drastically, since the Calcareniti di Floresta have a duration that gradually increases to eventually cover the whole hiatus. Sicily and southernmost Calabria: Burdigalian – Langhian; Aspromonte: Langhian; Sila: this unit is missing (see scheme in Fig. 79). This probably highlights a contrasting subsidence trend in the CPT, where uplift dominates to the north, while the southern sector is subsiding. This probably indicates that the SCOF was never sedimented North of the Catanzaro Line exhumation rate, as the paleotopography, inherited after the Early Jurassic rifting, was complex and the Quaternary uplift caused erosion of vast sectors of the Meso/Cenozoic depositional system. For example, in the Sant'Onofrio area, the upper Miocene succession onlaps the basement directly. One should bear in mind that 1500 m of Jurassic succession plus 4000 m of tectonic slice have been eroded away, and this would increase the erosional rate. Here, the basement constituted, since the Early Jurassic, the footwall of a Jurassic rift fault, being consequently raised with respect to the basement underlying the very thick hangingwall-block basinal successions by 1.5-2 km. Furthermore, the onlap wedge of the basinal units is extremely thin as the formations pinch out dramatically towards the basin margins, meaning a reduced lithostatic load.

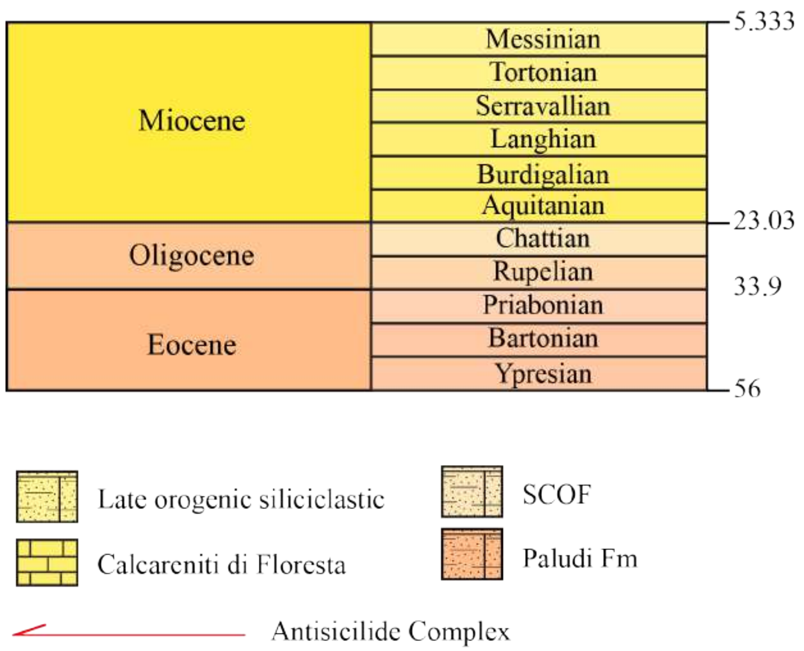

Figure 79: Stratigraphic scheme of the Eocene-Miocene interval of the three sectors of the CPT. Please note the extent of the hiatus in Northern Calabria (i.e., Longobucco Basin) that gradually decreases in the southern sector of the CPT.

The same occurs in the Caloveto area where the exposed basement, draped by thin carbonates and pelagites, is a "footwall-block basement". Furthermore, erosion affected a deformed structure, made of anticlines and synclines. In the Ortiano area, for example, the uppermost Miocene units onlap the Jurassic units directly, since erosion affected here the core of an anticline, while along its flanks the Miocene units onlap the Paludi Fm.

Starting in the Serravallian, the basin underwent another phase of steady subsidence, as testified by the gradual deepening of the depositional setting (Deltaic – coastal – marine). This trend experienced a sudden halt as evidenced by an erosive phase affecting both the Upper Miocene deposits and the Meosozoic cover. As testified by burial history, another phase of major erosion, responsible for the removal of the upper Miocene succession, started in the Pleistocene/Holocene.

Age of compression in the Apennines: A review and a novel approach

The age of the beginning of Apennine subduction is still debated, and two main hypotheses exist in the literature: the "Young Apennines" and the "Ancient Apennines". The first hypothesis invokes the development of the Apenninic slab starting in the Oligocene, following the break-off of the Alpine slab. The second hypothesis envisages the occurrence of two coeval opposite subductions since the Eocene (or earlier: Cretaceous). Regardless of which is the more robust one, both hypotheses consider the subduction process as a continuum.

We will now briefly review, based on literature, the ages of compression detected across the Apennines, and will attempt to frame the evolution of the Longobucco basin within this more general context.

The age of compressive tectonics has been classically established through the dating of foredeep and wedge-top deposits by means of biostratigraphic analyses. In recent years attempts to date thrusts by means of radiometric dating (U-Pb dating of calcite veins or K/Ar dating of sin-kinematic minerals) have been carried out. The results of these innovative methods are not always satisfactory (see Curzi *et alii*, 2020 and this Thesis for two examples), therefore it is safe to state that paleontological data still represent the most reliable tool.

CENTRAL APENNINES

The seminal work of Cipollari & Cosentino (1995) summarizes the ages of the foredeep/wedge top deposits along a transect crossing the Central Apennines (Fig. 80), highlighting that the innermost sector of the chain (Volsci range) started to deform in the Tortonian, and deformation migrated progressively and continuously toward the east, where the thrusting phase is dated as Zanclean in the Majella area.

SOUTHERN APENNINES AND CPT

In the CPT, a Cretaceous/Eocene event is documented by metamorphism in the Coastal chain (see introduction). Other main compressive lineaments developed at the Oligo/Miocene boundary, generating the deformation of the Calabride Complex and marking the collision of this exotic crustal fragment with the African plate. Such phase is constrained essentially by the age of the late-orogenic deposits of the Stilo-Capo d'Orlando Fm, extensively exposed in the whole SCPT (see introduction), and by the Calcareniti di Floresta Fm, that seals thrust faults in the Peloritani mountains.

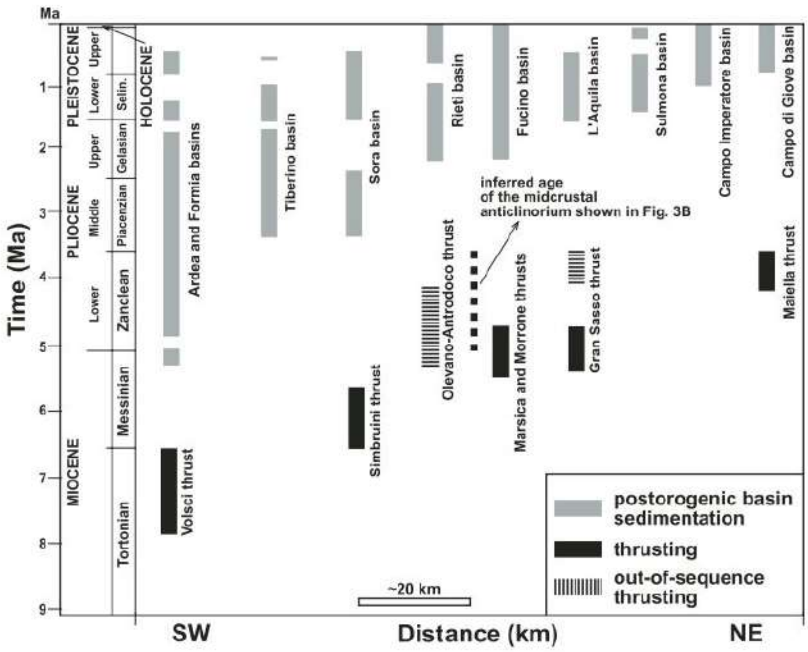

Figure 80: Summary of the ages of thrusting in Central Apennines (Billi & Tiberti, 2009).

The external sector of the CPT is, at a later time, deformed by a post Tortonian phase that progressively migrated toward the Ionian sea (see: Tripodi *et alii,* 2018 for southern Calabria; Polonia *et alii*, 2011 for the Ionian Sea). It must be highlighted that in the CPT a major role in the Neogene tectonic evolution was played by transcurrent faults (see Tripodi *et alii*, 2018).

Moving across the southern Apennines, other interesting and partly contrasting data, with respect to the Central Apennines, can be picked out:

- in the Zannone island, thrusting was dated at 22.01 Ma (Aquitanian) (Curzi *et alii*, 2020);
- in the Lungro-Verbicaro unit (belonging to the so-called Northern Calabrese Units of the Calabro-Lucanian border) compression occurred in the Aquitanian (Iannace *et alii*, 2007);
- in the Pollino Massif and in the Cilento Peninsula the compressive event occurred in the Burdigalian (Sabbatino *et alii*, 2020).

