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A geological survey of the Farma River Valley (Monticiano-Roccastrada metamorphic core complex, Tuscan Palaeozoic, Italy) – preliminary results and considerations

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Abstract: A geological survey at the 1:10,000 scale was performed in the Farma River Valley, the central part of the late Palaeogene–Quaternary Monticiano-Roccastrada metamorphic core complex (MRMCC), Tuscany, Italy. The area is subdivided by complex first-order faults into tectono-stratigraphically homogeneous subzones, characterised by distinctive Carboniferous lithofacies and by late Palaeogene–Quaternary deformation and metamorphism. The uniform stratigraphic base is formed by a condensed interstratification of carbonous mudstones, fine-grained greywackes and cherts (late Emsian–early Tournaisian), which were deposited in a shallow epicontinental starved basin. Subsequent extensional fragmentation produced coeval, predominantly siliciclastic depositional areas. The sediments form progradational highstand systems tracts above a downlap surface, which represents the floor of the shelf, the lower slope, the base of the slope and the margin of the basin floor. The mobilisation of the sediments via tempestite-turbidite cycles was probably triggered by a hothouse pulse, related to the Hangenberg Crisis. The coeval units created were covered uniformly by regressive, littoral-deltaic and continental siliciclastics (Permian to middle Triassic) and subsequently formed the substrate of the Mesozoic, transgressive, passive margin deposits of the Adriatic microplate. Oligocene to Miocene thrusting- and subduction-related tectonic burial to a depth of ca. 25 km, followed by Miocene to Pliocene extensional exhumation, was controlled by several shear zones, including the reactivated Carboniferous normal faults mentioned above. This resulted in the formation of closely spaced interfering metamorphic core complexes. Post-Messinian oroclinal bending caused the intense fragmentation of the subzones. The Farma River Valley is interpreted as the geomorphological expression of exhumed tear faults, which were part of the fracture system controlling the late Tortonian–Quaternary eastward-migrating Tuscan Magmatic Province.

Kurzfassung: Vom Farmatal, dem zentralen Teil des jungpaläogenen-neogenen Monticiano-Roccastrada metamorphen Kernkomplexes, ist eine geologische Karte im Maßstab 1:10.000 erstellt worden. Das Gebiet wird von komplex gebauten Hauptstörungen in tektono-stratigraphisch homogene Subzonen unterteilt, die charakterisiert sind durch spezifische karbonzeitliche Lithofazien und darauf zurückzuführende spezifische jungpaläogen-quartärzeitliche Deformation und Metamorphose. Den homogenen stratigraphischen Sockel bildet eine kondensierte Wechselfolge bituminöser Ton- und Siltsteine, feinkörniger Grauwacken und Kieselschiefer (Oberes Ems–Unteres Tournais), die in einem flachen, epikontinentalen Becken mit stark reduziertem Sedimenteintrag abgelagert worden sind. Durch seine nachfolgende, von Extensionsbrüchen gesteuerte Segmentierung entstanden zeitgleich bestehende Ablagerungsräume, gekennzeichnet durch vorwiegend siliziklastischen Eintrag. Bei Meeresspiegelhochstand progradierten diese Sedimente auf den Schelf, den unteren Kontinentalabhang und den Rand der Beckenebene. Die Sediment-Mobilisierung via Tempestit-Turbidit-Zyklen wurde vermutlich durch eine kurzdauernde klimatische Heißphase initiiert, die mit der Hangenberg-Krise korreliert ist. Die Formationen wurden einheitlich von regressiven, littoral-deltaischen und terrestrischen Siliziklastika überlagert. Anschließend erfolgte ihre Integration in den Sockel der meso-känozoischen, transgressiven Ablagerungen am passiven Rand der Adriatischen Mikroplatte. Oligozäne bis frühmiozäne Überschiebungen und durch Subduktion gesteuerte tektonische Versenkung bis ca. 25 km Tiefe, gefolgt von mio-pliozäner extensionsbedingter Exhumation waren von mehreren Scherzonensystemen gesteuert; wie z. B. den reaktivierten, oben genannten karbonzeitlichen Abschiebungen. Daraus entstanden die interferierenden, nahe aneinander grenzenden metamorphen Kernkomplexe des Monticiano-Roccastrada Gebietes. Während der Bildung der pliozän-quartären nordapenninischen Orokline kam es zu einer intensiven Zerblockung der Subzonen. Das Farmatal wird interpretiert als geomorphologischer Ausdruck exhumierter Transferstörungen, die Teil des Bruchsystems waren, welches die Ausbreitung der nach Osten migrierenden Toskanischen Magmenprovinz steuerten.

Keywords: Tuscan Palaeozoic, Carboniferous extensional basin, Hangenberg Crisis, late Palaeogene–Quaternary closely spaced interfering metamorphic core complexes, reactivated faults, oroclinal bending

Schlüsselwörter: südtoskanisches Jungpaläozoikum, karbonzeitliches Extensionsbecken, Hangenberg Krise, jungpaläogene-quartärzeitliche, interferierende, nahe aneinander grenzende metamorphe Kernkomplexe, reaktivierte Störungen, Orokline

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1. Introduction

The architecture of orogenic interiors is often complex because of protracted deformation produced over millions of years. Consequently, the structural style of these internal provinces may be difficult to unravel. However, detailed geological mapping of these settings provides decisive information, which helps to reconstruct the local geological history. This paper presents a detailed geological map of a large part of the Farma River Valley (see Electronic Supplement). It is a geological key area in the central part of the Monticiano-Roccastrada metamorphic core complex (MRMCC) in SW Tuscany (Northern Apennines, Italy).

The history of creation of the map at a 1:10,000 scale took many years. It started in 1979 with an EU-funded antimony prospection (Dehm et al. 1983) in Southern Tuscany, carried out in the Department of Geochemistry and Science of Ore Deposits, hosted in the Institute of General and Applied Geology, Ludwig Maximilians University in Munich, Germany. This project included a geological survey in the MRMCC with the aim to identify and map Early Palaeozoic formations. It was expected that they probably contain relics of time- and strata-bound stibnite mineralisations, which might have been the source of late Neogene epigenetic epizonal hydrothermal Sb_2S_3-pocket-ores. The latter occurred at the eastern and northern margins of the MRMCC and were mined (Müller & Klemm 1989).

Independently, I carried out the geological survey in the Farma River Valley – covering ca. 30 km^2 – from 1985–1990 and, after a long interruption, again since 2013. The exploration of this area was seen as ideally suited to reveal the formation and geological structure of the MRMCC. The current state of the geological exploration of the Farma River Valley is provided herein. Because of restrictions on the precision of the observations, the interpretations provided below should be regarded as advanced, although non-conclusive yet.

The geological map of the Farma River Valley (see Electronic Supplement) comprises an area occupied by the river loops present between the Ponte di Torniella at the western margin of the MRMCC and the Ponte di Petriolo at its eastern boundary. The valley displays over a linear distance of 13 km a conspicuous general west–east direction. This peculiarity will also be explained here because a comparable geomorphological element is absent in regions such as the Apuan Alps and Pisan Mountains to the north or the Monte Leoni to the south of the MRMCC; these regions are considered to be geological equivalents of the latter (Fig. 1).

Pyrite, stibnite, quartzous sand, lignite and travertine were mined in the Farma River Valley in the first half of the 20th century (Stea 1971; Tanelli 1983); kaolin and alum ore (Sammuri & Scapigliati 2022), as well as hydrothermal energy (Capezzuoli & Gandin 2005), are actually utilised at its western and eastern boundaries, respectively.

2. Geological setting

The working area is positioned in the central part of the late Palaeogene–Neogene low- to medium-grade MRMCC. Together with other discretely exhumed geological equivalents, it forms the Mid-Tuscan Ridge: an asymmetric, nonuniform bent arc, which marks the eastern boundary of the internal zone of the late Palaeogene–Neogene Northern Apennines fold-thrust belt (Boccaletti et al. 1987; Carmignani et al. 1996). This belt, which measures ~ 460 km along strike, developed in a back-arc setting at the Adriatic suture (Doglioni et al. 1991) and is confined between the orogenic segments of the Maritime Western Alps/Ligurian Knot and the Central-Southern Apennines (Vai 2001). The simplified structural zonation of the nappe edifice is illustrated in Fig. 1. The internal zone is the emerged, easternmost part of the North Tyrrhenian rift basin (Vannoli et al. 2021). Its recent geotectonic state has been described in detail by Brogi et al. (2005, cum lit.), Finetti (2006, cum lit.), Pascucci et al. (2007, cum lit.), Bellani (2018, cum lit.), De Franco et al. (2019, cum lit.), Liotta et al. (2021, cum lit.), Sillitoe & Brogi (2021, cum lit.); these investigations revealed that this part of the fold-thrust belt has undergone intense extension since the late Palaeogene.

Meso-Cainozoic geodynamic development: Mentioned tectonic nappes, which constitute the Tuscan mainland, are assigned to different lithofacial domains and correspond to oceanic (Ligurian) and epicontinental (Subligurian, Tuscan and Umbromarchean) depositional environments of the west-Mediterranean Neotethys, established between the Corsica-Sardinia and Adria microplates (Molli 2008). Their relative motions were constrained by the drifting directions of the encircling megaplates of Gondwana and Europe (Stampfli 2005; Le Breton et al. 2017). The orogenic cycle comprised the development, transformation and destruction of the abovementioned depositional realms (Fig. 2):

– Triassic to early Cretaceous divergence: The deposition of deepening-upward, onlapping first-order sequences onto the passive margin of the Adria microplate (Umbromarchean, Tuscan and Subligurian units) occurred, as well as the formation of the Ligurian oceanic crust and deposition of its pelagic cover sediments since the Dogger (Carmignani et al. 2001).

– Late Cretaceous to early Eocene: The convergence of both microplates via the oceanic subduction of the Ligurian crust below the Adria microplate occurred and the deposition of orogenic clastics in the trench began (Boccaletti et al. 1980; Stampfli 2005).

– Late Eocene to early Oligocene: The onset of the continental collision of the mentioned microplates and progressive subduction polarity reversal occurred (Malusà et al. 2015).

– Middle Oligocene to early Miocene: The eastward drift of the Corsica-Sardinia microplate, the continental subduction of the Adriatic lithosphere below the Corsica-Sardinia microplate and the beginning of the delamination of the lower plate occurred. The rollback of the Adriatic slab resulted in upwelling of the asthenosphere and the onset of back-arc extension, gravitational collapse and the thinning of

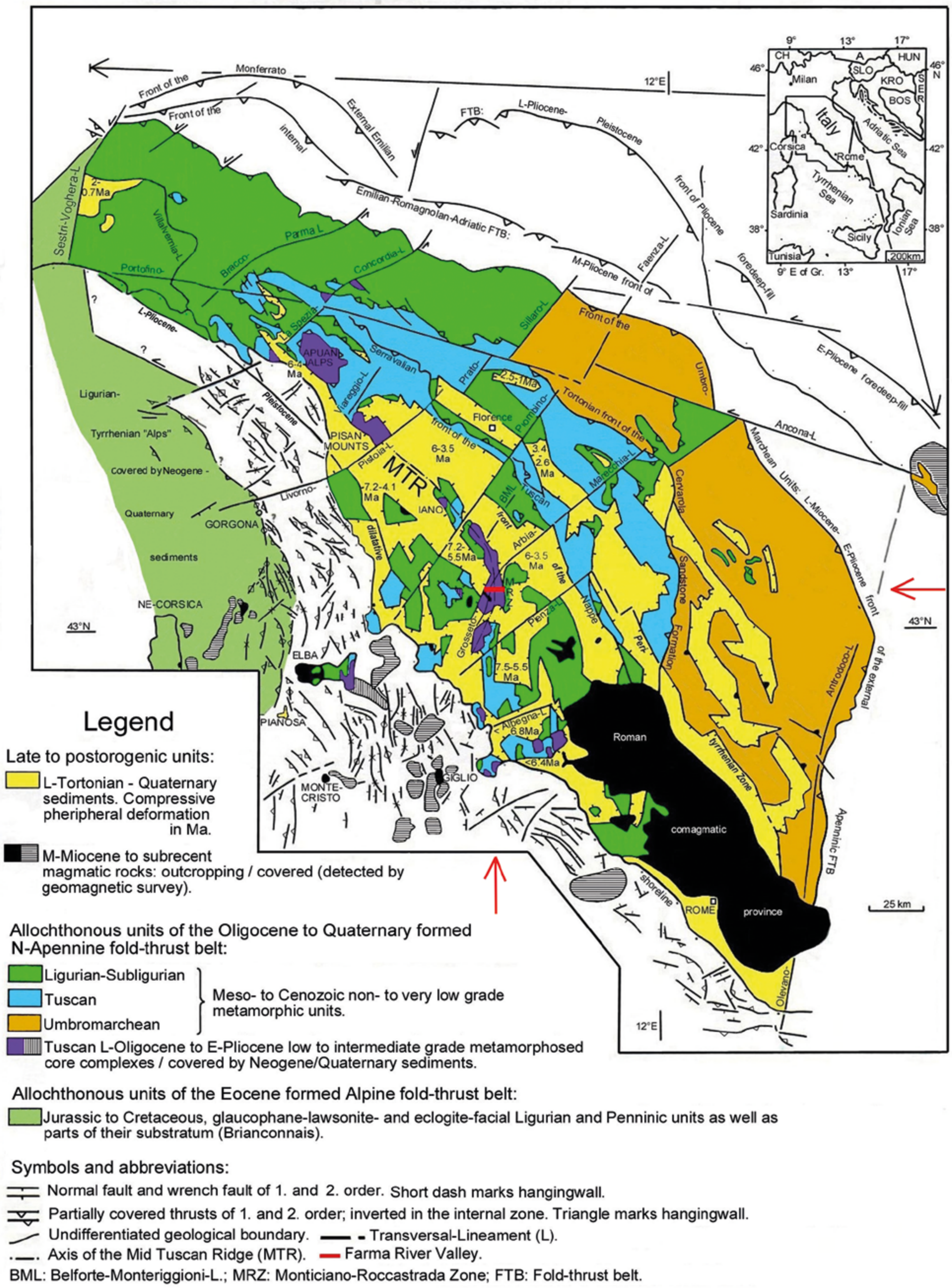


Fig. 1: Structural setting of the Northern Apennines fold-thrust belt. Ages indicate late Miocene to Pliocene compressive events, which were derived from seismically or stratigraphically determined unconformities present in deformed Neogene basins (Engelbrecht 1997b, cum lit.; Bonini et al. 2014).