Interestingly, in Southern Apennines a younger cluster of thrusting ages can also be detected, migrating progressively from the Sorrento Peninsula to Mt Alpi (*via* the Aurunci, Matese and Carseolani ranges) since the earliest Tortonian until the Pliocene.

NORTHERN APENNINES

As highlighted in the synthesis of Conti *et alii* (2020), in Northern Apennines three distinct compressive events can be detected (Fig. 81). The older, named by the authors “Ligurian phase” (upper Cretaceous – Bartonian), represents the closure of the Ligurian – Piedmont Ocean, due to a southward directed subduction.

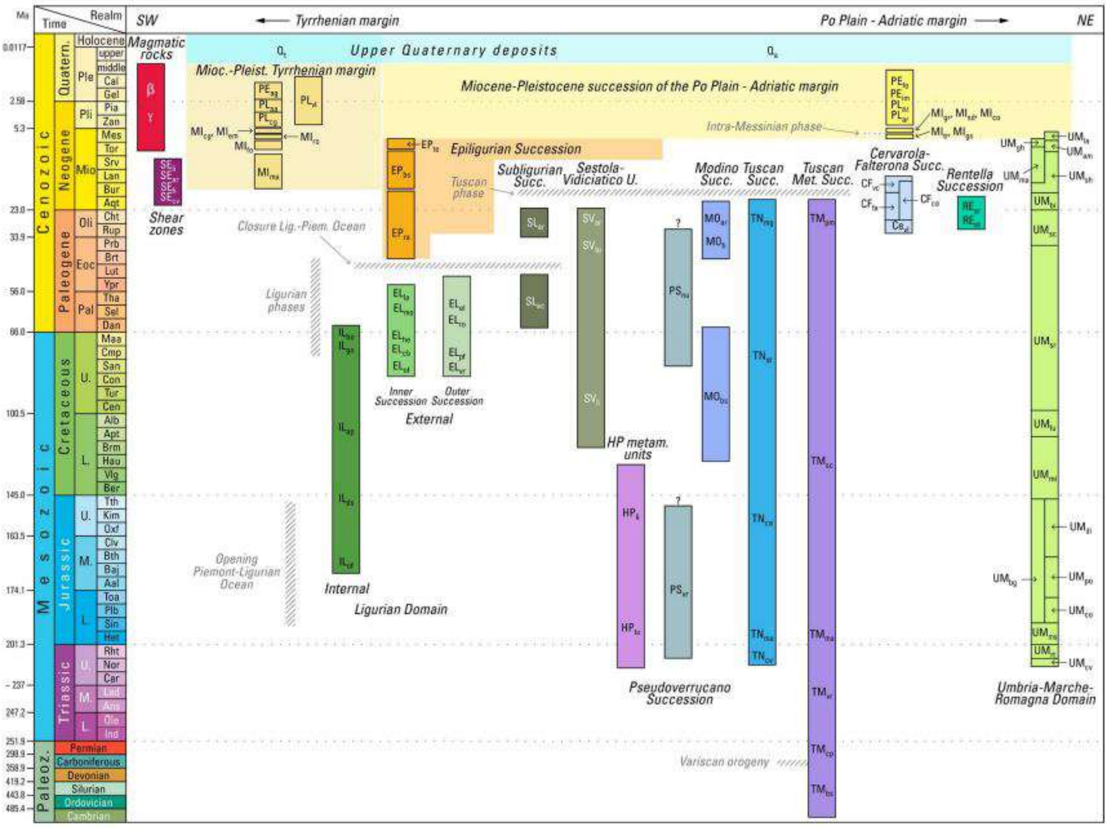

Figure 81: Main tectonic events in Northern Apennines (Conti et alii*, 2020).*

The younger phase, named "Tuscan phase", corresponds to a switch in the polarity of the subduction and the development of a West-directed slab. This phase deforms the late/post orogenic deposits (upper Oligocene – Aquitanian) of the Ligurian accretionary prism (Monte Modino Sandstones), and the foredeep deposits (Macigno Fm) of the newly formed subduction system. In this phase (Aquitanian) a HP/LT metamorphic (Alpi Apuane) event accompanies compression. This episode marks the emplacement of the Ligurian and Subligurian Units above the Tuscan Units. Starting in the Serravallian, a further compressive event, migrating progressively toward the Adriatic, deformed the external domains.

Fresh data, based on K/Ar geochronology, published by Viola *et alii* (2022) on the Zuccale fault (Elba Island), reveal that this lineament experienced compression in the Aquitanian.

Considerations – What does it all mean?

Based on this brief compilation of data, we can detect the existence of three compressive events:

- Cretaceous-Eocene → detected both in the CPT and Northern Apennines;
- Oligocene – Burdigalian → detected in Southern Apennines, CPT and Northern Apennines;
- Tortonian – Recent → detected in the above sectors and in Central Apennines.

Due to the number of open questions regarding these deformative phases (*i.e.* structure vergence, subduction polarity, interaction among subductions ecc.), in this preliminary discussion we propose to separate these phases by using as the sole reference line – a sort of "conceptual/structural divide" - the rotation (inception and halt) of the Sardinia-Corsica block, as this process is firmly time-constrained and universally accepted (Gattacceca *et alii,* 2007). By doing so, we are now able to clearly discriminate among distinct events, and consequently different orogens. While this may sound trivial at first, as it is well known that Alps and Apennines, for example, have different crustal "engines" as their generators so they are undoubtedly distinct orogens, during this Doctorate it became increasingly evident that the "geographic Apennines" embed a major first-order structural theme that, as it will be seen, has been largely underestimated or simply went unnoticed. We will use below the nomenclature proposed by Finetti *et alii*, 2005 and the by Sicilian Authors to describe the evolution of the Peloritani area (Lentini *et alii*, 2006) during the CROP project, to avoid the proliferation of new terms with the same meaning.

- The term "Alpine" refers to the events predating the Oligocene (beginning of the rifting phase of the Liguro-Provencal basin), and related to the convergence of the African and European plates;
- The term "Balearic phase" is herein used to define the events occurring during / caused by the rotation of the Sardinia-Corsica block (late Oligocene – Langhian);
- The term "Tyrrhenian phase" is used to define the events that postdate the halt of the Sardinia-Corsica rotation and follow the opening of the Tyrrhenian basin (Tortonian onwards);

In this light, we can detect the following:

In the innermost sector of the Apennines, in the CPT and in northern Tuscany, the coexistence of all three phases;

In Southern Apennines the "Balearic" and "Tyrrhenian" phases;

In Central Apennines the "Tyrrhenian" phase.

In the past (Rossetti *et alii*, 2001; Vignaroli *et alii*, 2012; Vitale *et alii*, 2019; Sabbatino *et alii*, 2021) most authors have considered such phases as a continuum, and consequently the product of the same geodynamic process, therefore representing one orogen. We want to stress here – this is the core of our approach - that these phases rather represent discrete processes, being the products of distinct geodynamic engines, so we are in fact dealing with two separate (but welded) orogens (see also Barchi *et alii*, 2012).

One first-order evidence is the presence of two widespread gaps between such events. The older gap spans the late Eocene to Oligocene (10 ma) while the second gap spans part of the

Burdigalian and part of the Serravallian (6 Ma). This latter gap is also beautifully highlighted by the isochrons of the foredeep deposits of Southern Apennines (Sabbatino *et alii*, 2021) (Fig. 82).

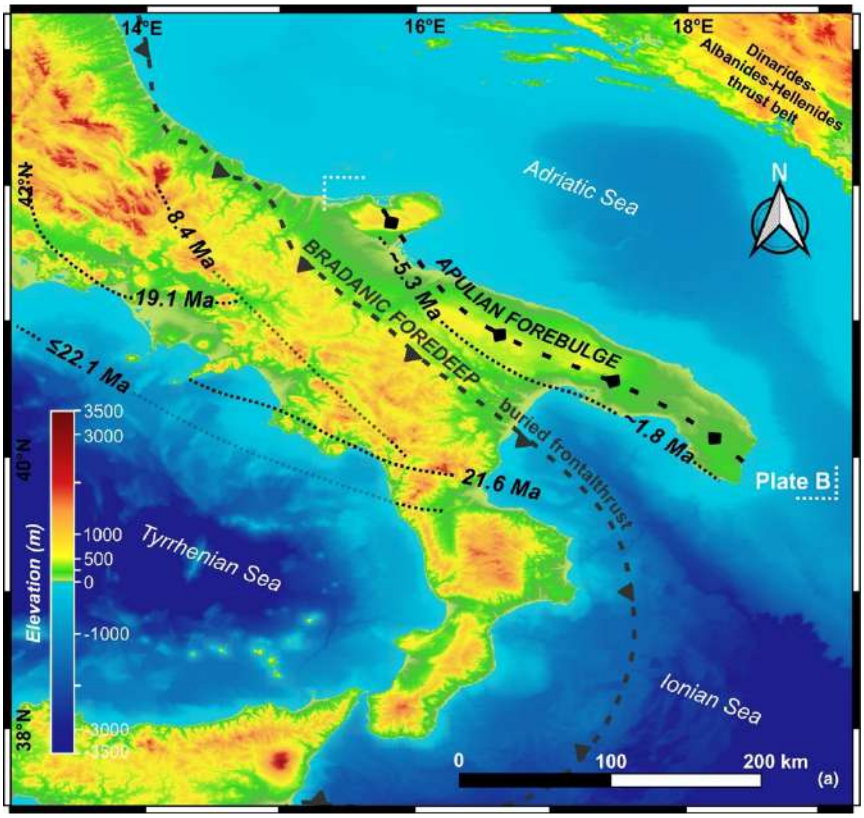

Figure 82: Isochrones of the Apenninic foredeep (From Sabbatino et alii*, 2021).*