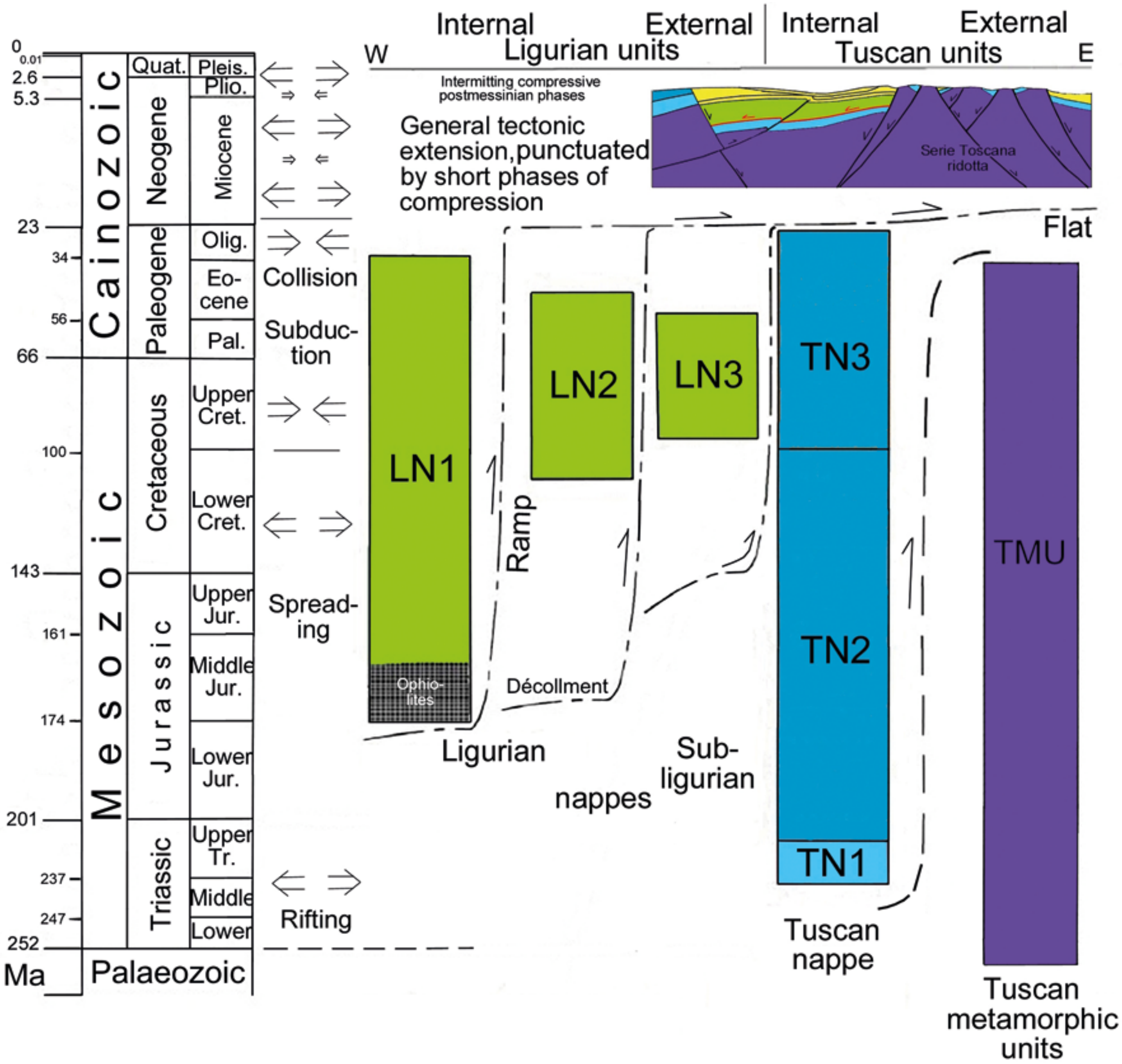
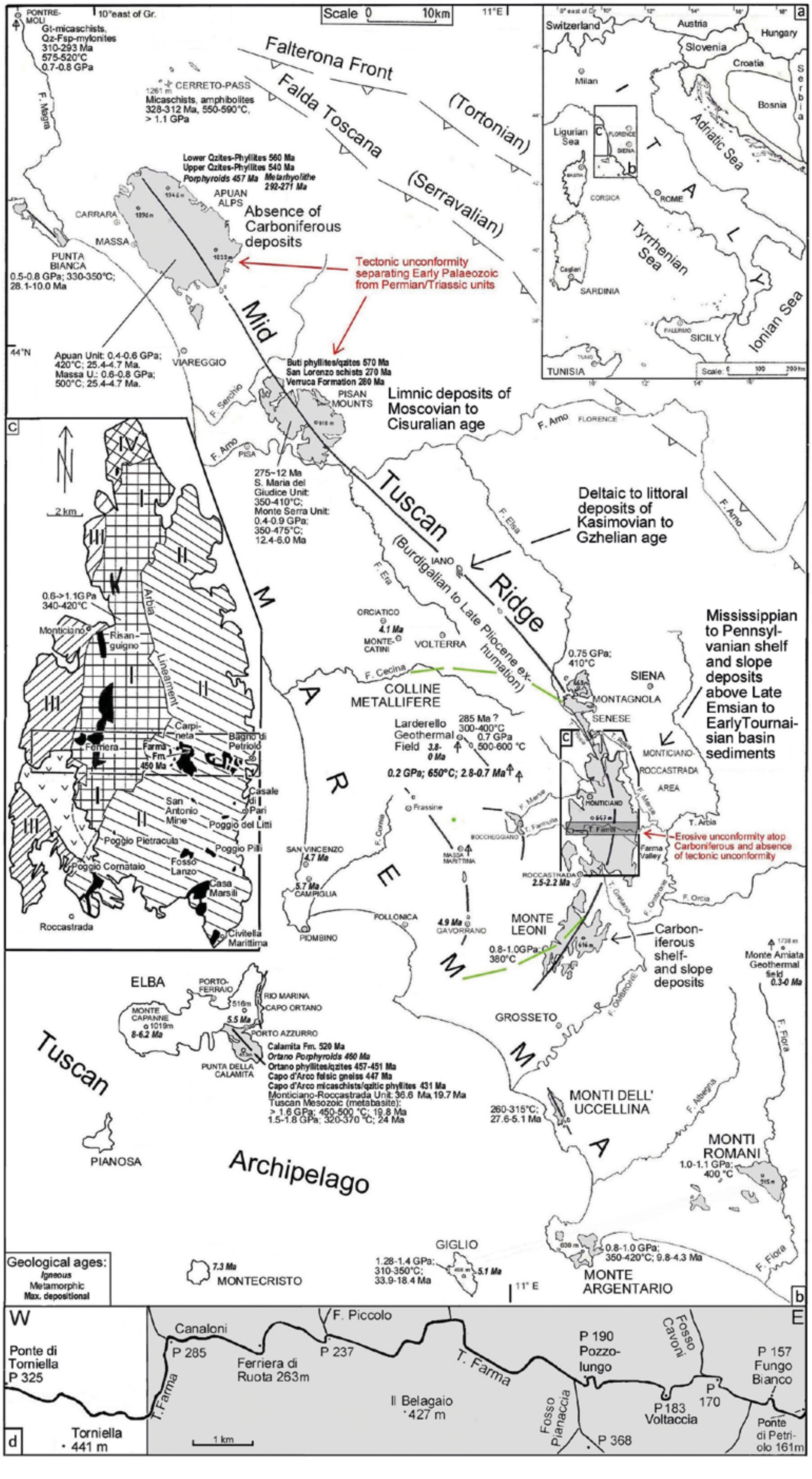


Fig. 2: Scheme representing the relationship between the Ligurian and Tuscan units of the internal zone and their Mesozoic and Cainozoic deformation phases (see text): Late Cretaceous–early Miocene tectonic stacking of the units via compression and thrusting, followed by Miocene–Quaternary multiphase extension, intermitted by short phases of compression. Decollement horizons developed in zones of crustal weakness, e.g. ophiolites and late Triassic evaporites. Abbreviations: LN – Ligurian nappe; LN1 – Ophiolite unit; LN2 – Santa Fiora unit; LN3 – Canetolo unit; TN – Tuscan nappe; TN1 – Late Triassic evaporites; TN2 – Late Triassic-Cretaceous carbonate and siliceous deposits; TN3 – Late Cretaceous–early Miocene turbidites; TMU – Tuscan Metamorphic units (Palaeozoic–Eocene). Additional symbols: yellow: Neoautochthonous graben deposits of late Miocene–late Pliocene age; red line: low-angle normal fault of middle Miocene age. Modified from Liotta (2002) and Brogi et al. (2014).

Fig. 3: General view of the Tuscan metamorphic units. Main structural, palaeotectonic and palaeogeographical data and relations. (**a**) Geographic overview. (**b**) Tuscan metamorphic units, late Neogene–Quaternary magmatites, drill sites and working area. Grey: metamorphic units (Cambrian–late Oligocene sediments, including Ordovician, Lopingian and middle Triassic volcanics), relics of the Tuscan nappe and subordinately Neogene and Quaternary cover. Dark grey: working area. Abbreviations: F. – Fiume (river); T. – Torrent. Triangle with vertical rod: drill site. Some igneous, metamorphic and maximum depositional ages are given according to Borsi et al. (1967a), Borsi et al. (1967b), Di Sabatino et al. (1978), Del Moro et al. (1982), Bertini et al. (1994), Brunet et al. (2000), Rossetti et al. (2008), Musumeci et al. (2011), Bonini et al. (2014), Lo Pò et al. (2015, 2018), Bianco et al. (2015, 2019), Papeschi (2015 cum lit.), Papeschi et al. (2020, 2022), Sirevaag et al. (2016), Paoli et al. (2016), Fulignati (2018), Vezzoni et al. (2018) and Ryan et al. (2021). Green semicircle and dot: geometric centre of the oroclinal arc comprising the southernmost part of the Mid-Tuscan Ridge. (**c**) MRMCC, dissected into subzones by transversal lineaments. Black: Late Palaeozoic units (Devonian–Carboniferous). Hatchings with numbers: tectono-stratigraphic homogeneous subzones I–IV: areas with differing degrees of tectonic deformation and metamorphism, consisting of Late Palaeozoic to late middle Triassic siliciclastics, covered by very low-grade to non-metamorphosed late Triassic evaporites, carbonates, siltstones, Miocene tectonic breccias and Pliocene to Quaternary sediments. V-signature: rhyolites of Torniella-Roccastrada (2.3–2.5 Ma). (**d**) Relevant geographical sites of the working area. Grey: mapped part of it. Abbreviations: F. – Fosso; T – Torrent; P – topographic reference point.


a
Switzerland
Austria
Hungary
Slovenia
Croatia
Bosnia
Serbia
Milan
Ligurian Sea
FLORENCE
SIENA
CORSICA
ROME
SARDINIA
Adriatic Sea
Tyrrhenian Sea
Ionian Sea
ITALY
PALERMO
SICILY
TUNISIA
Scale: 0 100 200 km
b
PONTREMOLI
10°east of Gr.
Gt-micaschists, Qz-Fsp-mylonites 310-293 Ma 575-520°C 0.7-0.8 GPa
F. Magra
Scale 0 10km
11°E
Falterona Front
Falda Toscana
(Tortonian)
(Serravalian)
1261 m CERRETO-PASS
Micaschists, amphibolites 328-312 Ma, 550-590°C, > 1.1 GPa
Lower Qzites-Phyllites 560 Ma
Upper Qzites-Phyllites 540 Ma
Porphyroids 457 Ma
Metarhyolithe 292-271 Ma
APUAN ALPS
Absence of Carboniferous deposits
CARRARA
MASSA
PUNTA BIANCA
0.5-0.8 GPa; 330-350°C; 28.1-10.0 Ma
Tectonic unconformity separating Early Palaeozoic from Permian/Triassic units
44°N
Apuan Unit: 0.4-0.6 GPa; 420°C; 25.4-4.7 Ma.
Massa U.: 0.6-0.8 GPa; 500°C; 25.4-4.7 Ma.
VIAREGGIO
Buti phyllites/qzites 570 Ma
San Lorenzo schists 270 Ma
Verruca Formation 280 Ma
PISAN MOUNTS
Limnic deposits of Moscovian to Cisuralian age
F. Serchio
F. Arno
PISA
FLORENCE
275~12 Ma S. Maria del Giudice Unit: 350-410°C; Monte Serra Unit: 0.4-0.9 GPa; 350-475°C; 12.4-6.0 Ma
Mid Tuscan Ridge
(Burdigalian to Late Pliocene exhumation)
Deltaic to littoral deposits of Kasimovian to Gzhelian age
IANO
F. Elsa
F. Era
ORCIATICO
4.1 Ma
MONTECATINI
VOLTERRA
F. Cecina
COLLINE METALLIFERE
0.75 GPa; 410°C
SIENA
MONTAGNOLA SENESE
Mississippian to Pennsylvanian shelf and slope deposits above Late Emsian to EarlyTournaisian basin sediments
Larderello Geothermal Field 3.8-0 Ma
285 Ma ? 300-400°C
0.7 GPa 500-600 °C
0.2 GPa; 650°C; 2.8-0.7 Ma
MONTICIANO-ROCCASTRADA AREA
Frassine
F. Merse
T. Farmulla
BOCCHEGGIANO
MONTICIANO
T. Farma
T. Arbia
Farma Valley
Erosive unconformity atop Carboniferous and absence of tectonic unconformity
MASSA MARITTIMA
ROCCASTRADA
2.5-2.2 Ma
F. Orcia
SAN VINCENZO 4.7 Ma
F. Cornia
5.7 Ma CAMPIGLIA
PIOMBINO
FOLLONICA
4.9 Ma GAVORRANO
MONTE LEONI
0.8-1.0GPa; 380°C
Carboniferous shelf- and slope deposits
F. Ombrone
1738 m Monte Amiata Geothermal field 0.3-0 Ma
MAREMMA
ELBA
PORTOFERRAIO
MONTE CAPANNE 1019m 8-6.2 Ma
516m
5.5 Ma
RIO MARINA
CAPO ORTANO
PORTO AZZURRO
PUNTA DELLA CALAMITA
Calamita Fm. 520 Ma
Ortano Porphyroids 460 Ma
Ortano phyllites/qzites 457-451 Ma
Capo d'Arco felsic gneiss 447 Ma
Capo d'Arco micaschists/qzitic phyllites 431 Ma
Monticiano-Roccastrada Unit: 36.6 Ma, 19.7 Ma
Tuscan Mesozoic (metabasite): > 1.6 GPa; 450-500 °C; 19.8 Ma
1.5-1.8 GPa; 320-370 °C; 24 Ma
GROSSETO
260-315°C; 27.6-5.1 Ma
MONTI DELL' UCCELLINA
F. Albegna
F. Fiora
MONTI ROMANI
1.0-1.1 GPa; 400 °C
Tuscan Archipelago
PIANOSA
7.3 Ma MONTECRISTO
GIGLIO
1.28-1.4 GPa; 310-350°C; 33.9-18.4 Ma
5.1 Ma
11° E
630 m
0.8-1.0 GPa; 350-420°C; 9.8-4.3 Ma
MONTE ARGENTARIO
Geological ages:
Igneous
Metamorphic
Max. depositional
c
N
2 km
IV
I
II
III
0.6->1.1GPa 340-420°C
Monticiano
Risanguigno
Arbia Lineament
Carpineta
Bagno di Petriolo
Ferriera
Farma Fm. 450 Ma
Casale di Pari
San Antonio Mine
Poggio del Litti
Poggio Pietraculo
Poggio Pilli
Fosso Lanzo
Poggio Comataio
Casa Marsili
Civitella Marittima
Roccastrada
d
W
E
Ponte di Torniella P 325
Torniella • 441 m
T. Farma
Canaloni
P 285
F. Piccolo
P 237
Ferriera di Ruota 263m
Il Belagaio • 427 m
T. Farma
Fosso Pianaccia
P 368
P 190 Pozzolungo
Fosso Cavoni
P 183 Voltaccia
P 170
P 157 Fungo Bianco
Ponte di Petriolo 161m
1 km

the internal part of the Apenninic fold-thrust belt (Carmignani et al. 2001; Vignaroli et al. 2009; Tavarnelli et al. 2021).