The picture taken from the paper of Sabbatino *et alii* (2021) is also interesting as it clearly shows that the 8.4 Ma isochron cross-cuts the older isochrons and is obliquely oriented with respect to the 19.1 Ma isochron. This is an indication of the presence of two distinct geological objects as opposed to one.

We will now focus our attention on the "Balearic" phase. This is the most underinvestigated phase as it is often confused, as we mentioned, with the "Tyrrhenian" phase having the abovementioned gap between Burdigalian and Tortonian not been adequately taken into account by previous authors.

Alvarez *et alii* (1974) had already assumed the presence of this phase on a purely theoretical basis, invoking that the Sardinia-Corsica block rotation must have produced a first compressive phase, with Apenninic polarity. This hypothesis has been later (see Speranza *et alii*, 2000 for an example) re-proposed sporadically in the literature. Later, Sicilian authors (see Lentini & Carbone, 2014 for an extensive review) recognised in eastern Sicily a distinct compressive phase in the Calabride Domain that they named "Balearic" as it was simultaneous with the opening of the Balearic Basin. In the frame of the CROP project this phase was also identified by Finetti (2005). Authors relate this phase to the onset of a westward directed subduction of the Alpine Tethys beneath Corsica, which progressively started to retreat as the Liguro-Provencal Basin (Balearic Basin) started to open. The halt of this phase (~15 Ma) is related to the complete consumption of the Tethyan oceanic crust through the collision of the Sardinia-Corsica-Calabria element with Africa.

Two are the existing models (Fig. 83):

- the first (Finetti, 2005) considers Africa and Adria as separated, and the Ionian Ocean and the Alpine Tethys being connected. The collision of Sardinia-Corsica-Calabria with the so called Etrurian Corner and Partenopean Corner allowed the subduction of the Ionian slab, whose retreat guided the following "Tyrrhenian" phase;
- the second (Rosembaum *et alii*, 2004) considered the two Oceans (Ionian and Alpine) as separated, being the Africa plate connected to Adria.

As stated by Finetti (2005), the "Balearic" orogen during the following "Tyrrhenian" phase, related to the retreat of the Ionian slab and to the opening of the Tyrrhenian basin, was dissected and passively transported at the rear of the Apenninic wedge. In this stage, the CPT was affected essentially by transcurrent tectonics.

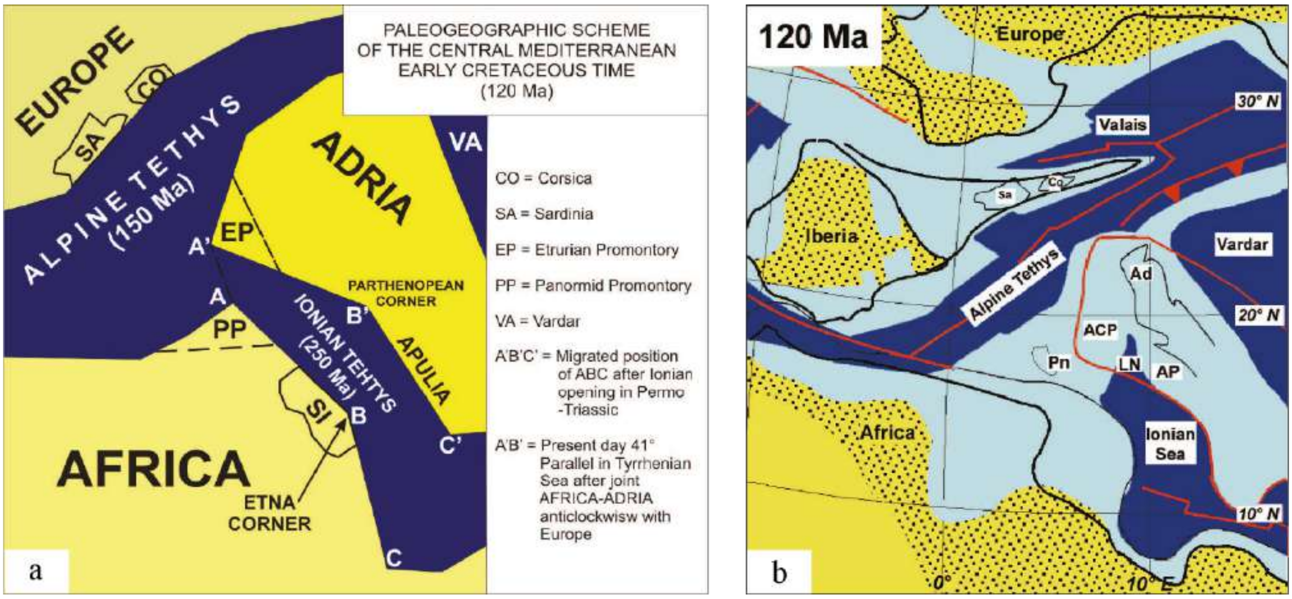

Figure 83: a) paleogeographic reconstruction representing a scenario where the Alpine Tethys and the Ionian Tethys are connected (after Finetti, 2005); b) Alpine Tethys and Ionian Sea are separated by the Adria promontory (Rosenbaum et alii*, 2004).*

THE PALUDI BASIN

The Paludi Fm is an Ypresian/Lutethian unit, consequently, following the terminology proposed herein and due to its markedly clastic nature, it must be considered as the product of the "Alpine" phase. In the literature it has been interpreted as a foredeep deposit, but as discussed in the relevant chapter, the sedimentological features of the deposits are not those of a typical turbiditic basin.

We highlight the following:

- the presence of a continuous megaclastic wedge-shaped belt, resting unconformably on the basement and pinching-out toward the North (observable in the lowermost tectonic unit);
- In the higher tectonic units the arenites rest unconformably on different Mesozoic units;
- The megaclastics were sourced by repeated catastrophic collapses (rock-fall/avalanching, grain flows and debris flows) that were predominant in an initial stage of the basin development;
- No evidence of sourcing from the rising Alpine chain, as the sourcing is completely local;
- Progressively such deposits were replaced by sand-rich bodies, the products of sandy dense flows and not of true turbidity currents.

* The basal contacts indicate that the substrate underwent uplift and erosion until the complete exhumation of the Paleozoic basement in the lowermost tectonic unit;
* The triggering mechanisms for the emplacement of the megaclastic deposits were conceivably repeated earthquakes;
* The sandy dense flows started to prevail when tectonic activity started to decrease or vanished.

We might explain the exhumation of the pre-Paludi substrate by inferring the growth of anticlines, but this hypothesis does not fit our field data. In fact, the presence of the megaclastic wedge is restricted to the lowermost tectonic unit, while both in the backlimb and forelimb of the mapped anticline such deposits are missing. In the higher tectonic units only the marly and sandy deposits pertaining to the Paludi Fm can be detected, with the exception of the Section 9 and along the Colognati River Valley, where the distal part of the gravelly dense flows has been mapped onlapping a preserved Jurassic basin-margin escarpment (Santantonio *et alii*, 2016). The above indicates that the basin morphology, inherited from the Early Jurassic rift, had undergone an uplift but was essentially undisturbed and there was only one principal uplifted area that sourced the system. This area was not the Alpine chain, as a provenance from high grade metamorphic rocks or ophiolites typical of the Calabrian Alpine chain has never been detected in the Paludi Fm (see also Zuffa & De Rosa, 1978).

We therefore interpret the Eocene Paludi basin as a part of an Alpine foredeep, in particular in a forebulge setting (but see below). Geological mapping clearly indicates that the megaclastic wedge abuts the basement *via* a steep onlap unconformity corresponding to a submarine escarpment rooted in a synsedimentary normal fault which bordered the bulge crest. The footwall blocks experienced major uplift, resulting in erosion of the Jurassic succession and eventually unroofing of the basement. Similar examples have been detected in central Apennines (Fabbi, 2016; Tavani *et alii*, 2015b).