– Middle Miocene to Recent: Ongoing Adriatic slab retreat caused the eastward migration of a paired compressional-extensional front, where the thrust front is closely followed by an extensional deformation front: compression generated basement-involved, thick-skinned, far-travelled, rootless nappes, whose transport was controlled by zones of crustal weakness, present in the stratigraphy as incompetent lithologies and in the tectonic structure as palaeotectonic faults separating differing depositional systems (Kligfield 1979; Tavarnelli 1996; Engelbrecht 1997a; Cosentino et al. 2010). Extension resulted in the origin of the North Tyrrhenian back-arc basin (Loreto et al. 2021), the continued thinning of the internal part of the fold-thrust belt and eastward-migrating magmatism (Rossetti et al. 2008; Poli & Peccerillo 2016). Thinning first occurred via low-angle normal faults. This led to the unroofing and lateral segmentation of Meso-Cainozoic Tuscan and Ligurian units above late Triassic evaporites and stratigraphic gaps (‘serie Toscana ridotta’), as well as footwall uplift and the surface emergence of metamorphic core complexes (Signorini 1947; Brogi et al. 2005; Brogi 2008; Barchi 2010; Bonini et al. 2014). The dip directions of the mentioned low angle normal faults were oriented perpendicular to the NW–SE strike direction of the uplifting metamorphic chain. Since the latest middle Miocene, an eastward-younging horst and graben system, fanning out southeastward, with vertical throws up to several km, developed (Martini et al. 2001; Pascucci et al. 2007). Short compressive pulses between the late Miocene and Pliocene deformed peripherally the basin fills of the internal zone (Bonini et al. 2014; Viola et al. 2018).

Metamorphic core complexes, as defined by Platt et al. (2015) and Roche et al. (2018), originated in the inner zone via forelandward-propagating extensional pulses/events, caused by exhumation in the early Oligocene, early Miocene and Messinian to early Pliocene. These events occurred with an eastward-younging trend, and their peaks of subduction-related metamorphism decreased from (U)HP-LT grade in the Eocene-Oligocene to HP-LT grade in the Neogene (Vignaroli et al. 2009). Their regional distributions are illustrated in Fig. 3. The core complexes are interpreted as interfering metamorphic core complex chains, as defined by Tirel et al. (2009), because they share a shear zone established in late Triassic evaporites and formed above a lower crust and sub-Moho mantle of very low strength.

The post-Messinian partitioning of the Adriatic-Ionian slab caused belt-parallel compression, the oroclinal bending of parts of the longitudinal ridges of the Northern Apennines and the development of transfer faults (Lucente & Speranza 2001; Lucente et al. 2006; Viti et al. 2015; Lo Bue et al. 2021). This bending caused additional tectonic deformation in the internal zone and the curvature of the southernmost quarter of the Mid-Tuscan Ridge. The MRMCC is positioned at the apex of this arc, where its general strike direction turns in its southern quarter over a short distance of 30 km by 75° from Apenninic (NW) to ca. anti-Apenninic (NNE) directions.

The Palaeozoic history of the substratum of the working area is only briefly summarised because of the extreme complexity of the processes which acted near the boundaries separating the megaplates Gondwana and Europe.

The substratum, on which the geological formations in the working area were deposited, belongs to the Adria Microplate. During the late Ordovician, it formed a constituent of the eastern part of the northern Perigondwanan shelf, adjacent to the Armorican Spur (Stephan et al. 2019). Silurian closure of the Rheic Ocean, back-arc spreading and opening of the Palaeotethys caused rifting of the Adria Microplate from its mainland and its northward drift along with other terranes. Late Devonian–Pennsylvanian convergence and continental collision resulted in the integration of the Adria Microplate into the southern part of the European Variscides, made up of a complex amalgamation of Gondwana-derived terranes (Stampfli 2005; Sirevaag et al. 2016). Late Palaeozoic gravitational collapse of orogenically thickened crustal parts caused the formation of extensional basins. The filling history of one of these basins is described below.

3. Previous work in the Farma River Valley

Geological investigations in the Colline Metallifere, which include the Farma River Valley, go back many centuries and were initiated by the exploration of raw materials (Sammuri & Pantaloni 2018). The identification of strata of Carboniferous age in the Farma River Valley entailed the ‘Sondaggio del Belagaio’ borehole, which was carried out in 1941 to explore hard coal seams; however, the drill was unsuccessful (Redini 1942; Novarese 1946; Lazzarotto & Moretti 1973). Redini (1958: 611) recognised a bipartition of Carboniferous coeval marine clastic deposits in the Farma River Valley: strata representing shallow water environments occupy the western region and those representing deep water environments crop out in its eastern region. He stated a gradual passage upward into the Verrucano Group. Cocozza (1965) confirmed via macrofossil and microfossil determinations the age attribution given by Redini but postulated a late Hercynian compressive event and thus a tectonic unconformity separating the Carboniferous formations from the Verrucano Group of late Permian to early Triassic age. Consequently, Cocozza et al. (1974) interpreted the Carboniferous formations as flysch and molasse deposits (Farma and Carpineta formations): relics of an intensely folded foreland basin fill. This was specified by Cocozza et al. (1975), Azzaro et al. (1976), Cocozza et al. (1978) and Cocozza et al. (1987); the Carboniferous dichotomy ascertained by Redini (1958) was rejected. Refined palaeontological investigations further constrained the Carboniferous ages attributed to the late Moscovian Farma Formation, the Viséan–late Moscovian Carpineta Formation and the Moscovian Sant’ Antonio Limestone Formation and equivalents (Pasini 1978a, b; Pasini 1979a, b; Pasini & Winkler Prins 1981; Pasini & Vai 1997).

Costantini et al. (1988), however, re-established the bipartition of the Late Palaeozoic units in the working area and postulated a shallow west-dipping major thrust. It separates two tectonic and lithofacial subunits, in which several Hercynian and Alpine compressive events were thought to have occurred. The internal subunit, containing shallow marine early Mississippian strata (Poggio al Carpino Formation), was interpreted to be thrust eastward onto the external subunit with deep marine deposits of a Carboniferous foreland basin; see also Conti et al. (1991) and Musumeci et al. (2011). Thus, the finding of Signorini (1967), confirmed by Lazzarotto & Moretti (1973), that the fault juxtaposing the two subunits consists of a steeply east-dipping normal fault offsetting east-directed thrust surfaces was rejected.

Engelbrecht (1997a) agreed on the abovementioned subdivision into subunits, which, however, he found to be separated by steep, complex faults, offsetting late Palaeogene–early Neogene thrusts. Instead of Hercynian deformation, intense Alpine contractional deformation was recognised in the Palaeozoic formations. The stratigraphic successions of the subunits start with a condensed uniform base, consisting of the Risanguigno Formation (Bagnoli & Tongiorgi 1979) of Emsian to probable early Tournaisian age. It is overlain with erosive unconformities by the coeval Poggio al Carpino, Carpineta and Farma formations. These are interpreted as a storm-influenced Carboniferous tempestite-turbidite megasequence deposited on the shelf, slope and at the margin of the basin floor of an epicontinental extensional basin. The megasequence is covered by regressive, shallow marine Permian siliciclastics (Civitella Marittima Formation) and the siliciclastic Verrucano Group.

Spina et al. (2001) reported that palynomorphs sampled in the Poggio al Carpino Formation are Guadalupian–Lopingian in age. The resulting stratigraphic gap downward to the Risanguigno Formation was seen as an emersion unconformity (Lazzarotto et al. 2003; Decandia et al. 2004). Engelbrecht (2008) specified that the deposits of the mentioned extensional basin, interpreted by him as Carboniferous failed rift filling, were generated by highstand-shedding and that patches of late Gzhelian to Asselian Fusulina Limestone (Pasini 1991) were present; they were partially redeposited in the Verrucano Group.

Aldinucci et al. (2008a) and Spina et al. (2021) reported that sporomorphs sampled in the Farma and Carpineta formations are Guadalupian–Lopingian in age. Capezzuoli et al. (2021) stated that palynomorphs sampled in the Risanguigno Formation are of early–middle Mississippian age. Brogi et al. (2023: 17–20, Fig. 14) postulated a uniform and homogenised, isochronous hiatus that separates the mentioned Guadalupian–Lopingian formations from the Verrucano Group (middle to early late Triassic).

Giuntoli & Viola (2022) determined the peak P-T conditions to which a mesoscopic compressive duplex structure (Casini et al. 2007) of the internal subzone was subjected during Eocene-Oligocene subduction: 1.1 GPa and 350 °C, thus confirming the results of Brogi & Giorgetti (2012).

Further interpretations of the geological ages and palaeotectonic development of the units are given in meta-analyses by Perrone et al. (2006), Cassinis et al. (2012), Cassinis et al. (2018) and Molli et al. (2020).

4. Methods

The Farma River Valley was surveyed by applying the classic methods described in Compton (2017). Several topographic maps of the Istituto Geografico Militare (Florence) were used for localisation. A GPS handheld device was also used. Tectonic measurements were collected with a geological compass (Clar type). Terminologies concerning sediments and tectonic elements follow those of Miall (2010) and Ramsay & Huber (1987), respectively. The geological observations were annotated in several data books and later transferred to drafts of preliminary map parts. The digital compilation of the latter and the final preparation of the map and stratigraphic profile were accomplished with the graphics program Adobe PhotoDeluxe Business Edition 1.

5. Stratigraphic succession

The formations are described from the tectono-stratigraphically homogeneous subzones I–II, which are characterised by distinctive Carboniferous lithofacies. The subzones are separated by a major fault, attributed to the Arbia lineament (Liotta 1991; Brogi et al. 2014). Their lithostratigraphic successions are detailed below (Fig. 4).The stratigraphic features of formations cropping out in subzone III are not dealt with below, and will be presented in a forthcoming contribution. Remark: Subduction-related glaucophane schist metamorphism generated pre-Carnian metasediments. In order to avoid unnecessary repetitions, the prefix ‘meta’ is omitted in the following section.

Tuscan metamorphic units
Mesozonal metamorphic units
Monticiano-Roccastrada Unit

5.1 Risanguigno Formation (late Emsian to early Tournaisian) (ex Boccheggiano Formation)

The lithologies present are chert, lydite, thin-layered marble, black coloured, in places nodular limestones, black shale, pyritiferous alginite-bituminite, arenaceous clay schist, dark grey siltstone and fine- to medium-grained greywacke; see also Bagnoli & Tongiorgi (1979), Bartoletti & Tongiorgi (1982), Costantini et al. (1988), Engelbrecht (1997a, 2008), Engelbrecht et al. (1988) and Capezzuoli et al. (2021). Its maximum exposed thickness is estimated at 40 m. Sample nr. 535 of the limestones yielded relics of crinoids, cephalopods, conularids and the conodont species belonging to the families of the Gnathodontidae and Polygnathidae; the species indicate late Eifelian (M. Piecha, written communication). A detailed contribution about these findings is in preparation.

The stratigraphic upper boundary is an erosive unconformity (Fig. 5a–c) and is defined by the sudden onset of

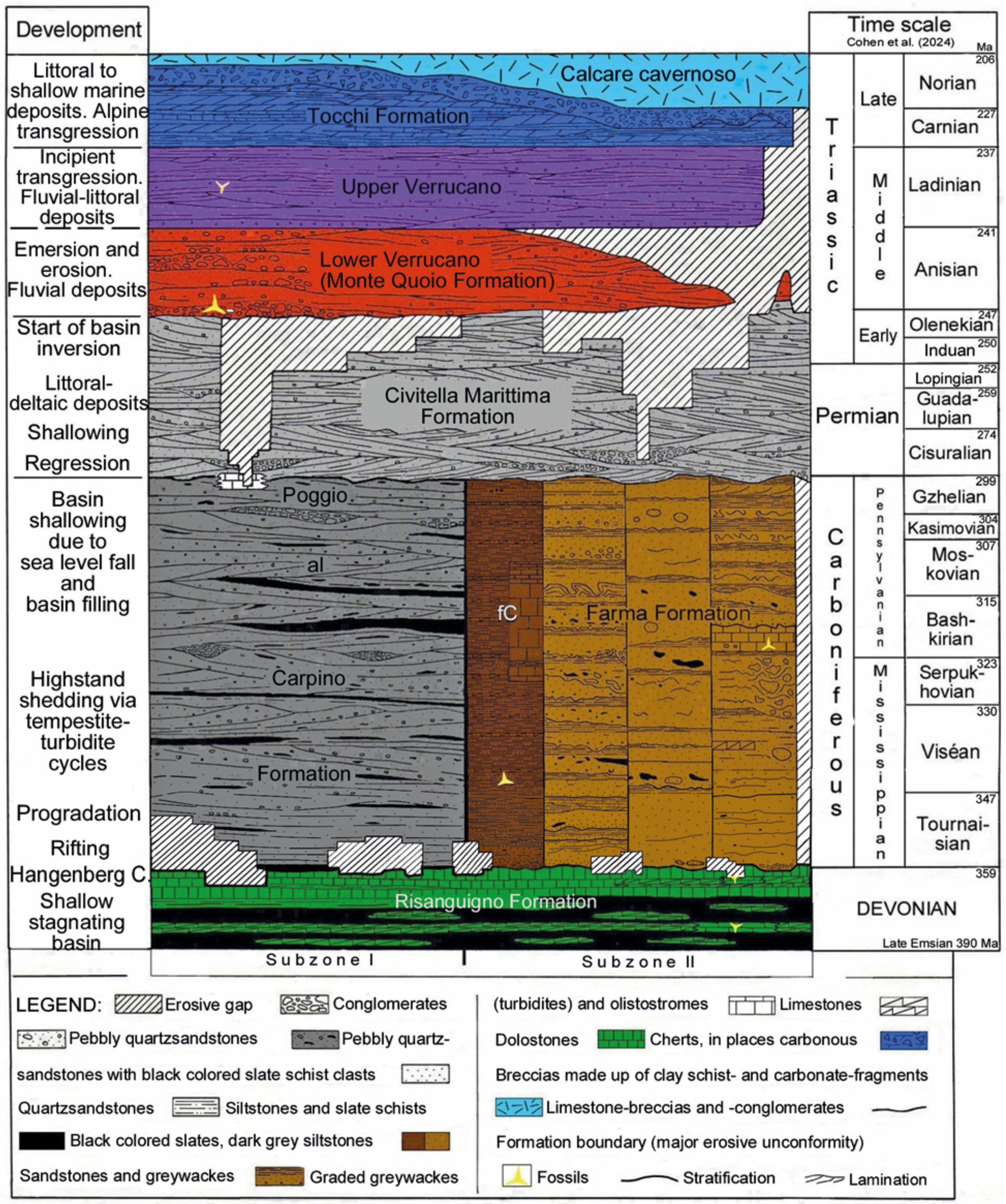


Fig. 4: Schematic stratigraphic subdivision of the Late Palaeozoic to late Triassic formations in the Farma River Valley (MRMCC). Time scale according to Cohen et al. (2025). Abbreviations: fC – Carpineta Formation; C. – Crisis.

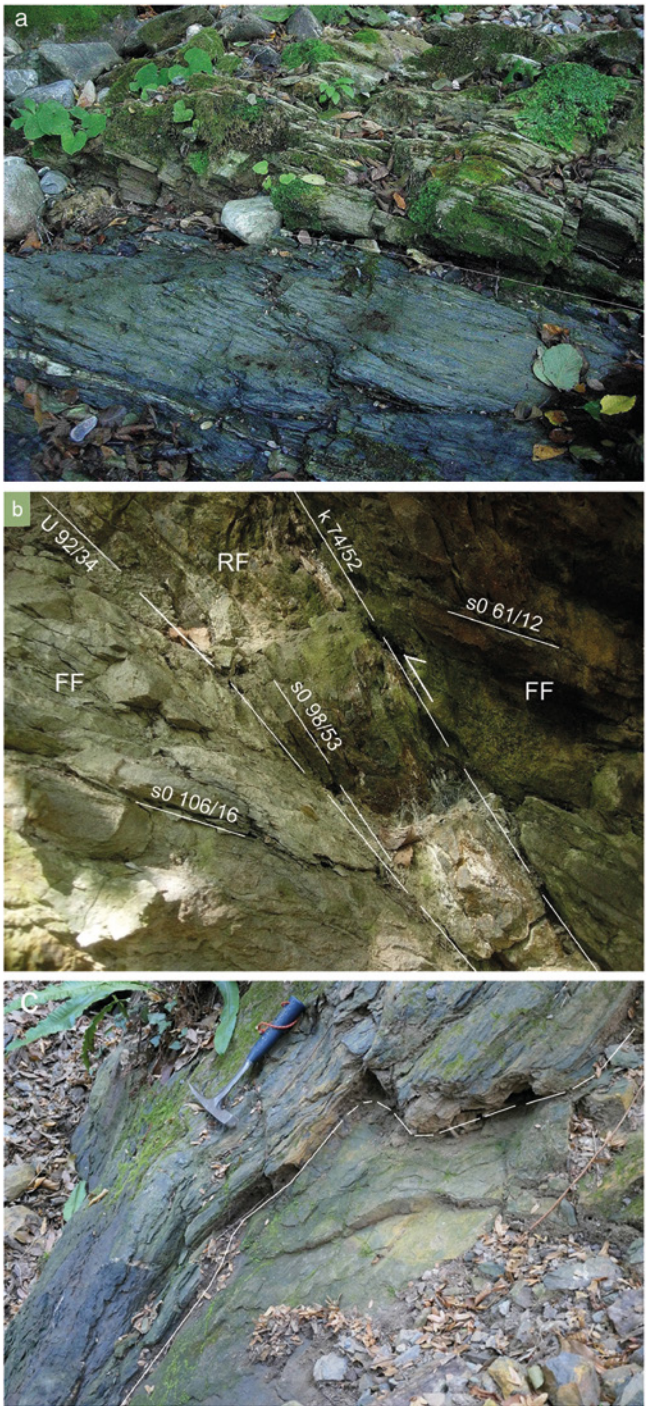


Fig. 5: Lithostratigraphic boundaries between Late Palaeozoic formations. (**a**) Dark grey to blackish, mm-scale-laminated, schistous siltstones and cherts of the Risanguigno Formation (s_0 280/35) below light grey, pebbly quartz sandstones of the Poggio al Carpino Formation (s_0 283/37). Loc.: ca. 450 m WNW of Ferriera di Ruota (P 263). Width of outcrop: 1.2 m. View west. (**b**) Centre: 1.0–1.5 cm thick chert laminae of the Risanguigno Formation (s_0 98/53) above thick-bedded, pebbly quartz greywackes of the distal part of the Farma Formation (s_0 106/16). Erosive unconformity U 92/34. Strata overturned by Oligocene–early Miocene tectogenesis. Thrust (k 74/52) postdates overturning. The tectonic hanging wall consists of the distal part of the Farma Formation (s_0 61/12). Loc.: 70 m to the east of the confluence between Fosso Cavoni and the Farma River. View north. Outcrop width: 0.8 m. (**c**) mm-scale alternation of chert and black schistous siltstone of the Risanguigno Formation (s_0 260/48) above quartz greywackes of the Carpineta Formation (s_0 262/44). Strata overturned by Oligocene–early Miocene tectogenesis. Loc.: Fosso Pianaccia at 345 m elevation, 120 m north of P 368. View north. Scale: Hammer 28 cm.

coarse-grained siliciclastics of the coeval Carboniferous depositional systems described below.

The contacts are slightly to moderately erosive unconformities. The presence of chert clasts in the stratigraphic cover of the Risanguigno Formation indicates that more pronounced erosive contacts exist (Engelbrecht 1997b; Capezzuoli et al. 2021). The base of the Risanguigno Formation is not exposed.

According to the lithologies and sparse microfauna, the depositional setting of the formation is interpreted as a low-energy, anoxic to dysoxic, partially starved marine-epicontinental, extensional basin with strongly reduced, predominantly fine-grained clastic input alternating with the deposition of inorganic and/or organic chert. The ubiquitous lamination at the cm and mm scales indicates that burrowing fauna was nearly absent. The conodont species determined by Bagnoli & Tongiorgi (1979) preferred palaeoenvironments as follows: Ozarcodinins inhabited zones between the peritidal and open marine zones and *Panderodus unicostatus* (Branson & Mehl 1933) lived in shallow subtidal zones (Murdock & Smith 2021; Terrill et al. 2022). According to these few indications, the formation probably originated in a shallow, subtidal, epicontinental basin, where deposition occurred predominantly below the storm wave base.

Carboniferous

Carboniferous of subzone I

5.2 Poggio al Carpino Formation (Tournaisian to late Pennsylvanian) (ex Ferriera Formation)

It was formally established by Cocozza et al. (1975). The lithologies are the following:

– Polymictic and oligomictic, < 1 m thick, coarse-grained, locally matrix-supported, lens-shaped, poorly organised quartz conglomerate layers, containing locally abundant intraclasts.

– White to dark grey, coarse-grained, up to 2 m thick, pebbly quartz-sandstones and quartzites rarely containing cm-thick phytoclastic accumulations.

– Dark grey to black-coloured, < 2 m thick arenaceous siltstone and silty pelite layers with interstratified mm-thick chert and dark grey, slightly carbonous and siliceous, < 35 cm thick boudinaged dolostones.

More information is given in Engelbrecht (2000) and Spina et al. (2001). The stratigraphic upper boundary of the Poggio al Carpino Formation exhibits a diffuse passage to the Civitella Marittima Formation, since the dark grey to black mudstones and clasts gradually disappear. The tempestite- and tsunamite-influenced deposits were laid down at the tropical inner shelf and lower shoreface. Large internal solitary waves (Pomar et al. 2012) probably reworked these sediments. The lower boundary is a slight erosive unconformity to the Risanguigno Formation (Fig. 5a).

Its true thickness, which is masked by intense Oligocene-Pliocene deformation, is estimated to be ca. 300 m.

Carboniferous of subzone II

5.3 Farma Formation (Tournaisian to Pennsylvanian)

It was established by Cocozza et al. (1974). An additional lithofacial subdivision according to the proximality trend of its heteropic subunits is presented herein. Each subunit crops out at specific sites along the course of the Farma River whose elevations above sea level are indicated as P 190 (proximal subunit), P 183 (intermediate subunit), P 170 West (distal subunit), and P 170 East (ultradistal subunit), respectively.

5.3.1 P 190 – site Pozzolungo (proximal subunit)

It consists of grey to dark grey subgreywackes, grading vertically into black siltstones and mudstones (Bouma cycles). The stratal thickness varies from 8 to 62 cm and is, on average, 30 cm. Metre-sized channelling at erosive bases is rare. Sedimentary structures consist of erosive ripple lamination, cm-sized convolutions, load casts with bulges up to 25 cm in amplitude, slumps, flame structures, clay wisps, sedimentary sills, injections, flute casts and scour, prod, skip and groove casts. Rarely, cm-thick phytoclastic accumulations are contained in arenaceous parts. The quantity of the mudstone interval varies between 25 and 70%. Rarely, laterally thinning, <2 m thick, normally graded, polymictic conglomerate layers are present: these contain boulders consisting of rolled-up siltstone, coarse-grained greywacke and bioclastic limestone. More details can be found in Aldinucci et al. (2008a) and Engelbrecht (2019). The strata near the locality P 190 of this subunit are Moscovian in age (Pasini 1979b).

The sediments are interpreted as slightly water-undersaturated inner fan turbidites and channelised debris flows or fluxoturbidites deposited on the lower slope.

At the stratigraphic top of the subunit, a moderately erosive unconformity separates it from the overlying light grey, coarse-grained quartz sandstones of the Civitella Marittima Formation (Fig. 6a).

The true thickness of this subunit is difficult to calculate because of intense late Palaeogene–Neogene deformation; it is tentatively estimated to be ca. 450 m.

5.3.2 P 183 – site Voltaccia (intermediate subunit)

It is made up of grey to dark grey, locally microconglomeratic, often massive and indistinctly graded, medium- to coarse-grained subgreywackes, passing upward predominantly with hiati into volumetrically suppressed, dark grey to black mudstones (incomplete Bouma cycles). The layers exhibit tabular, wavy, ripple and flaser lamination; occasionally hydroplastic deformations such as load casts, convolutions, ball-and-pillow structures and drag marks occur. The strata thicknesses vary from 0.16 to 7.92 m; on average, the thickness is 0.95 m. The fossil content consists of crinoidal detritus, phytoclasts < 1 cm, driftwood fragments < 30 cm and ichnofauna.

The sediments are interpreted as water-saturated middle fan deposits of unconfined frontal splays and rare grain flows at the slope-basin transition, probably representing the depocentre of the Farma turbidite fan system.

The true thickness can hardly be calculated reliably because of complex late Palaeogene–Neogene deformation; it is tentatively estimated to be ca. 550 m.

A faintly erosive unconformity at the stratigraphic top separates it from the Civitella Marittima Formation above (Fig. 6b).

5.3.3 P 170 West (distal subunit)

The stratigraphic lower part is made up of medium to very thick bedded turbidites (< 2.10 m) consisting of coarse-grained, locally pebbly, either massive or faintly graded subgreywackes, interlayered with laterally thinning, maximum 1.2 m thick sedimentary breccias, mélanges and conglomerates, which contain intraclasts up to boulder size interspersed in varying amounts of sheared mudstone matrix. Near the base of this subunit the Bouma gradation is partially not developed. The stratigraphic upper part of the subunit is composed of turbidites consisting of grey to dark grey subgreywackes that grade into black mudstones, which volumetrically constitute 10–75% of the cycles. The stratal thickness varies from 0.10 to 0.58 m and is on average 0.21 m. The coarse-grained basal parts of the strata are occasionally disrupted and develop laterally into sedimentary mélanges.

Separated by a fault, another part of this subunit consists of < 63 cm thick, graded or massive subgreywacke layers often fragmented by networks of mm- to cm-thick fractures filled with injections consisting of shale and limestone, interstratified with disorganised, coarse-grained conglomerates containing intraclasts, white quartzite and quartz in a convoluted mudstone matrix.

This subunit is interpreted as the thinning- and fining-upward section of a distal outer fan part with basal debris flows and suspension currents, covered by water-oversaturated turbidites, deposited at the basin margin's floor. Porewater overpressure locally caused the fragmentation of strata into mélanges (Festa et al. 2012).

A slightly erosive unconformity at the stratigraphic top of this subunit separates it from the Civitella Marittima Formation above (Fig. 6c).

Fig. 6: Lithostratigraphic upper boundaries of the Farma Formation; brachiopods contained in the Carpineta Formation. (**a**) Slightly erosive unconformity between dark grey turbidites of the proximal part of the Farma Formation (s_0 17/22) below and light grey quartz sandstones of the Civitella Marittima Formation (s_0 358/31). Loc.: Fosso Pianaccia, westernmost branch at 425 m elevation. View NNW. Scale: Hammer 28 cm. (**b**) Thin-layered, light grey quartz sandstones of the Civitella Marittima Formation (s_0 271/38) below dark grey turbidites of the intermediate part of the Farma Formation (s_0 310/53). Strata overturned by Oligocene–early Miocene tectogenesis. Loc.: 90 m SE of P 183. View WNW. (**c**) Dark grey turbidites of the distal part of the Farma Formation below the light grey, thick-bedded quartz sandstones of the Civitella Marittima Formation. Strata subhorizontal. Loc.: near the confluence of Fosso Cavoni with the Farma River. View SW. Photo taken in April 2016. (**d**) Slightly erosive unconformity between black siltstones and fine-grained quartz greywackes of the Carpineta Formation above and light grey, fine-grained quartz sandstones of the Civitella Marittima Formation. Both dip shallowly to the west; strata overturned by Oligocene–early Miocene tectogenesis. View WNW. Loc.: 130 m ENE P 190. (**e**) Order Orthotetida: Genus *Orthotetes* Fischer de Waldheim, 1829 (Carboniferous–Cisuralian): Ventral internal mould with a delthyrial chamber and robust median septum; external cast. Loc.: Fosso Pianaccia, easternmost branch at 345 m elevation. (**f**) Order Productida: Genus *Antiquatonia* Miloradovich, 1945 (Viséan-Serpukhovian): Ventral internal mould, which displays a muscle attachment area; external cast. Loc.: see Fig. 6e.

5.3.4 P 170 East (ultradistal subunit)

It consists of monotonous, dark grey, fine-grained greywackes and mudstones containing rarely mm-sized chert laminae, < 1 dm thick carbonate layers and graded subgreywackes. Locally, the latter carry intraclasts, as well as carbonate pebbles and boulders. 8–10 m of thick-bedded, black-coloured limestones containing accessorily quartzite boulders are intercalated. Conodonts were isolated from these calciturbidites: Sample nr. 356 contained a fragment of a species belonging to the family of the Palmatolepidae, which indicates a late Devonian age; and samples nr. 357–358 yielded besides crinoid fragments, ramiform conodonts and plant relics two species belonging to the order of the Conodontophorida, which indicate a Bashkirian age of this part of this subunit (M. Piecha, written communication). The Palmatolepidae specimen was probably reworked from the upper part of the Risanguigno Formation (see Fig. 4). A detailed contribution about these findings is in preparation.

The subunit is interpreted as a distal fan part deposited at the basin's margin floor.

Due to intense deformation, the true thickness of this ultradistal subunit cannot be reliably calculated; it is estimated to be ca. 400 m.

5.4 Carpineta Formation (Tournaisian to late Pennsylvanian)

It was established by Cocozza et al. (1974). The major constituents consist of grey, dark grey to black, dm to 1 m scale layered, mm to 2 cm scale tabular to low-angle cross-laminated, monotonous, in part calcareous siltstones and fine-grained greywackes, which are interstratified with thin, dark grey, slightly silicified boudinaged dolostones. Ca. 1.5 m above the stratigraphic base, a condensed succession of faintly silicified, dark grey siltstones and fine-grained greywackes occurs, which contain unsorted accumulations of brachiopods, rarely bivalvular lamellibranches, crinoid fragments and plant debris. The brachiopods consist predominantly of orthotetids and productids. The genera identified are *Antiquatonia* Miloradovich, 1945 (indicative of Viséan–Serpukhovian), and *Orthotetes* Fischer de Waldheim, 1829 (Carboniferous–Cisuralian; U. Jansen, Senckenberg Institute, Frankfurt on the Main, Germany, written communication; Fig. 6e, f). A detailed contribution about these findings is in preparation. Accessorily up to 10 m thick black-coloured limestones, which are at the base massive and towards the top plane stratified, are present; they contain foraminifers and calcareous algae of middle Pennsylvanian age (Pasini 1978b).