Also in the other tectonic units the basal contact of the Paludi Fm is marked by an unconformity. The extent of the hiatus varies strongly, from locality to locality. In the literature such unconformity in similar settings is named “forebulge unconformity” and is essentially related to the flexuring of the lithosphere.

Although in the literature models also envisage complex erosional contacts with unconformities, these are generally low-angle unconformities, consistent with a broad arching (White *et alii*, 2002) (Fig. 84). In our case, in contrast, we were able to map several instances of high-angle

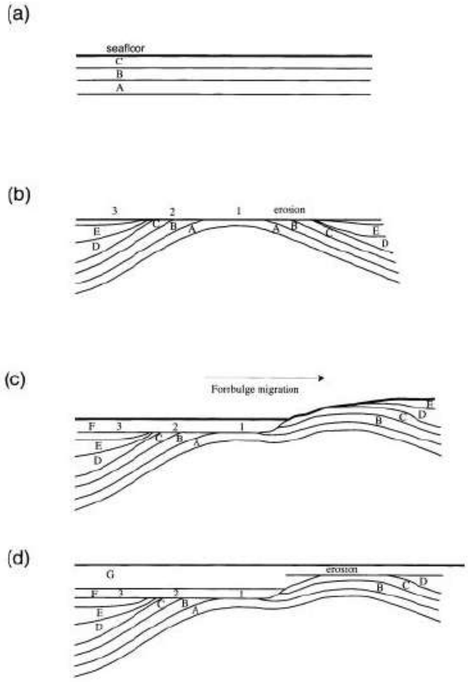

Figure 84: Model showing the development of complex basal forebulge unconformity related to the action of uplift paired with submarine erosion in the crest area of the bulge, with the progressive migration of the bulge and the coeval sedimentation (from White et alii*, 2002).*

contacts in different localities, which suggests that normal faulting accompanying the bulging process was most effective in Sila, possibly also due to the presence of a network of pre-existing Jurassic rift faults underlying the Paludi Basin. At least in one case (Caloveto area), an Eocene normal fault clearly re-utilizes a late Sinemurian/early Pliensbachian rift fault zone. In conclusion, and in the light of the local unroofing of the Hercynian basement, erosion linked with deposition of the Paludi Fm greatly benefited from the growth of syn-sedimentary normal faults. While lower angle unconformities with the sedimentary substrate can be interpreted as unconformities related to broad arching (forebulge unconformities), the high angle contacts with various Jurassic units or with the basement are best interpreted as the products of local synsedimentary extension.

The widespread absence of post Hauterivian to lower Eocene deposits in Sila, with a stratigraphic gap of >80 Ma, was noted by several authors (Magri *et alii*, 1965; Dubois, 1970). It is important to note that sparse bits of the missing succession cited above, *i.e.,* covering the time span of the hiatus, can locally be found reworked in the Paludi Fm as plastically deformed elements.

Another important piece of evidence, that led us to invoke the presence of normal faulting, can be found in the Colognati Valley. Here a Jurassic paleomargin directed ~NE-SW, as mentioned above, was identified by Santantonio *et alii* (2016).

This inherited margin is onlapped by the Paludi Fm through a complex erosional surface running parallel to the margin, sealing the latest Jurassic onlap against the Carboniferous granite. This disagrees with the expected dynamics and geometries of migration and erosion of a bulge crest, which is generally orthogonal with respect to the direction of orogenic transport. In the case of the Paludi Basin, the orogenic transport was roughly towards the northeastern quadrants, and was therefore subparallel to the local basin-margin onlap along the Colognati Valley, while the bulge must have had a circa NW-SE strike. In this particular case, uplift of an element (exhumed Jurassic basin margin) striking orthogonally with respect to the strike of the migrating bulge suggests the presence of a normal fault orthogonal with respect to the master faults bordering the bulge.

Such transversal faults, oriented perpendicularly with respect to the trend of the peripheral bulge, are reported in the literature from these settings (Tavani *et alii*, 2015a) (Fig. 85).

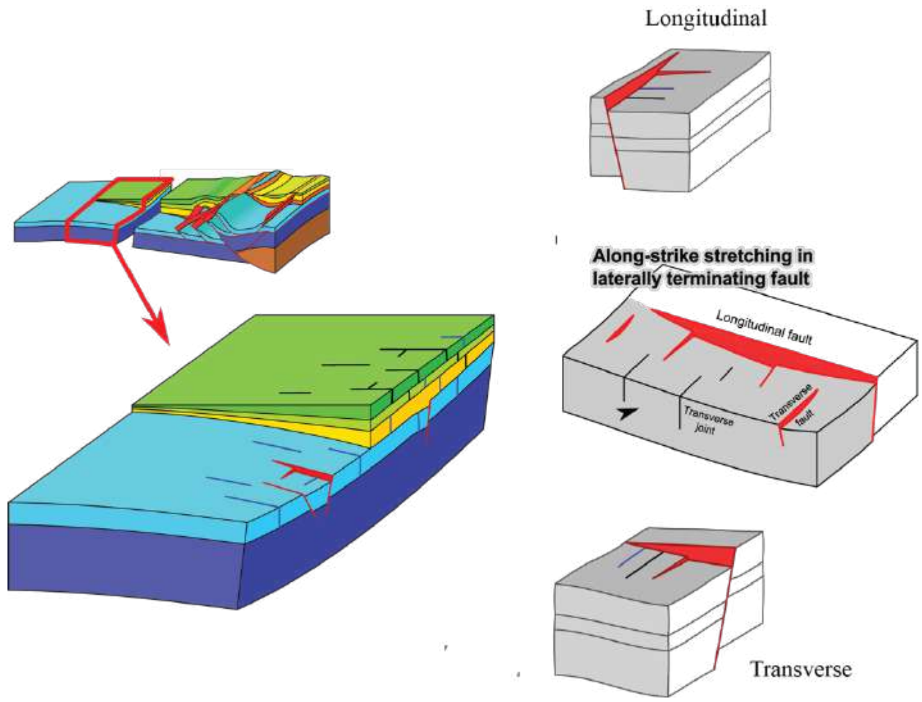

Figure 85: Typologies of extensional faults developing in the bulge area (Tavani et alii*, 2015a).*

We speculate that the early Jurassic fault/s that bordered the Colognati Valley was reactivated during the migration of the bulge system, being an inherited weakness zone of the lithosphere. Given the geometric relationships between the Paludi Fm and the Jurassic paleomargin, this Cenozoic fault would have acted as a transfer fault. As such, a steeper dip than that the Jurassic rift fault would be expected. This would configure a situation where the new fault runs parallel to the strike of the Jurassic fault, but being steeper it preserves the termination of the Mesozoic onlap wedge against the granite (*i.e.*, it runs "offshore" with respect to the mappable Jurassic onlap line). Unfortunately, some crucial information cannot be collected as only a small onlap wedge of the Paludi Fm, overlying the exhumed Jurassic onlap wedge mentioned above, is preserved. A thrust plane, juxtaposing the Hercynian granites above the Meso/Cenozoic succession, buries the basinward continuation of the basin-margin beds.

A nice analogue comes from the Briançonnais domain (Michard & Martinotti, 2002). Here an unconformity between the Lower Paleocene and Priabonian deposits was mapped, and normal faults detected in the field. Such elements have been interpreted as related to the peripheral bulge active at the front of the Alpine chain. There exist similarities between the Longobucco Basin and the Briançonnais Domain also in the Jurassic. These two areas, along with Sardinia and Corsica are paleogeographically similar as they represent a belt of stable basement-made topographic highs during the Triassic and Hettangian, while other sectors of the Tethys were strongly subsiding (Santantonio & Carminati, 2011; Passeri *et alii*, 2014; Costamagna, 2016; Santantonio *et alii*, 2016). These similarities would lead us to place the Longobucco Basin in the European plate, as the Briançonnais domain. Another important consideration concerns the tectonic setting of the whole Calabrian Arc. The Paludi basin helps also in shedding light on the subduction polarity in this part of the Chain at the time of the Alpine Orogeny. In the models dealing with the development of the bulge in a westward directed slab, the forebulge is expected to migrate toward the foreland (toward the east), along the flexuring subsiding plate (see DeCelles & Gilles, 1996) (Fig. 86).