The stratigraphic contact of the Carpineta Formation upward to the Civitella Marittima Formation is either sharp (Fig. 6d) or consists of a < 1 m thick resedimentation horizon. The stratigraphic contact of the Carpineta Formation downward to the Risanguigno Formation represents a slight erosive unconformity (Fig. 5c).

The true thickness of the Carpineta Formation can hardly be reliably calculated because of complex Palaeogene–Neogene deformation; it is estimated to be ca. 350 m.

The Carpineta Formation probably represents an epicontinental, predominantly low-energy depositional area below the storm wave base: the outer shelf region. The rarity of fossil fauna, scarcity of bioturbation and presence of organic matter in the intensely laminated strata point to a prevailing stagnating, dysoxic milieu, most likely caused by eutrophication. However, during the deposition of the biogenic limestones, interpreted here as small-scale biostromes containing rhodophycea and foraminifera, a better oxygenated milieu temporarily existed. Rare ruditic and accessory fossiliferous layers constitute mobilised and redeposited matter transported by bottom currents, which originated during tempestite events (Aigner 1985) or surging internal waves (Pomar et al. 2012).

5.5 Civitella Marittima Formation (Permian–early Triassic)

It was established by Costantini et al. (1988) and represents the uniform stratigraphic cover of the Carboniferous heteropic units. The formation consists of:

– Oligomictic, clast-supported, poorly sorted, wedging-out, thick-bedded (< 4 m), massive to indistinctly laminated, white to light-grey quartz conglomerates with clast sizes of approximately 2 cm on average (max. 20 cm). Occasionally irregular interstratifications consist of grey, cm-thick siltstones and fine-grained quartz sandstones.

– Grey-green, moderately sorted, tabular to high-angle cross-bedded pebbly quartzites, quartz sandstones and phyllitic quartz sandstones.

– Centimetre-thick, light grey to yellow, boudinaged dolomitic caliche layers.

In the uppermost part of the formation, the colours merge into differing degrees of bluish-grey, indicating a terrestrial milieu. The upper stratigraphic boundary of the Civitella Marittima Formation consists of moderate to strong erosive unconformities. Considerable portions were reworked into the Lower Verrucano Formation (Engelbrecht 1997a, 1997b). This is also indicated by conspicuous thickness variations. In subzone I, the stratigraphic cover of the Civitella Marittima Formation is regularly formed by the Lower Verrucano; in subzone II, however, its cover consists locally of the Upper Verrucano Formation and rarely of the Tocchi Formation. The inferred hiati can be explained by erosion, non-deposition or replacement. The latter is suggested by locally very high thicknesses of the Civitella Marittima Formation in subzone II.

The stratigraphic contacts downward are characterised by slight to moderate erosive unconformities (see above). Boulder-sized lydite clasts (maximum diameter of 16 cm) in the quartz conglomerates suggest that locally the Risanguigno Formation might have formed the stratigraphic substratum of the Civitella Marittima Formation, indicating an intense erosive unconformity.

The true thickness of the Civitella Marittima Formation cannot be reliably calculated because of its deformation; it is estimated to range from 5 to 300 m.

Its sedimentary fabric and the fact that it forms a uniform stratigraphic cover above the coeval Carboniferous formations imply that after the Carboniferous progradational phase, the depositional area developed into a regressive, shallowing, littoral-deltaic zone characterised by highly differing energetic indexes. These deposits may have sealed a Carboniferous marine failed rift (Vai 1991; Engelbrecht 2008).

To date, no fossils have been reported from this formation. According to its stratigraphic position at the transition between the marine Carboniferous and the middle Triassic terrestrial Verrucano Group, the Civitella Marittima Formation is probably Permian to early Triassic in age. This corresponds partly to the age attribution suggested by Costantini et al. (1991).

5.6 Verrucano Group (Middle Triassic)

5.6.1 Lower Verrucano Formation (Monte Quoio Formation) (Middle Triassic)

The formation, defined by Cocozza et al. (1975), consists of:

– Quartz conglomerate layers in alternation with pebbly, coarse-grained quartz sandstone and quartzite layers, which are massive or irregularly stratified and laminated. The strata reach thicknesses up to ca. 3.5 m. They thin laterally into violet quartz sandstones or are intercalated with erosive unconformities into the latter. The proximality trend of the quartz conglomerates corresponds to the present E–W direction: polymictic, poorly sorted, thick-bedded, laterally discontinuous lithologies of subzone I develop into better matured, polymictic to oligomictic lithologies of subzone II.

– Very thick, violet to bluish-grey-coloured, monotonous, silty, fine- to coarse-grained schistous quartz sandstones.

– Rarely grey, yellow-grey and reddish-brown, mm- and cm-sized boudinaged dolomitic caliche layers.

More details can be found in Costantini et al. (1988), Engelbrecht et al. (1988) and Engelbrecht (1997b). The erosive contact of the Lower Verruccano Formation to its stratigraphic base is either sharp or diffuse; the latter occurs within several metre thick oligomictic quartz conglomerate layers, in which the amount of reddish quartz pebbles gradually increases upward. The contact of the Lower Verrucano Formation with its stratigraphic cover is diffuse and locally hiatal.

The true thickness of the Lower Verrucano Formation cannot be given precisely because of intense deformation; it is tentatively estimated at max. 280 m.

To date, no in-situ fossils or traces of organic matter have been reported from the Lower Verrucano Formation. Its relative age is constrained by fossiliferous carbonate pebbles, which form accessory constituents in polygenic conglomerates. Cocozza et al. (1975) attributed its poorly preserved microfauna to the early and early middle Triassic. This view has been adopted in the stratigraphy (Fig. 4).

The sedimentary fabric points to regressive depositional conditions: ephemeral fluvial mass transport in a semiarid climate and proximal to intermediate fluvial deposits, consisting of coarse-grained channel fills and flood plain sediments, interstratified with caliches. Conspicuous redeposition from proximal sources occurred.

5.6.2 Upper Verrucano Formation (Ladinian)

This formation is seen partly as a lithological and temporal equivalent of the upper part of the Verrucano Group in the Pisan Mounts, e.g. of the ‘Anageniti minute’ (Costantini et al. 1988).

The formation consists of:

– Small-grained, clast-supported, oligomictic to monomictic, maximum 2.2 m thick quartz conglomerate layers, which are tabular or low-angle cross-laminated at the cm and dm scales.

– Whitish, grey-pink, light-greenish and reddish to violet-coloured, fine- to coarse-grained, locally pebbly, tabular and cross-bedded, laminated quartzites and quartz sandstones up to 40 cm thick.

– Bluish-grey, light grey, rarely greenish-grey and beige-coloured, up to several tens of metres thick arenaceous siltstones, containing thin lenses of micro-quartz conglomerate and occasionally cm-thick boudinaged dolomitic caliche layers.

The contact of the Upper Verrucano Formation with its transgressive stratigraphic cover is slightly unconformable and probably diachronous. Hiati exist in subzone II. Its true thickness, estimated at max. 300 m, cannot be given precisely because of intense deformation. To date, only ichnofauna consisting of cm-sized burrows has been identified.

According to the stratigraphic position of the Upper Verrucano Formation, a Ladinian age is inferred. It accommodates the passage between the regressive Lower Verrucano Formation and the marine-transgressive late Triassic evaporites.

The fabric of the sediment and its elevated maturity point to multiple recycling events in a distal fluvial and/or littoral-deltaic setting under semiarid climatic conditions.

Epizonal metamorphic unit

5.7 Tocchi Formation (Carnian)

This formation was first described by Signorini (1947). The base consists of homogeneous, pale, light greenish, very low-grade metamorphosed claystones and siltstones. Above follows an alternation of mudstones with mm- to cm-thick, tabular, light grey-yellowish, sometimes grey, cellular, intensely fissured dolostones and limestones. Further upwards, the carbonate layers thicken up to 1 m, whereas the mudstone interlayers diminish. The top of the formation consists of a breccia made of dark grey, mm- to cm-sized dolostone and rarely gypsum clasts, as well as oligomictic breccias consisting of varying amounts of unsorted angular dolostone, limestone and mudstone fragments with highly differing clast sizes dispersed in a limestone matrix.

Because of intense deformation, the true thickness of the Tocchi Formation cannot be calculated reliably; it is estimated at max. 200 m. Whether the local absence of this formation in subzone II is due to tectonic effects, non-deposition/erosion or substitution by Calcare cavernoso could not be clarified.

The age, depositional setting, palaeoclimate and tectonic, metamorphic and physico-chemical transformations of its subsurface equivalent (Burano Formation) were described in Costantini et al. (1980), in Martini et al. (1989) and in Barchi (2010).

The formation is interpreted as a marine-transgressive unit: an encroachment of the Mesozoic Tethys on the Adriatic passive continental margin, marking the onset of the Alpine orogenic cycle. It consisted of a first-order deepening-upward megasequence, which buried its pre-Carnian substratum and marked an epochal turning point from a predominantly siliciclastic to an evaporitic-carbonatic-silicic-marly depositional mode. A context with the Carnian pluvial event (Lu et al. 2021, and references therein) is suggested.

Tectonic unconformity
Tuscan nappe
Unmetamorphosed unit

5.8 Calcare cavernoso (Norian)

The formation is made up of massive, vacuolar carbonate breccias with highly variable amounts of powdery carbonate matrix. The breccia rarely contains xenoclasts; it interfingers with poorly lithified and unsorted, monomictic, clast-supported pseudo-conglomerates consisting of rounded rauhwacke clasts up to boulder size. More details can be found in Cornamusini et al. (2024), who interpreted the formation as a tectonic-chaotic unit that developed along a Miocene dilatational major décollement horizon. The thickness of the Calcare cavernoso, inextricably intermixed with cataclasites (see 5.10 below), is estimated at a maximum of 130 m. It is capped by either a pronounced erosive unconformity segregating it from postorogenic deposits of late Pliocene to Quaternary age ('serie Toscana ridotta') or the mentioned décollement, which separates it from the Ligurian units.

The depositional setting, age and tectonic as well as physico-chemical transformations of its subsurface equivalent, Carnian–Norian evaporites and dolostones (Burano Formation), are described in Martini et al. (1989) and in Gandin et al. (2000).

Tectonic unconformity
Allochthonous domain
Ligurian unit (Upper ophiolitic unit)

5.9 Galestri e Palombini (Cretaceous)

The unit consists of an unmetamorphosed tectonic mélange: flysch-type lithologies, composed of fractured and faulted, dark grey claystones and siltstones ('galestri') and bluish-grey, cherty limestones ('palombini') of Cretaceous age (Servizio Geologico d'Italia 2004). Its thickness is estimated to have a maximum of 80 m.

Tectonic unconformity
Allochthonous domain
Unmetamorphosed deposit

5.10 Late Miocene

Tectonised components (e.g. cataclasites, pseudo-conglomerates) of the Calcare cavernoso (see 5.8 above), with which they are intermixed inextricably. These reworked/tectonised/replaced derivates are cartographically not separable at a scale 1:10,000.

Erosive unconformity
Neoautochthonous domain
Unmetamorphosed deposit

5.11 Pliocene

There are grey and dark grey, maximum 35 m thick, sometimes reddish-brown altered marls and fine-grained, calcareous, friable, unconsolidated sandstones containing slightly redeposited faunas: lamellibranches (*Ostrea lamellosa*, *Pecten* sp.) gastropods (*Turritella* sp.), arthropods (*Balanus* sp.), echinoids and foraminifers (*Florilus* sp., *Cibicides* sp., *Textularia* sp.) (Signorini 1967; Klingsohr 1984).

5.12 Late Pliocene to early Quaternary

There are peneplains and fluvial terraces at different topographic elevations, covered with subrecent detritus and rarely with relics of fluvial gravel and boulders. These areas indicate the uplift stages and erosional states of individual tectonic compartments (see below).

Ignimbritic rhyolite flows (2.3–2.5 Ma) crop out in the westernmost part of the study area.

Relics of fluvial and lacustrine gravels and sands containing lithotypes of the Upper Verrucano Formation, the Ligurian units and the Neoautochthonous are present. These deposits that may be up to 20 m thick, are isolated from the recent fluvial drainage system and indicate that the latter forms an antecedent water gap.

5.13 Quaternary

The deposits consist of: subrecent talus and alluvium; fluvial gravel and sand forming high-, intermediate- and low-lying terraces; travertine and calcareous tufa; brownish-red, up to several metre thick lateritic soils; boulders; and landslide deposits.

6. Tectonics

The tectonic features described below are based on the measurements of ca. 3,800 planar and linear structures, which, together with stratigraphic data, serve to reconstruct the structural contours of geological boundaries and the orientation of β-axes (sensu Ramsay & Huber 1987).

Two tectono-stratigraphically homogeneous subzones separated by a segmented, steeply west-dipping, extensional first-order lineament (Arbia Lineament; see Electronic Supplement: geological map of the Farma River Valley) exist, which is offset by several transcurrent faults. This fault divides subzone I, hosting the Carboniferous inner shelf deposits in the west, from subzone II, containing coeval outer shelf–slope–basin margin sediments. The vertical throw at the subzone boundary is estimated at a maximum of 1.2 km.

Subzone I is characterised by an east-vergent, imbricated structure, where open frontal flexures and ramp folds, which attain amplitudes of up to ca. 200 m, are less frequent. The maximum thickness of the thrust sheets is estimated at 350 m. Asymmetric isoclinal, north–south trending anticlines of probable late Oligocene–early Miocene age are developed in competent lithologies (Fig. 7a, c), but the corresponding synclines are less frequently observed. The maximum thrust-induced stratigraphic throw appears to have transported the Risanguigno Formation onto the Lower Verrucano Formation. Because cut-off lines cannot be drawn, schistosities in coarse-grained sediments scatter and geopetal structures are rarely present, the thicknesses and along-strike persistences of overturned parts of the fold limbs remained undefined. The scattering of the bedding surface attitudes resulted from heterogeneous sediment lithologies, from synsedimentary, fault-controlled thickness fluctuations of the clastic formations and from post-Messinian NW–SE-directed compression. The latter probably produced open asymmetric fold trains with ß-axes dipping shallow to moderately W-SW and with fold amplitudes <3.0 m. N–S, NNE–SSW, NNW–SSE, NW–SE, WNW–ESE and W–E striking sets of normal faults and complex shear fractures of probable late Pliocene to Quaternary ages dissect the fold-thrust system. With few exceptions, the west–east striking fracture systems offset those striking north–south. A set of steeply east-dipping major normal faults (Fig. 7b) is supposed to form the western flank of the normal fault-bounded graben at ‘Il Belagaio’, which contains relics of marine deposits of Pliocene age, and to have dissected east-directed thrusts inferred from the presence of east-vergent folds (Fig. 7a) and from the sequence encountered in the ‘Sondaggio del Belagaio’ borehole. In addition, up to 5 cm sized tension gashes (joint 65/84) filled with cataclasite made up of quartz and quartzite fragments in the ankerite matrix are present. The normal fault immediately to the west of ‘Il Belagaio’, separating late Triassic evaporites from the Upper Verrucano Formation, constitutes a north–south striking, steeply east-dipping tectonic boundary, which was offset by several secondary WNW–ESE striking faults. The presence of silicified zones indicates that fluids ascended into units subjected to Pliocene-Quaternary extension.

Subzone II displays a significantly larger variation concerning the thrust directions: the fold axes reconstructed through the analysis of bedding surface measurements comprise domains with non-, east, SE, SSW and predominantly NE vergence of probable Oligocene–Pliocene age. The variability of vergence along the strike trends of ß-axes, as well as their curved trends, is more pronounced than in subzone I and may be caused by changes in the lithologies and thicknesses of the corresponding formations (Greenhalgh et al. 2015). The age relationships between these domains of different directions of thrust movements could not be determined. Open, tight and isoclinal anticlines and synclines – some of them probably developed at ramps – with predominantly shallow but also moderate and steep β-axes exist; the folds may reach amplitudes of up to 350 m. Overturned limbs are better developed than in subzone I. These reach thicknesses of up to 300 m and persist along strike up to 700 m. Thrust-related stratigraphic throw is highly variable: Middle Triassic but also Carboniferous formations were thrust onto late Triassic evaporites (Fig. 7d). Similar to subzone I, the tectonic superposition of Devonian onto middle Triassic strata marks the maximum thrust-induced throw. Small-scale boudin trains were observed that resulted from multilayer-normal shortening and related stretching in the fold limbs; e.g. accordingly deformed cm-thick dolostone nodule horizons contained in mudstones of the Carpineta Formation (Fig. 7e) probably represent drawn boudins (Goscombe et al. 2004). The mentioned domains of thrust directions are separated by normal faults and other complex faults of probable Pliocene to Quaternary ages, which can be grouped into sets displaying N–S, NNE–SSW, NE–SW, E–W, ESE–WSW and SE–NW directions. The age relationships of these fault sets cannot be determined, because mapping indicated that they intersect in a non-systematic fashion.

Both subzones are segmented by numerous steeply dipping normal faults, strike-slip faults and other faults of probable late Pliocene or Quaternary age into compartments (measuring 0.01–1.2 km^2), composed of stratigraphic subsequences; similar structures were identified and referred to by Signorini (1967) as ‘Struttura Toscana’. Thickness fluctuations of the Lower Verrucano and Civitella Marittima formations in adjoining compartments indicate that at least some of the fractures separating them may represent synsedimentary normal faults of Permian and/or middle Triassic age that were reactivated in the late Miocene to Quaternary. Each of these compartments is characterised by a specific style of tectonic deformation; this will be detailed in a forthcoming contribution. The highly variable orientations of the above-mentioned normal faults and shear fractures correspond largely with the Merse (N–S), Arbia (NNE–SSW), Farma (W–E) and Lanzo (NW–SE) directions of the MRMCC, which determine the regional morphotectonic-hydrographic pattern.

According to the mapping results, the relative age relationships between these groups of faults are difficult to establish, remain uncertain and cannot be summarised at present in a regular pattern. Possible explanations are the adverse outcrop conditions and the particularly complex tectonic situation: late Oligocene to Miocene east-directed thrusting was followed by polyphasic and variably oriented extensional and compressional events of Pliocene–Quaternary age. An attempt at unravelling and constraining the relative chronology of development of structures with differing geometries and trends is beyond the scope of this paper, and the problem will be addressed in a forthcoming contribution.

7. Discussion

7.1 Age and depositional conditions

Upper Verrucano Formation

The hiati in the Upper Verrucano Formation of subzone II may be explained by erosion, non-deposition or replacement with the Civitella Marittima Formation. However, due to a lack of evidence, the latter two interpretations were not adopted in the stratigraphic column (Fig. 4).

Lower Verrucano Formation

The hiati in the Lower Verrucano Formation of subzone II are interpreted as erosive gaps. The general fluvial transport directions (in recent coordinates) reported by Canuti & Sagri (1974) – westbound – and by Costantini et al. (1988) – southbound – in the MRMCC were not confirmed referring to the

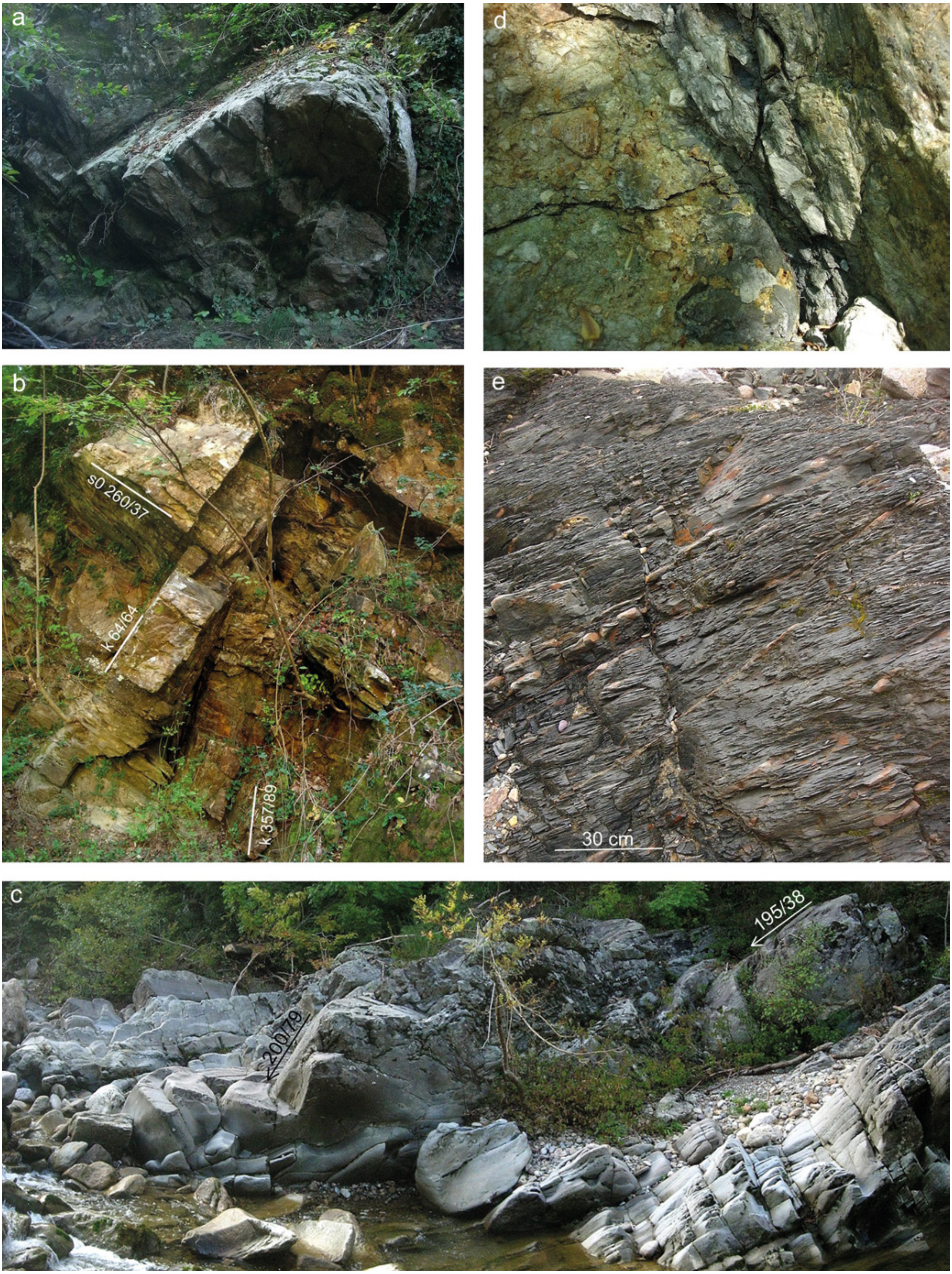

a
d
b
s0 260/37
k 64/64
k 357/89
e
30 cm
c
195/38
200/79

Fig. 7: Tectonics. (**a**) Tight east-vergent antiform developed in oligomictic, dm-thick, white-pink coloured microquartz conglomerates of the Upper Verrucano Formation, subzone I. β_0 174/22; flanks: s_0 221/40 normal; s_0 252/71 overturned. Loc.: 310 m ESE of the confluence of Fossato Piccolo with Farma River. View north. (**b**) dm- to m-thick layers of quartz sandstone, quartzite and small-grained quartz conglomerate of the Upper Verrucano Formation, s_0 260/37. En échelon joint set k 64/64 and fault k 357/89. The throw is indeterminable because of the lack of a marker horizon. Subzone I. Loc.: 60 m ESE of the confluence of Fossato Piccolo with the Farma River. View SE. (**c**) East-vergent isoclinic antiform in competent, coarse-grained quartz sandstones of the Poggio al Carpino Formation; flanks: s_0 252/46 normal; s_0 248/47 overturned; β_0 195/38, bent at a distance of 9 m southward to β_0 200/79. Subzone I. View NW. Loc.: 290 m NE of Ferriera di Ruota P 263. Bending of the β-axis may result from post-Messinian compression. (**d**) Thrust plane (k 213/46) separating schistous greywackes of the Farma Formation (s_0 268/65 overturned) from grey-yellow coloured carbonatic breccias of the Tocchi Formation (s_0 222/31). Subzone II. View NE. Loc.: Nameless ditch 200 m SW of 'Fungo Bianco' (P 157). Width of outcrop: 35 cm. View ENE. (**e**) Part of an overturned fold limb consisting of schistous mudstones (s_1 270/49) containing cm-thick boudin trains indicating stratification s_0 270/57. Sinistral, < 12 cm measuring offsets of the boudin trains at fault k 185/78. Subzone II. Scale: Hammer 28 cm. View west. Loc.: 110 m NE of P 190.

situation in the Farma River Valley: there the general fluvial transport course proceeds parallel to the general Carboniferous marine transport direction.

Considerations of the geological age of the Lower Verrucano Formation: newly sampled clasts of the same lithotype at the same location described in Cocozza et al. (1975) yielded better preserved microfossils consisting of fusulinid associations of Gzhelian to Cisuralian age (Engelbrecht et al. 1988). Because clasts containing Triassic fossils could not be ascertained by Engelbrecht (1997b) and the microfauna described in Cocozza et al. (1975) may also be Permian in age (E. Fluegel, Univ Erlangen-Nuremberg; pers. comm. 03.1983), the Lower Verrucano Formation might be, similar to the South Alpine Verrucano Formation (Cassinis et al. 2018), of Lopingian age. U-Pb age determinations of detrital zircons revealed 280 Ma (late Cisuralian) as the maximum depositional age of the Verruca Formation of nearby occurrences (Sirevaag et al. 2016; Paoli et al. 2016). The fusulinid-bearing carbonate pebbles ensure that locally the erosive base of the Lower Verrucano Formation reached down at least into Gzhelian open marine shelf deposits; a redeposition of these carbonate pebbles from the Civitella Marittima Formation is excluded according to its clast inventory (Engelbrecht 1997b: 75–76; Aldinucci et al. 2001, 2003). According to Pasini (1991) and Pasini & Vai (1997), these marine sediments, together with fusulinid-bearing deposits of Cisuralian age from Eastern Elba and of Guadalupian age from a drill site in the Monte Amiata geothermal area, indicate that relics of a marine Permian carbonate platform exist.

The Ladinian age of the Lower Verrucano Formation proposed by Cassinis et al. (2018) is called into question in this work according to the facts already mentioned and because no fossils sustaining this age attribution were described from the Lower Verrucano Formation and nearby equivalents.

Consequently the author favours a late Guadalupian–Lopingian age of the Lower Verrucano Formation, but adopted in this contribution the classic interpretation of its middle Triassic age.

Civitella Marittima Formation

It is considered by Costantini et al. (1988), Aldinucci et al. (2001, 2008b), Pandeli et al. (2004), Perrone et al. (2006: 21) and Cassinis et al. (2012) as part of the Verrucano Group. This is doubted here, because the typical sediment property of the Verrucano Group (terrestrial redbeds) is, except in the uppermost section of this formation, absent. Its interpretation as deposits of an early-middle Triassic failed rift (Cassinis et al. 2018: 491, Fig. 7) is called into question here because these sediments cover Carboniferous marine failed rift deposits.

Carpineta Formation

Engelbrecht (2002, 2008), Aldinucci et al. (2008a), Conti et al. (2020) and Spina et al. (2021) interpreted the stratigraphic position of this formation differently. This study revealed that the stratigraphic boundaries of the Carpineta Formation are conform with those of the coeval formations mentioned.

Spina et al. (2021) reported that the macrofossils found by Pasini in the Carpineta Formation were contained in its dolostone nodules. However, Pasini (1978b: 74, line 6) explicitly wrote that these were contained in "… dark-grey siltitic shales" (literal citation). A Permian age attribution to this formation, as proposed by several authors mentioned in paragraph 3, is doubted here (see paragraph 5.4).

Farma Formation

Pasini (1979b) concluded according to the relative uniformity of the age attributions (early–late Moscovian) of the fusulinids in the proximal part of the Farma Formation that their reworking occurred soon after their lithification, but not significantly later. The Guadalupian–Lopingian age of this formation postulated by several authors mentioned in paragraph 3 and the fact that a clear separation of the interturbidite layers from the mud-division of the turbidites was often not feasible (Aldinucci et al. 2008a) implies that redeposited microfossils with ages between the early Moscovian and Lopingian should be present in their fossil record. The fact that this is not the case is problematic. Aldinucci et al. (2008a) further reported that this formation was deposited with a hiatus above the late Pennsylvanian Spirifer Schists. However, according to our own observations, the latter are stratigraphically overlain by the Civitella Marittima Formation (Engelbrecht 2008), and the Farma Formation was deposited above the Risanguigno Formation. Therefore, the Guadalupian–Lopingian age proposed by above mentioned authors and others for the Farma Formation is questioned herein. The significantly different thermal alteration indexes of the sporomorphs discernible in Aldinucci et al. (2008a: fig. 8) are hardly compatible with the HP-LT regional metamorphism relevant to this formation. Thus, the proposed species attributions are doubted.

U-Pb detrital zircon ages from this formation revealed that its depositional age is late Mississippian to early Pennsylvanian and that it cannot be significantly younger than its maximum depositional age (Paoli et al. 2016). This is conform with the ages of the conodonts determined in the ultradistal subunit of the Farma Formation.

According to the tempestite-turbidite models of Walker (1984) and Aigner (1985) applied by Engelbrecht (2000), the microfloras are also subject to redeposition. Consequently, some of the 11 species of sporomorphs mentioned by Spina et al. (2001) from the inner shelf should also be present in the 17 species described by Aldinucci et al. (2008a) from the lower slope; however, none of the species described from the inner shelf correspond with those from the lower slope.

Aldinucci et al. (2008a) reported that the organic matter content in the Farma Formation is exclusively of terrestrial origin. However, they mention burrowed layers (ibid, p. 586, fig. 4, stratum no. 11). Moreover, marine microfaunas have been ascertained from the proximal part of this formation and coeval formations (see paragraph 3). Ichnofaunas have been documented from the proximal and intermediate parts of the Farma Formation (Engelbrecht 2019).

Poggio al Carpino Formation

The Guadalupian to early Triassic ages attributed to this formation (see paragraph 3) are called into question here because: (1) the stratigraphic contact displayed in Fig. 5a does not reveal distinctive marks that justify the interpretation of a hiatus of 85 Ma; (2) fossiliferous coeval formations of subzone II were classified via refined investigations as Carboniferous; (3) the black mudstones (containing local chert laminae) of this formation display close similarities with the sediments of its substratum. Their continuous deposition above the stratigraphic top of the Risanguigno Formation is inconsistent with the postulated pronounced hiatus.