Consequently, it is impossible to place the Longobucco Basin in the upper European Plate with a westward directed subduction as proposed by some authors as a bulge would not form in that place, in this hypothesis. Furthermore, we know that the Calabride Complex overlies the Liguride Complex, consequently, considering a westward directed long-lasting subduction, the Longobucco Basin cannot be placed on the Africa plate, being the sense of tectonic transport toward the East, thus resulting in the juxtaposition of the oceanic domain on the continental crust. For the same reason, in the context of the internal organization of the Calabride Complex, the Longobucco occupies the highest position in that Complex, consequently it must have been located, in Alpine times, westward with respect to the other units pertaining to the Calabride Complex.

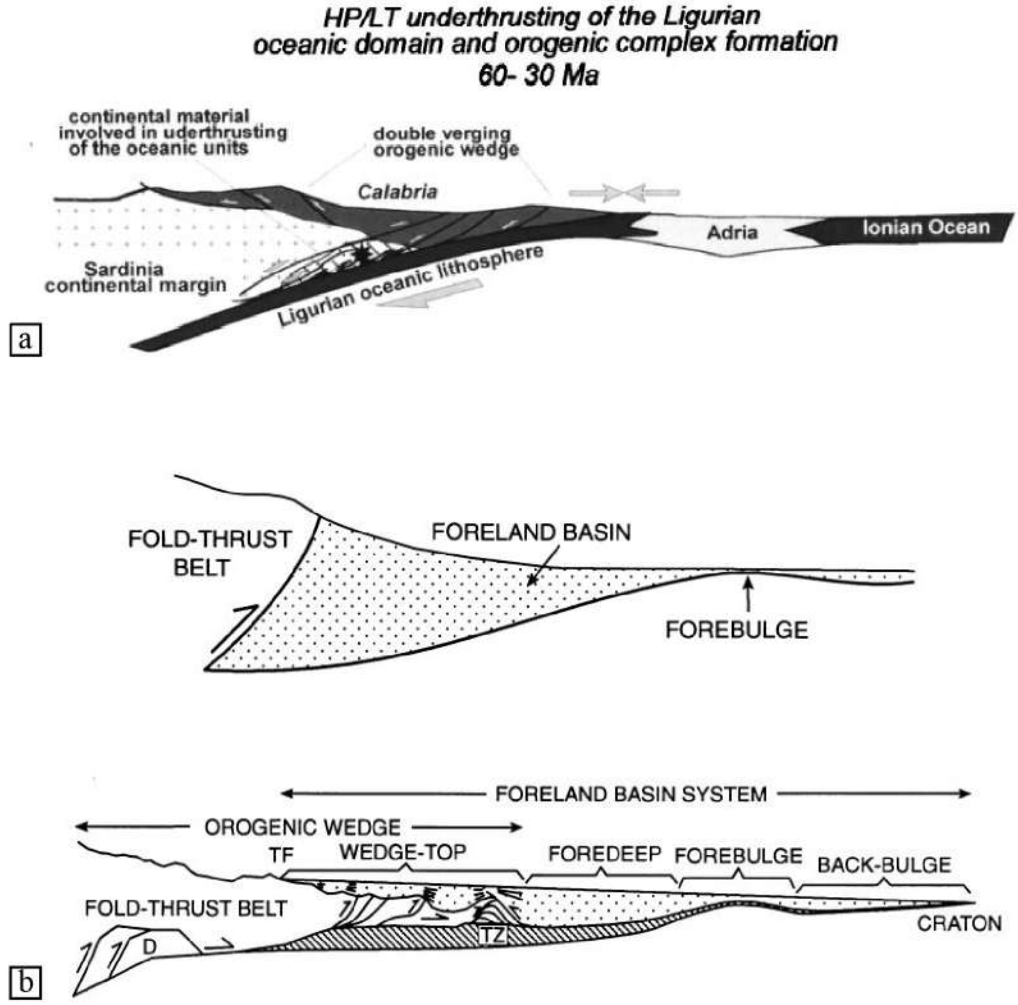

Figure 86: a) The Longobucco basin is placed in the overriding European plate in a westward subduction, where however a bulge system would not develop. The bulge area is indeed expected toward the foreland (Adria) (Rossetti et alii*, 2001); b) Scheme representing the depozone of a foredeep (DeCelles & Gilles, 1996), in a westward directed subduction the bulge area is located toward the East.*

We therefore propose that the Longobucco Basin in the Alpine phase was part of the European plate being located in the periphery of this orogen as testified by the presence of the peripheral bulge system.

As mentioned repeatedly in this Thesis, the Longobucco/Paludi basin “has more gaps than stratigraphic record” (both in a temporal and geometrical sense), being a kind of snapshot of a discrete moment of the geological history of this sector of the Tethys. Concerning the architecture of the bulge basin, we did detect the existence of a main fault directed toward the South. We do not know if other faults, with opposite dip, existed in the Eocene as the basin abruptly ends, being covered by the Late Miocene deposits. We know that the alpine bulge should migrate toward the north and major extensional faults bordering the forebulge area should have been northward dipping. This is in contrast with our data, so we can propose one twofold alternative:

- The basin was located in the back-bulge and not in the forebulge;

- As a consequence of the above point, there must be other, antithetic faults dipping toward the North that are now eroded/buried;

From the Apennines we learn that thrusts commonly deform their own foredeep. For example, in the Simbruini Range of Central Italy the foredeep deposits (lower Tortonian/lower Messinian) are deformed by late Messinian thrusts, this shortening involving also a Tortonian bulge (Fabbi, 2016). The time span separating the formation of the foredeep and its deformation is thus very short, as expected with known computed thrust migration rates in this region of 15 mm/yr (Sabbatino *et alii*, 2021).

In the case of the Paludi Basin, thanks to basin modelling, we demonstrate that the basin experienced compression around the Oligo/Miocene boundary. The time span intercurred between the deposition of the Paludi Fm and the thrusting phase would be of 24 Ma, that is an unreasonably long time span if compared with the abovementioned timings. Consequently, we cannot ascribe the compressive phase to the same process that guided the formation of the foredeep and the migration of the bulge. We refer the compressive phase described in the Longobucco basin to a Balearic phase, thus related to the westward subduction of the remaining part of the Ligurian-Piedmont Ocean below the European plate. Such phase was characterised by a progressive migration of the thrust front toward the eastern quadrants. We constrained the time between the burial related to compression and the exhumation by means of AFT analysis related to the eastward propagation of such thrusts. In the Longobucco area, an exhumation by means of a low angle extensional detachment, recognised in other sector of the region, can be confidently excluded being the calculated exhumation rate slow and thus referrable to denudation only. This Balearic orogen was then passively transported at the rear of the “Tyrrhenian” orogenic wedge that is today active in the Ionian Sea.

A similar scenario can be found in the Tuscany area, where the products of the “Alpine”, “Balearic” and “Tyrrhenian” phases coexist. In other areas of the Apennines, only the last two phases can be detected (Southern Apennines, Southern Calabria, Sicily), while in central Apennines the evidence of the “Tyrrhenian” phase only is detected.

The Longobucco/Paludi basin evolution suggests that the subduction history of what we now call the “Apennines” was not an almost continuous and steady process, but rather the product of episodic pulses referrable to distinct and discontinuous events, eventually leading to the welding of different orogens. The noticeable gap separating a population of thrusts older than 20 Ma from another population of thrusts younger than 15-10 Ma reflects the halt of rotation of Sardinia-Corsica, and jump of the zone of back-arc extension from West to East of said Block. The U/Pb age, (8 Ma) detected from calcite veins, are in this light interpreted as the product of the Tyrrhenian phase, that produced a phase of major uplift.

Establishing an early Eocene age for the Paludi Fm has a paramount impact on our understanding of the tectonic evolution of the CPT. Being well aware of Bonardi *et alii*'s (2005) paper, we were genuinely expecting that our new sampling campaign would have confirmed it. As clear indications for an age not younger than the Eocene started to emerge, we sampled and re-sampled every possible outcrop, and cross-checked with different fossil groups, only to eventually reinforce the result that any evidence for an Oligocene or early Miocene age was positively missing. Any reworked forms, mainly planktonic forminifera where present, were all of pre-Eocene (e.g., Cretaceous) age. As it was thoroughly discussed in the "Age assessment" section, the LBF assemblages were subjected to a strict scrutiny exactly with the goal of detecting even the slightest piece of evidence for an age younger than the Eocene, but with no success: the assemblages are all age-consistent, within the (narrow) limits of biostratigraphic resolution. We must therefore conclude that, if we wished to rule any paleontological error out, the allegedly Aquitanian samples of Bonardi *et alii* (2005) must have been collected in some outcrop documenting a post-Eocene, pre-Serravallian basin that is otherwise undocumented in our study area, although this latter virtually comprises the complete areal extent of the Paludi Fm.