Risanguigno Formation:

The early–middle Mississippian age of this formation, cited in paragraph 3 and the related statement that the late early Devonian microfauna described by Bagnoli & Tongiorgi (1979) is reworked, is disputed: To date, neither redeposited fossils of middle Devonian to early Mississippian age nor sedimentary structures typical of reworking have been mentioned from this formation. The palynomorphs in Capezzuoli et al. (2021: 10–11, 15–16) are described as "… strongly degraded …"; their systematics and specific sculptural characteristics were not provided. Some species attributions are doubted, e.g. the sculptures of the specimens termed *Spelaeotriletes balteatus* (Playford) Higgs 1975 and *S. pretiosus* (Playford) Neves and Belt 1970 in Capezzuoli et al. (2021: 11, fig. 7.9, 7.12) differ significantly from the referring species in Brittain & Higgs (2007: 112, 120–123).

7.2 Palaeoenvironmental setting

Further investigation is necessary to determine the subsidence history of the Risanguigno Basin and the reasons for the strongly reduced sediment input, the low species diversity and abundance, the suppression of carbonate production (otherwise widespread in the Devonian period) and the persistent dysoxic and anoxic conditions, which indicate a long-lasting palaeoecological crisis. Drivers might have been the effects of greenhouse gases emitted from distant large igneous provinces (LIPS) and from arc and rift volcanism, as well as the spread of vascular land plants, which caused enhanced silicate weathering and the eutrophication of marine-epicontinental areas (Kaiser et al. 2015; Racki 2020; Kabanov et al. 2023). Devonian LIPS were correlated with oceanic anoxia and biotic crises (Ernst et al. 2019). Crasquin & Horne (2018) proposed that oxygen minimum zones in pelagic areas (resulting from increased nutrient input) expand upward into shallow epicontinental seas. Alternatively, terrestrial nutrient runoff may have caused photic zone anoxia/dysoxia in shallow epicontinental basins and the spilling of eutrophicated water layers onto a well-ventilated deeper ocean water column (Carmichael et al. 2019).

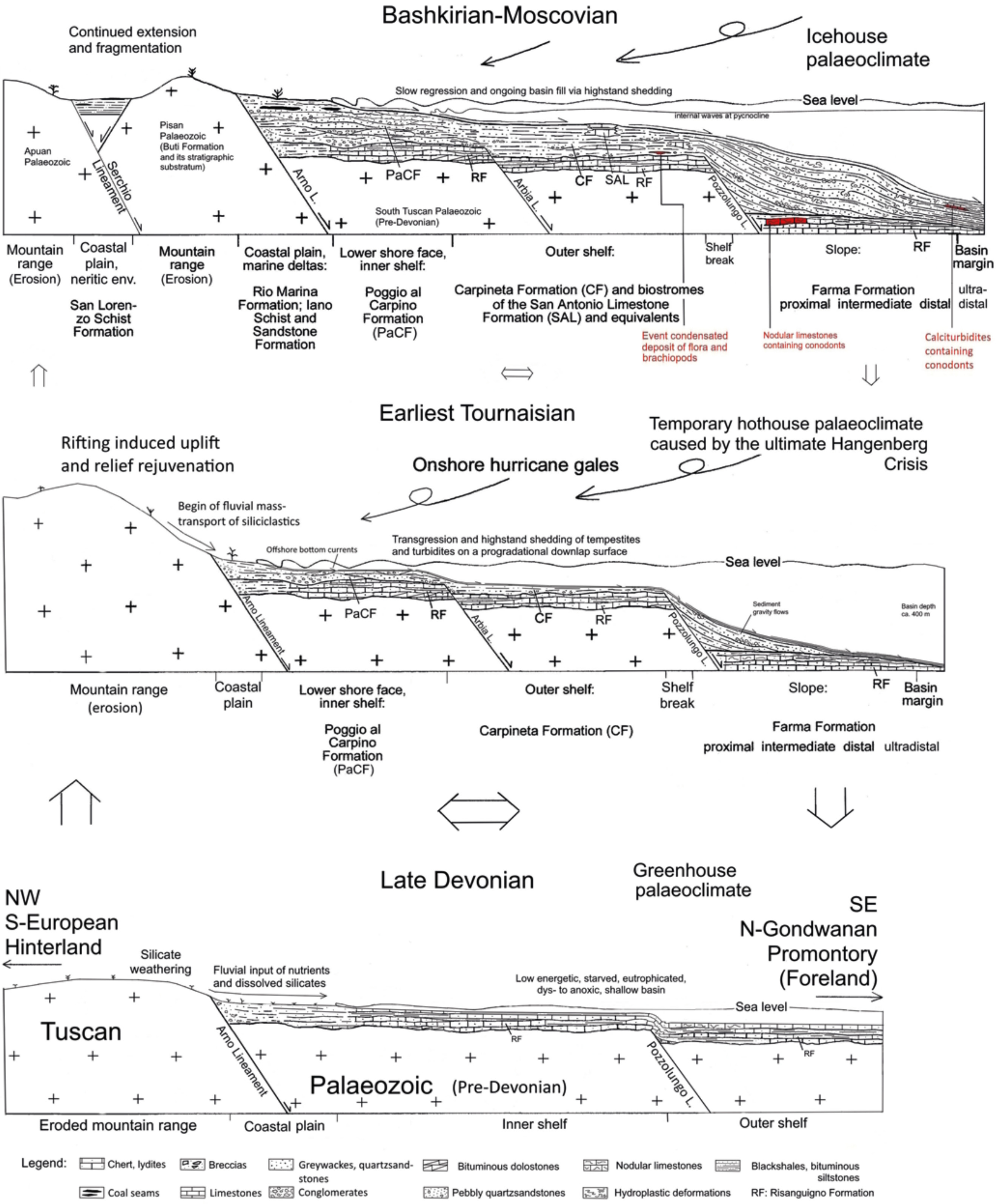
Proxies, attainable via systematic biochemical, geochemical and isotope analyses in the sediments of the Risanguigno Formation, could contribute to the understanding of how this palaeoenvironmentally critical and complex period of the Tuscan Late Palaeozoic developed and how it could be integrated into a global context (Pisarzowska et al. 2020; Kabanov et al. 2023).

The conspicuous change in the depositional mode at the stratigraphic top of the Risanguigno Formation can be explained by chronologically closely related events:

– A Tournaisian rift pulse, which deformed the basin by a staircase of normal faults with throws at the 100 m scale, occurred, thus creating a progradational downlap surface. Preexisting normal faults forming the Risanguigno Basin were probably reactivated (see e.g. the Pozzolungo Lineament postulated in Fig. 8).

– Palaeoenvironmental changes related to the Hangenberg Crisis triggered the sediment flux alimenting the prograding tempestite-turbidite system (Engelbrecht 2008).

Fig. 8: Schematic drafts – not to scale – of profiles of the Tuscan mainland, illustrating its late Devonian, earliest Tournaisian and Bashkirian-Moscovian palaeogeographic situations and depositional processes. Abbreviation: L. – Lineament. In red colour: approximate stratigraphic positions of new fossil findings. Modified and extended following Engelbrecht (2008: 300).


Bashkirian-Moscovian
Icehouse palaeoclimate
Continued extension and fragmentation
Slow regression and ongoing basin fill via highstand shedding
Sea level
internal waves at pycnocline
Apuan Palaeozoic
Serchio Lineament
Pisan Palaeozoic (Buti Formation and its stratigraphic substratum)
Arno L.
PaCF
RF
Arbia L.
CF
SAL
RF
Pozzolungo L.
South Tuscan Palaeozoic (Pre-Devonian)
Mountain range (Erosion)
Coastal plain, neritic env.
San Lorenzo Schist Formation
Mountain range (Erosion)
Coastal plain, marine deltas:
Rio Marina Formation; Iano Schist and Sandstone Formation
Lower shore face, inner shelf:
Poggio al Carpino Formation (PaCF)
Outer shelf:
Carpineta Formation (CF) and biostromes of the San Antonio Limestone Formation (SAL) and equivalents
Shelf break
Slope:
RF
Basin margin
Farma Formation
proximal intermediate distal
ultra-distal
Event condensated deposit of flora and brachiopods
Nodular limestones containing conodonts
Calciturbidites containing conodonts
⇑
⇔
⇓
Earliest Tournaisian
Rifting induced uplift and relief rejuvenation
Onshore hurricane gales
Temporary hothouse palaeoclimate caused by the ultimate Hangenberg Crisis
Begin of fluvial mass-transport of siliciclastics
Offshore bottom currents
Transgression and highstand shedding of tempestites and turbidites on a progradational downlap surface
Sea level
Sediment gravity flows
Basin depth ca. 400 m
Arno Lineament
PaCF
RF
Arbia L.
CF
RF
Pozzolungo L.
Mountain range (erosion)
Coastal plain
Lower shore face, inner shelf:
Poggio al Carpino Formation (PaCF)
Outer shelf:
Carpineta Formation (CF)
Shelf break
Slope:
RF
Basin margin
Farma Formation
proximal intermediate distal ultradistal
⇑
⇔
⇓
Late Devonian
Greenhouse palaeoclimate
NW
S-European Hinterland
SE
N-Gondwanan Promontory (Foreland)
Silicate weathering
Fluvial input of nutrients and dissolved silicates
Low energetic, starved, eutrophicated, dys- to anoxic, shallow basin
Sea level
Tuscan
Arno Lineament
RF
Palaeozoic (Pre-Devonian)
Pozzolungo L.
RF
Eroded mountain range
Coastal plain
Inner shelf
Outer shelf
Legend:
Chert, lydites
Breccias
Greywackes, quartzsandstones
Bituminous dolostones
Nodular limestones
Blackshales, bituminous siltstones
Coal seams
Limestones
Conglomerates
Pebbly quartzsandstones
Hydroplastic deformations
RF: Risanguigno Formation

Aretz & Corradini (2021) reported that the Hangenberg Crisis occurred suddenly and disruptively and caused major perturbations in the geosphere. According to Kidder & Worsley (2010) and Kabanov et al. (2023), additional emissions of large amounts of volcanic CO_2 could have resulted in global overheating and a transition from greenhouse to temporary hothouse conditions, which enhanced hydrological cycling, thus boosting the terrigenous flux seaward. This mechanism (which probably caused the epochal change in the depositional mode) might have initiated the onset of that mass transport of siliciclastics, supplied from an eroding, rejuvenated relief onshore. Oxygen-deficient zones expanded and shallowed during hothouse conditions into the photic zone and its resulting contamination with euxinia (Kabanov et al. 2023) probably prevented the establishment of faunas in the inner shelf area. Elevated solar UVB radiation probably contributed during this phase to the degradation of photic zone biota (Marshall et al. 2020; Häder et al. 2015). The mentioned depositional style stabilised over the Carboniferous period, probably because subsequent increasing icehouse-climate conditions generated elevated equator-to-pole atmospheric temperature gradients, which enabled more frequent tropical tempestite events in the narrowing, W–E-trending Palaeotethys; the strike of its coastal lines favoured the physical conditions necessary for tropical cyclone generation (Duke 1985).

Further palaeoecological considerations concerning younger formations are beyond the scope of this paper and will be presented in a forthcoming contribution.

The late Emsian to late Ladinian palaeoecological conditions in the Tuscan part of the Adria microplate were adverse. The reasons for this should be investigated with greater detail.

7.3 Palaeotectonic development

The lithostratigraphic boundaries of the Carboniferous coeval formations are strongly favoured to be erosive unconformities. A superposition of Hercynian and Oligocene–early Miocene compressive tectogeneses (Costantini et al. 1988; Conti et al. 1991; Aldinucci et al. 2008c: 576–577; Molli et al. 2020: 05, fig. 2b and others) was not detectable. Accordingly, the Farma Basin represents neither a Permian transtensional basin (Brogi et al. 2023) nor a Carboniferous compressive foreland basin (Cocozza et al. 1987 and others). No relics of Early Palaeozoic or Late Precambrian nappes thrust onto Carboniferous deposits were identified. The Farma Basin is positioned in a Carboniferous normal fault-bounded marine-epicontinental extensional setting (Fig. 8). The implied tectonic structures (normal faults) are inferred from stratigraphic considerations. This basin formed at the southern Hercynian convergence front near the Palaeoadriatic suture at the thickened accretionary wedge, whose upper part disintegrated due to gravitational collapse, thus generating short-lived rift basins, as proposed by Vai (2001) and Stampfli (2005). The lack of the Bouma gradation in basal layers of the Farma Formation, as can be seen in Fig. 5b, may be caused therein by the fact that during the early stage of basin subsidence the physical requirements for deposition out of suspension were not yet given.

The partial inversion of the Farma Basin deposits started in the Lopingian and was caused by the combined effects of the rise of geotherms driven by nearby magmatism (Bagnoli et al. 1979; Costantini et al. 1991; Pandeli et al. 2004; Pieruccioni et al. 2020; Marini et al. 2020) and by the global sea level fall (Van der Meer et al. 2025). Consequently, the upper parts of the units (down to the latest Pennsylvanian deposits) were exposed to fluvial erosion due to surface emergence.

The postulated shear-zone-bound transtensional Guadalupian-Lopingian basins (having hosted the aforementioned heteropic formations) as well as their Oligocene–Miocene tectonic deformation, as described in Molli et al. (2020), are doubted.

Subzones I and II, characterised by specific Carboniferous lithostratigraphic successions and specific Palaeogene–Neogene deformation styles, constitute the working area. Therefore, the overall tectonic structure may consist in closely spaced, interfering metamorphic core complexes, as defined by Tirel et al. (2009) and Zuza & Cao (2023). The preconditions for this deformation style (e.g. high geothermal heat flow, extended and very thin crust) were already listed. The components of this part of the MRMCC have several shear zone systems in common that are bound to zones of crustal weakness, i.e. the Miocene decollement horizon developed in late Triassic evaporites, and reactivated Carboniferous normal faults and Devonian mudstones.

The postulated shallow west-dipping thrust proposed by Bodechtel (1969), Costantini et al. (1988), Conti et al. (1991), Liotta (2002), Lazzarotto et al. (2003), Capezzuoli (2021) and others, separating the subzones and the implied v-shaped bearing of the north–south striking thrust surface supposed to be located to the west of 'Il Belagaio' could not be identified, although the thrusting of strata of the Upper Verrucano Formation onto Calcare cavernoso was ascertained in the drill core at a depth of 75.80 m in the 'Sondaggio del Belagaio' borehole (Lazzarotto & Moretti 1973) and by several east-vergent folds (e.g. Fig. 7a) and east directed thrusts observed close to the drill site, confirming Oligocene–early Miocene tectonic transport eastward. The above proposed Oligocene–early Miocene thrust surface may have been dissected by late Miocene to Quaternary tear fractures and normal faults (Fig. 7b), which support the inference that the thrust surface was dissected and displaced (see profile 1-1' in the geological map of the Farma River Valley; Electronic Supplement).

The different tectono-metamorphic histories of both subzones, as well as their relationships with subzone III, are not addressed herein, and will be the object of a forthcoming contribution.