As an Eocene age for the Paludi Fm. was eventually being established, in agreement with virtually all the existing literature predating the paper of Bonardi *et alii* (2005), and field evidence firmly constrained the age of thrusts, many of the working hypotheses that had been initially considered when this Doctorate started evaporated, and had to be rejected by this writer. In return, a new picture started to materialise. One remarkable consequence of our novel interpretation is that while our thrusts, delimiting the four tectonic units, are "Balearic" – that is they have a late Oligocene/earliest Miocene age – any Balearic foredeep is unpreserved. This marks a major difference with the southern sectors of the CPT, where the about 600 thick Stilo-Capo d'Orlando unit represents exactly this paleogeographic element which is missing in northern Calabria, as discussed above. As we mentioned, in our study area the Balearic thrusts, linked with the emplacement of the Calabride Complex onto African/Adriatic crust, deform an alpine peripheral bulge. It has been repeatedly stressed in this work, that discrete phases of uplift and erosion have completely erased some "geologically necessary" parts of the Mesozoic depositional systems: for example the Toarcian-Middle Jurassic mixed carbonate-siliciclastic shelf that had to feed the turbidites found in the Trionto, Calcari Grigi con selce and Radiolarian cherts formations of the Longobucco Group, or the parent shallow-water factories which sourced the LBF-rich turbidites found in the Paludi Basin

Also in the light of quantitative basin analysis data (thermal maturity, burial history), we know that a no less than 4 km thick crustal section must have been eroded, which had to overlie the

Longobucco Basin in pre-Serravallian times. We speculate that this missing burden can have included the foredeep infill of the Balearic orogen.

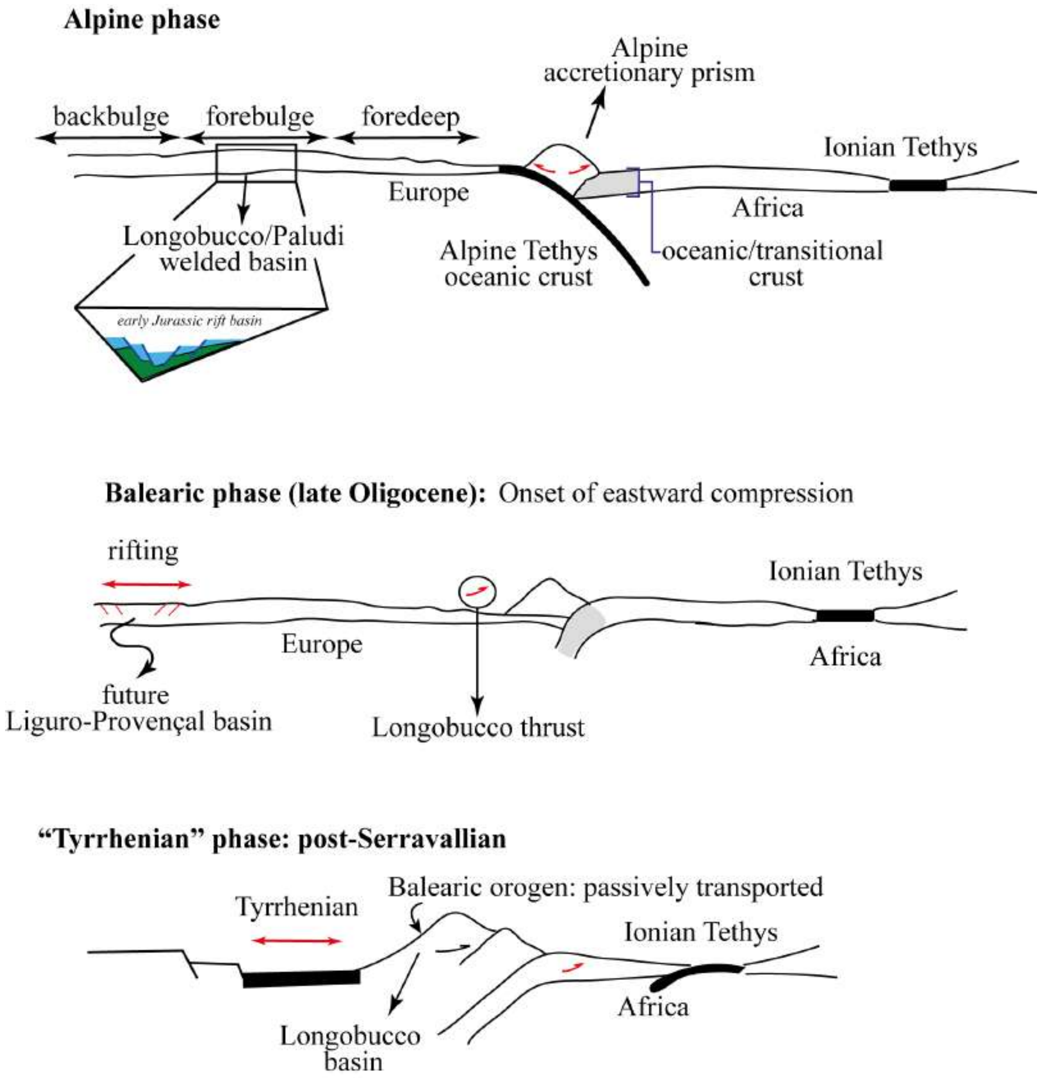

Figure 87: Main deformation phases affecting the Longobucco Basin.

Conclusions

In conclusion our results can be summarised as follows:

- The Paludi Fm has been dated as Ypresian/Lutetian (between the SBZ 11-12);
- The unit is characterised by three lithofacies displaying high lateral variability;
- The unit is herein reinterpreted as a very proximal product of the basal part of bipartite turbidity currents (*i.e.*, dense flows, debris flows, grain flows, rock-falls and avalanching) only eventually displaying true turbiditic deposits;
- The triggering mechanisms were essentially related to seismic shocks, related to the presence of extensional fault(s) bordering the basing toward the North (present-day coordinates) at least in the basal part of the succession; successively the mixed carbonate-siliciclastic shelf that had to back the basin dominantly fed with sand-sized resedimented material the sandy dense flows;
- The Paludi Fm is thoroughly characterised by Mass Transport Deposits and can be therefore be considered as a Mass Transport Complex;
- The Paludi basin is herein interpreted as a forebulge/backbulge setting of the Alpine foredeep resting on the European plate, close to the Briançonnais Domain and Sardinia;
- The Lower Cretaceous/Eocene hiatus can be related to flexuring of the lithosphere paired with the action of extensional faults;
- Jurassic extensional (rift) faults where partly reactivated in the bulge phase;
- The age of thrusting falls at the Oligo/Miocene boundary;
- The basin suffered a tectonic load of 4 km; this stack is today totally eroded due to a phase of strong uplift that erased most of the Post-Paludi record (second stratigraphic gap);
- While this sector of the basin was being vigorously uplifted, the southern sector of the CPT was instead subsiding (as testified by the Stilo Capo d'Orlando Fm);
- The compressive phase can be related to the "Balearic" phase referred to a switch in the polarity of the subduction, to the Sardinia-Corsica rotation and to the opening of the Liguro-Provençal basin;
- The Balearic orogen was, in the Tyrrhenian phase, passively transported at the rear of the Apenninic orogen.

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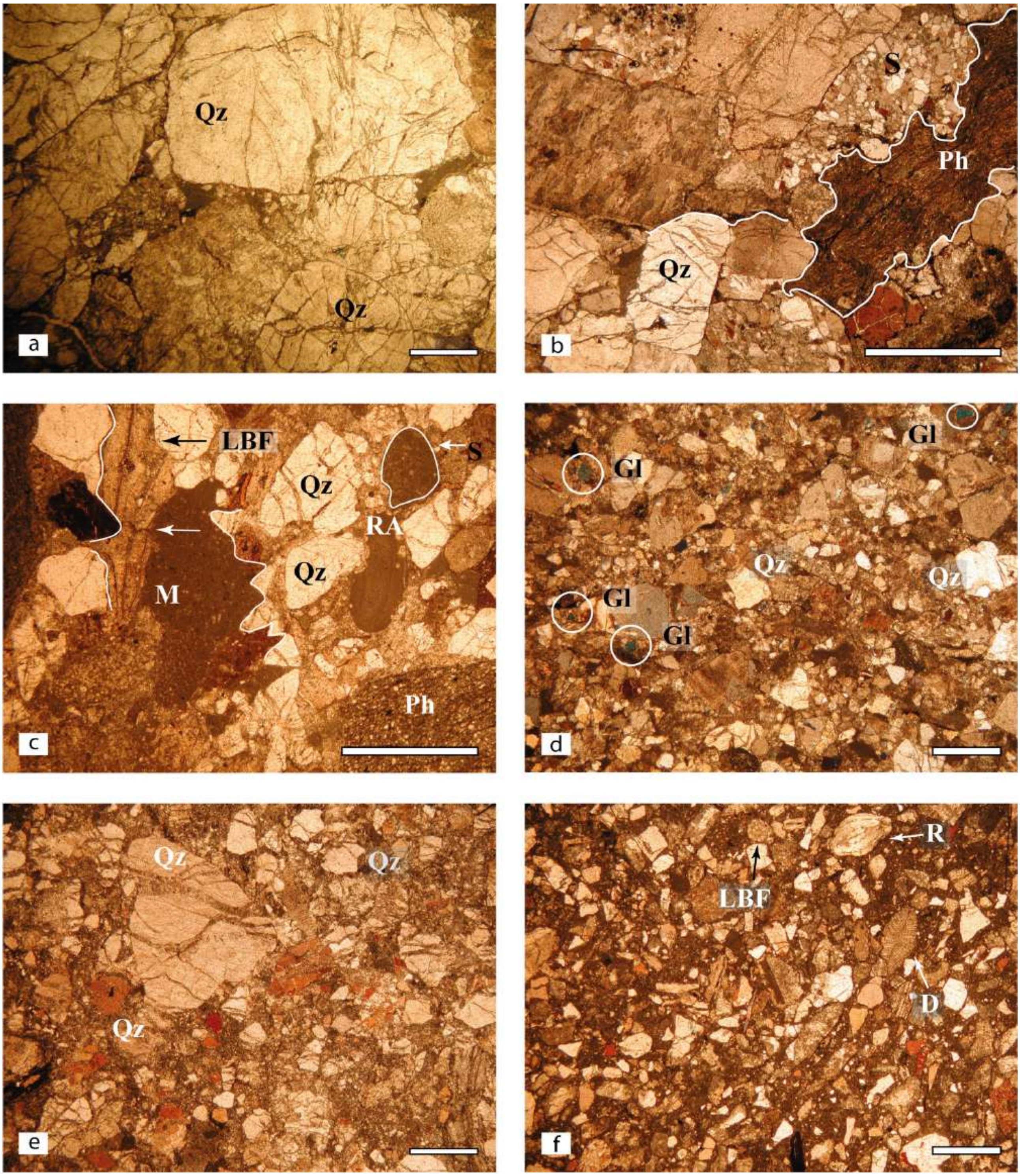
Plate I

a, b – Thin section of the clast-supported breccias where fine matrix is totally absent. The ground mass is represented by clasts with size > 1 mm. Sutured contacts among clasts are common, evidenced by the solid white line. Quartz (Qz), sandstone (S), phyllite (Ph) clasts can be observed. Scale bar 1 mm.

c – Thin section of the lithofacies 1 with a low percentage of silty matrix; lithoclasts (S) are present along with phyllites (Ph), subrounded quartz (Qz), and marly chips (M). Marls are deeply compenetrated by Qz grains with sutured contacts (white line). Top left a damaged LBF (*Discocyclina* sp.), with broken margins (white arrow), can be observed, compenetrated by Qz clast. A red algae fragment (RA) is also present. Scale bar 1 mm.

d, e – Clast-supported arenite/microbreccia with moderate amount of silty matrix. Contacts among clasts are uncommon. Glauconite (Gl) is circled in white. [Qz: quartz, subsrounded (larger clasts) to angular)] Scale bar 1 mm.

f – Thin section of a matrix-supported arenite/microbreccia. Macroforaminifera [both *Discocyclina* sp. (D) and Rotaliidae (R)] are present. Scale bar 1 mm.



Qz
Qz
a
S
Ph
Qz
b
LBF
S
Qz
RA
Qz
M
Ph
c
Gl
Gl
Qz
Qz
Gl
Gl
d
Qz
Qz
Qz
e
R
LBF
D
f

Plate II

a,d – General appearance of the Lithofacies 3, characterized by densely-packed, matrix poor, coarse hybrid arenites. LBF are deeply damaged and often compenetrated by subrounded Qz grains (as marked by the solid black line in a). The marginal part of the test (arrowed) is missing and the external whorls are abraded. Darker elements are mudstone lithoclasts. Glauconite can also be observed (Gl). Scale bar 1 mm.

b – Matrix-poor sample characterized by densely packed LBF, with subparallel equatorial planes that in several instances compenetrate each other. Glauconite (Gl) is present (circled). Scale bar 1 mm.

c – *Nummulites* sp. and *Discocyclina* sp. Scale bar 1 mm.

e – Lithofacies 3 with plastically deformed clasts of reddish marls (dashed), phyllites (Ph) and quartz (Qz). Scale bar 1 mm.

f – *Globotruncana* sp. found in a reworked clast embedded in the Lithofacies 3. Scale bar 1 mm.

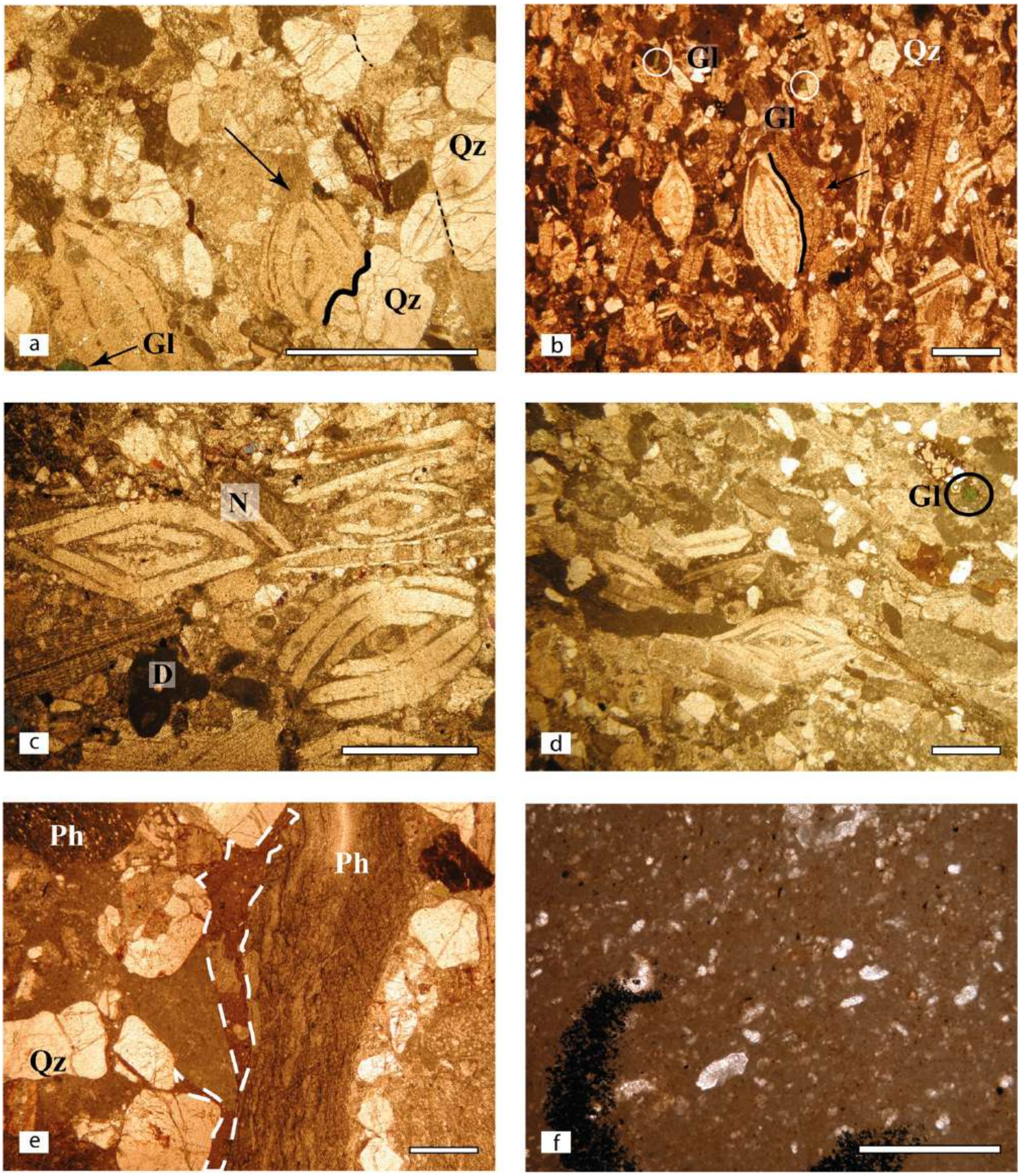

a
Gl
Qz
Qz
b
Gl
Gl
Qz
c
N
D
d
Gl
e
Ph
Ph
Qz
f

Plate III

a – *Globigerinelloides* sp. in the reworked Aptian bed-stack described in Fig. 23. Scale bar 1 mm.

b – *Granorotalia sublobata.* Scale bar 1 mm.

c, d – *Ornatorotalia granum*. Scale bar 1 mm.

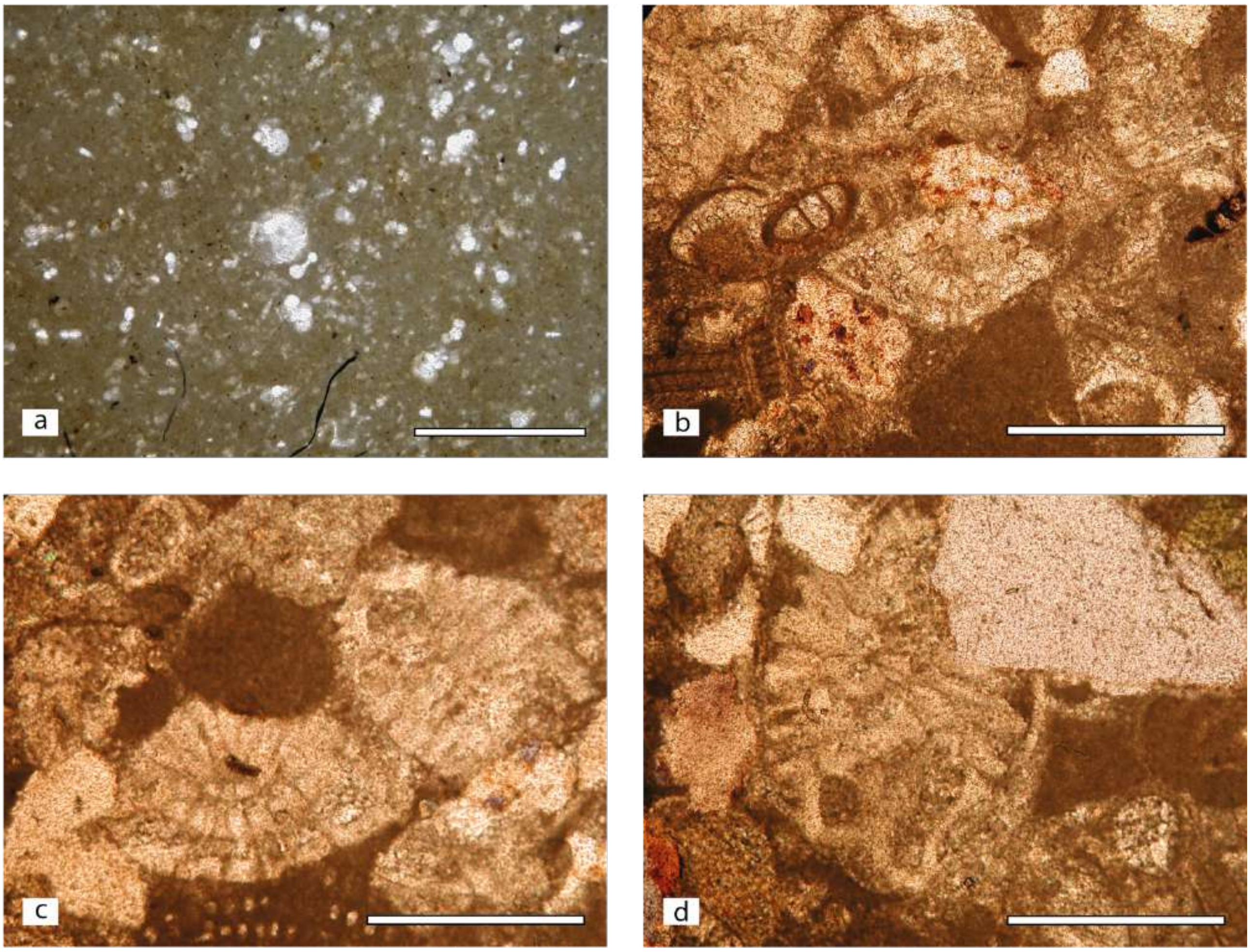
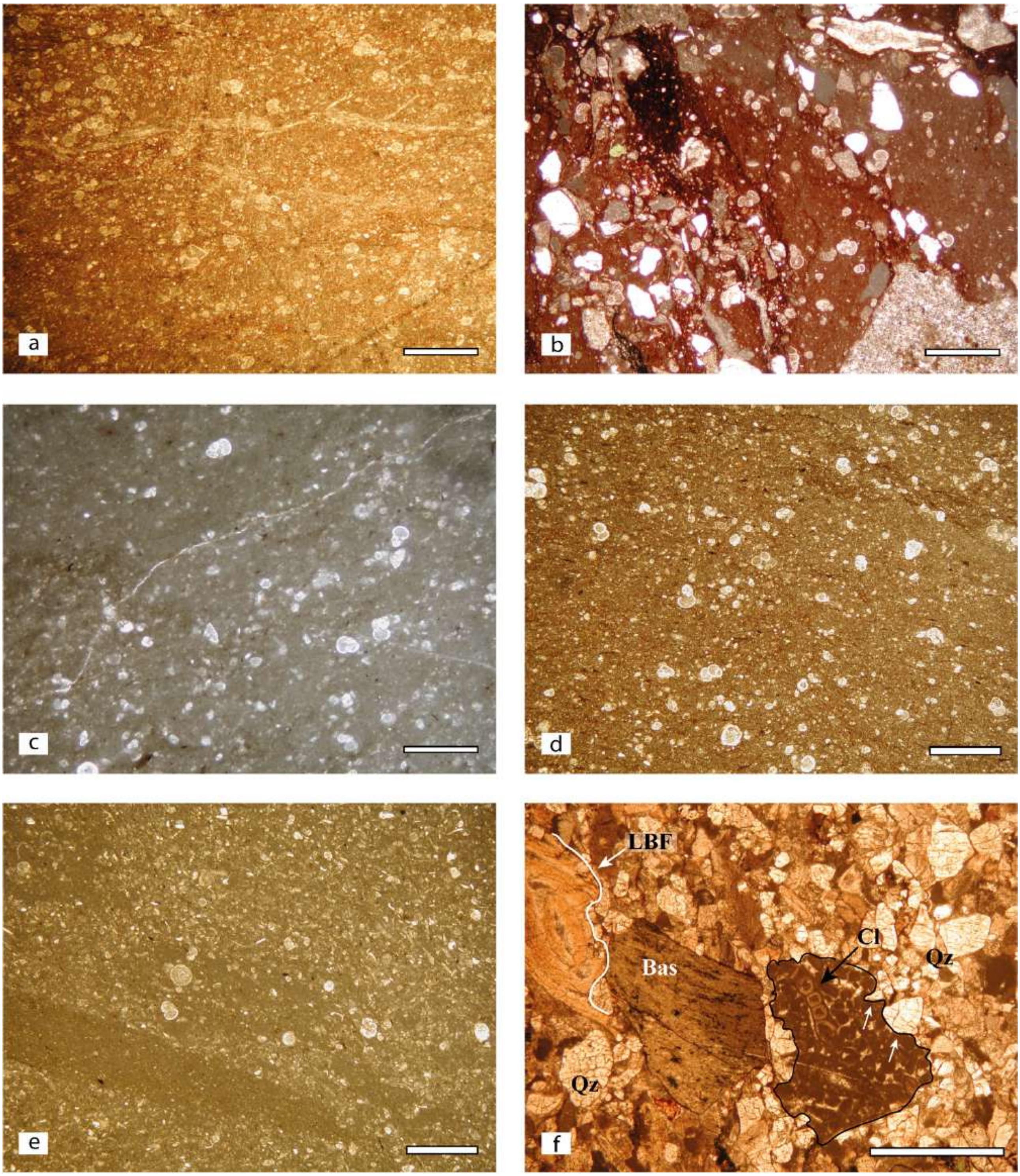

a
b
c
d

Plate IV

a, b, c, d, e – General appearence of the marly interbeds, with interbedded arenites (b), bearing planktonic foraminifera that are often badly preserved. It is possible to observe the genera *Akarinina, Subbotina and Morozovella*. In particular, the co-occurrence of LBF and *Morozovella cf. formosa* point to an Ypresian age for the deposit. Scale bar 1 mm.

f – LBF occur as loose grains within the hybrid arenites of the Lithofacies 3: this can be observed if we compare macroforaminifera with other older fossils like the *Clypeina jurassica* (Cl), that is clearly part of a bigger lithoclast. Quartz: Qz; Basement-derived clast: Bas. White arrows indicate sutured contacts. Scale bar 1 mm.


a
b
c
d
e
f
LBF
Bas
Cl
Qz
Qz