The morphotectonic feature of the Farma River Valley present between the extinct rhyolites of Torniella-Roccastrada (2.2–2.5 Ma) and the active hydrothermal system Bagno di Petriolo is conspicuous, mainly because the fault system forming that valley displays evidence pointing to a buried magmatic body emitting heat and fluids: In the central and eastern parts of the valley, travertine deposits, cold and warm (< 30 °C) springs precipitating calcareous tufa, travertine and ferric hydroxide, H_2S and CO_2 vents, hydrothermal quartz precipitates, intensely altered/bleached zones and rarely faint stibnite, cinnabar and pyrite mineralisations occur; many of these manifestations are controlled by faults that acted as pathways for the Pliocene to recent mineralised fluids and gases, supplied by an active magmatic reservoir (Montanari et al. 2023). Therefore, the Farma River Valley probably represents an exhumed system of steeply inclined tear faults that connects the already cooled magmatic body, which has fed the rhyolites of Torniella-Roccastrada at its western margin, with an active magmatic reservoir below Bagno di Petriolo at its eastern border. The onset of the activity of this geothermal system should be determined by dating the age of the basal travertine layer deposited there. Tear structures probably exerted control on the emplacement of the eastward-younging magmatic bodies of the Tuscan Magmatic Province (M. Lupi, University of Geneva, written communication; Nicollet et al. 2023).

8. Conclusions

The main results from the geological survey in the Farma River Valley are as follows:

– The Farma River Valley exposes a key section across the closely spaced interfering metamorphic core complexes of Monticiano-Roccastrada of the inner zone of the Northern Apennines.

– The deformation style, which is typical of the inner zone of the Northern Apennines, was further complicated by post-Messinian compressive phases, resulting in belt-parallel shortening and vigorous oroclinal bending of the southernmost quarter of the Mid-Tuscan Ridge.

– The observations support the ideas of an Emsian–early Tournaisian age of the stratigraphic base of the MRMCC and a Carboniferous age of the coeval formations following above with erosive unconformity. A final answer will be found in further palaeontological studies.

– Highstand shedding onto a normal-fault-fragmentated downlap surface of a uniform substratum started abruptly in the earliest Tournaisian by the combined effects of rifting-induced relief rejuvenation and of enhanced hydrological cycling caused by the Hangenberg Crisis. These chronologically closely related events provoked the progradation of predominantly siliciclastic tempestite-turbidite deposits, which covered the inner and outer shelf, slope and basin margin. Subsequently, this depositional mode prevailed until the late Pennsylvanian, probably because of increasing icehouse climate conditions.

– Regressive, littoral to terrestrial siliciclastics of Permian to middle Triassic age cover the marine failed rift deposits. At the stratigraphic top of the latter, patches of Fusulina limestone were interstratified. Partial basin inversion caused their reworking into the Verrucano Group.

– Zones of crustal weakness (Devonian carbonaceous mudstones, Carboniferous normal faults and late Triassic evaporites) controlled during the late Palaeogene the origin of detachments, which enabled thrusting and tectonic burial of the units and during the Neogene their dilatational exhumation from a depth of ca. 25 km. The reactivated Carboniferous normal faults confine tectono-stratigraphically homogeneous subzones: The latter are characterised by specific lithology and deformation, which form the central part of the closely spaced interfering metamorphic core complexes of Monticiano-Roccastrada.

– The conspicuous geomorphology of the east–west-directed Farma River Valley probably traces an exhumed system of tear faults that connects the cooled magmatic body below the rhyolites of Torniella-Roccastrada (2.3–2.5 Ma) at its western margin with an active magmatic reservoir feeding the hydrothermal system some kilometres below Bagno di Petriolo at its eastern boundary.

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10. References

Aigner, T. (1985). Storm depositional systems: Dynamic stratigraphy in modern and ancient shallow-marine sequences. *Lecture Notes in Earth Sciences, 3*, 1–174. https://doi.org/10.1007/BFb0011411

Aldinucci, M., Gandin, A., Pandeli, E., & Sandrelli, F. (2001). Sedimentary features, paleoenvironmental data and stratigraphic meaning of the Civitella Formation/Verrucano Group, Southern Tuscany, Italy). In Dipartimento di Scienze della Terra, Siena (Ed.), *Stratigraphic and structural evolution of the Late Carboniferous to Triassic continental and marine successions in*

Tuscany (Italy), regional reports and general correlation (pp. 12–14). Siena.

Aldinucci, A., Sandrelli, F., & Pandeli, E. (2003). Sedimentary features, depositional setting and stratigraphic meaning of Civitella Marittima Formation (Verrucano Group, southern Tuscany, Italy). *Bollettino della Società Geologica Italiana, 1*(Spec.), 37–48.

Aldinucci, M., Brogi, A., & Spina, A. (2008a). Middle-Late Permian sporomorphs from the Farma Formation (Monticiano-Roccastrada Ridge, southern Tuscany): New constraints for the tectono-sedimentary history of the Tuscan Domain. *Bollettino della Società Geologica Italiana, 127*(3), 581–597.

Aldinucci, A., Gandin, A., & Sandrelli, F. (2008b). The Mesozoic continental rifting in the Mediterranean area: Insights from the Verrucano tectofacies of southern Tuscany (Northern Apennines, Italy*). International Journal of Earth Sciences, 97*(6), 1247–1269. https://doi.org/10.1007/s00531-007-0208-9

Aldinucci, M., Pandeli, E., & Sandrelli, F. (2008c). Tectono-sedimentary evolution of the late Palaeozoic-early Mesozoic metasediments of the Monticiano-Roccastrada Ridge (southern Tuscany, Northern Apennines, Italy). *Italian Journal of Geosciences, 127*(3), 567–579.

Aretz, M., & Corradini, C. (2021). Global review of the Devonian-Carboniferous boundary: An introduction. *Palaeobiodiversity and Palaeoenvironments, 101*(2), 285–293. https://doi.org/10.1007/s12549-021-00499-8

Azzaro, E., Cocozza, T., Di Sabatino, B., Gasperi, G., Gelmini, R., & Lazzarotto, A. (1976). Geology and petrography of the Verrucano and Paleozoic formations of Southern Tuscany and Northern Latium (Italy). In H. Falke (Ed.), *The continental Permian in Central, West and Southern Europe* (pp. 181–195). Dordrecht: Reidel Publ. Comp. https://doi.org/10.1007/978-94-010-1461-8_15

Bagnoli, G., & Tongiorgi, M. (1979). New fossiliferous Silurian (Mt. Corchia) and Devonian (Monticiano) layers in the Tuscan Paleozoic. *Memorie della Società Geologica Italiana, 20*, 299–311.

Bagnoli, G., Gianelli, G., Puxeddu, M., Rau, A., Squarci, P., & Tongiorgi, M. (1979). A tentative stratigraphic reconstruction of the Tuscan Paleozoic Basement. *Memorie della Società Geologica Italiana, 20*, 99–116.

Barchi, M. (2010). The Neogene-Quaternary evolution of the Northern Apennines: crustal structure, style of deformation and seismicity. In M. Beltrando, A. Peccerillo, M. Mattei, S. Conticelli, & C. Doglioni (Eds.), *The geology of Italy: tectonics and life along plate margins. Journal of the Virtual Explorer, Electronic Edition, 36(11)*. https://doi.org/10.3809/jvirtex.2010.00220

Bartoletti, E., & Tongiorgi, M. (1982). Ricerche sulla geologia dei dintorni di Monticiano (Siena). In CNR – Consiglio Nazionale Ricerche (Ed.), *Energia geotermica: prospettive aperte dalle ricerche del Consiglio Nazionale delle Ricerche. Progetto Finalizzato Energetica, Sottoprogetto Energia Geotermica* (pp. 85–98). Rome: PEG Editrice.

Bellani, S. (2018). Update of heat flow data in the geothermal areas of Tuscany, Italy. *Geothermal Resources Council – Transactions, 42*(1), 861–870. https://publications.cnr.it/doc/396855

Bertini, G., Elter, F. M., & Talarico, F. (1994). Evidenze di una fase estensionale pre-triassica nel complesso degli gneiss nell'area geotermica di Larderello (Toscana Meridionale). *Studi Geologici Camerti, 1*, 129–137. http://193.204.8.201:8080/jspui/handle/1336/510

Bianco, C., Brogi, A., Caggianelli, A., Giorgetti, G., Liotta, D., & Meccheri, M. (2015). HP-LT metamorphism in Elba Island: Implications for the geodynamic evolution of the inner Northern Apennines (Italy). *Journal of Geodynamics, 91*, 13–25. https://doi.org/10.1016/j.jog.2015.08.001

Bianco, C., Godard, G., Halton, A. M., Brogi, A., Liotta, D., & Caggianelli, A. (2019). The lawsonite-glaucophane blueschists of Elba Island (Italy). *Lithos, 348–349*, 105198. https://doi.org/10.1016/j.lithos.2019.105198

Boccaletti, M., Coli, M., Decandia, F. A., Giannini, E., & Lazzarotto, A. (1980). Evoluzione dell'Appennino settentrionale secondo un nuovo modello strutturale. *Memorie della Società Geologica Italiana, 21*, 359–373.

Boccaletti, M., Decandia, F. A., Gasperi, G., Gelmini, R., Lazzarotto, A., & Zanzucchi, G. (1987). *Carta strutturale dell'Appennino Settentrionale, scala 1: 250.000. Note illustrative*. CNR, Sottoprogetto 5, Publicazione n. 429–1982. Tipografia Senese.

Bodechtel, J. F. (1969). Photogeologische Untersuchungen in der südlichen Toskana und Umbrien. *Bayerische Akademie der Wissenschaften, mathematisch-naturwissenschaftliche Klasse – Abhandlungen, 136*, 1–59.

Bonini, M., Sani, F., Stucchi, E. M., Moratti, G., Benvenuti, M., Menanno, G., & Tanini, C. (2014). Late Miocene shortening of the Northern Apennines back-arc. *Journal of Geodynamics, 74*, 1–31. https://doi.org/10.1016/j.jog.2013.11.002

Borsi, S., Ferrara, G., Rau, A., & Tongiorgi, E. (1967a). Determinazione con il metodo Rb/Sr dell'età delle Filladi e Quarziti listate di Buti (Monti Pisani). *Atti Società Toscana Scienze Naturali, A 73*, 632–646.

Borsi, S., Ferrara, G., & Tongiorgi, E. (1967b). Determinazione con il metodo K/Ar dell'età delle rocce magmatiche della Toscana. *Bollettino della Società Geologica Italiana, 86*, 403–410.

Branson, E. B., & Mehl, M. G. (1933). Conodonts from the Bainbridge (Silurian) of Missouri. *University of Missouri Studies, 8*, 39–52.

Brittain, J. M., & Higgs, K. T. (2007). The early Carboniferous *Spelaeotriletes balteatus – S. pretiosus* Miospore Complex: Defining the base of the *Spelaeotriletes pretiosus – Raistrickia clavata* (PC) Miospore Biozone. *Comunicações Geológicas, 94*, 109–123. https://www.lneg.pt/wp-content/uploads/2020/03/ComunGeol_V94_N1_Article_6.pdf

Brogi, A. (2008). Kinematics and geometry of Miocene low-angle detachments and exhumation of the metamorphic units in the hinterland of the Northern Apennines (Italy). *Journal of Structural Geology, 30*(1), 2–20. https://doi.org/10.1016/j.jsg.2007.09.012

Brogi, A., & Giorgetti, G. (2012). Tectono-metamorphic evolution of the siliciclastic units in the Middle Tuscan Range (inner Northern Apennines): Mg-carpholite bearing quartz veins related to syn-metamorphic syn-orogenic foliation. *Tectonophysics, 526–529*, 167–184. https://doi.org/10.1016/j.tecto.2011.09.015

Brogi, A., Lazzarotto, A., & Liotta, D., & the CROP Working Group. (2005). Structural features of southern Tuscany and geological interpretation of the CROP 18 Seismic Reflection Survey (Italy). *Bollettino della Società Geologica Italiana, 3*, 213–236.

Brogi, A., Capezzuoli, E., Martini, I., Picozzi, M., & Sandrelli, F. (2014). Late Quaternary tectonics in the inner Northern Apennines (Siena Basin, southern Tuscany, Italy) and their seismotectonic implication. *Journal of Geodynamics, 76*, 25–45. https://doi.org/10.1016/j.jog.2014.03.001

Brogi, A., Regoli, R., Spina, A., Capezzuoli, E., Zucchi, M., Lucci, F., . . . Cirilli, S. (2023). The Permian-Triassic succession of the Montagnola Senese Ridge (Middle Tuscan Ridge, Italy): A perspective for late Palaeozoic magmatism and continentalisation in the western Tethys. *International Geology Review, 66*(4), 897–926. https://doi.org/10.1080/00206814.2023.2220011

Brunet, C., Monié, P., Jolivet, L., & Cadet, J.-P. (2000). Migration of compression and extension in the Tyrrhenian Sea, insights from ^{40}Ar/^{39}Ar ages on micas along a transect from Corsica to Tuscany. *Tectonophysics, 321*(1), 127–155. https://doi.org/10.1016/S0040-1951(00)00067-6

Canuti, P., & Sagri, M. (1974). Ambiente di sedimentazione e provenienza di clasti nelle Anageniti del Verrucano (Appennino Settentrionale). *Bollettino della Società Geologica Italiana, 93*, 661–704.

Capezzuoli, E., & Gandin, A. (2005). *Facies distribution and microfacies of thermal-spring travertine from Tuscany*. Proceedings of 1st International Symposium on Travertine, 21–25/09/2005 (pp. 43–50), Pamukkale University, Denizli, Turkey.

Capezzuoli, E., Spina, A., Brogi, A., Liotta, D., Bagnoli, G., Zucchi, M., . . . Regoli, R. (2021). Reconsidering the Variscan Basement of Southern Tuscany (Inner Northern Apennines). *Geosciences, 11*(2), 84. https://doi.org/10.3390/geosciences11020084

Carmichael, S. K., Waters, J. A., Königshof, P., Suttner, T. J., & Kido, E. (2019). Paleogeography and paleoenvironments of the Late Devonian Kellwasser Event: A review of its sedimentological and geochemical expression. *Global and Planetary Change, 183*, 102984. https://doi.org/10.1016/j.gloplacha.2019.102984

Carmignani, L., Decandia, F. A., Disperati, L., Fantozzi, P. L., Lazzarotto, A., Liotta, D., & Tavarnelli, E. (1996). Relazioni tra il Bacino Balearico, il Tirreno Settentrionale e l'evoluzione neogenica dell'Appennino Settentrionale. *Studi Geologici Camerti. Volume Speciale, 1995*(1), 255–268.

Carmignani, L., Decandia, F. A., Disperati, L., Fantozzi, P. L., Kligfield, R., Lazzarotto, A., . . . Meccheri, M. (2001). Inner Northern Apennines. In G. B. Vai & I. P. Martini (Eds.), *Anatomy of an orogen: the Apennines and adjacent Mediterranean basins* (pp. 197–213). Dordrecht: Springer. https://doi.org/10.1007/978-94-015-9829-3_14

Casini, G., Decandia, F. A., & Tavarnelli, E. (2007). Analysis of a mesoscopic duplex in SW Tuscany, Italy: Implications for thrust system development during positive tectonic inversion. *Special Publication – Geological Society of London, 272*(1), 437–446. https://doi.org/10.1144/GSL.SP.2007.272.01.22

Cassinis, G., Perotti, C. R., & Ronchi, A. (2012). Permian continental basins in the Southern Alps (Italy) and peri-Mediterranean correlations. *International Journal of Earth Sciences, 101*(1), 129–157. https://doi.org/10.1007/s00531-011-0642-6

Cassinis, G., Perotti, C., & Santi, G. (2018). Post-Variscan Verrucano-like deposits in Italy, and the onset of the alpine tectono-sedimentary cycle. *Earth-Science Reviews, 185*, 476–497. https://doi.org/10.1016/j.earscirev.2018.06.021

Cocozza, T. (1965). Il Carbonifero nel Gruppo Monticiano-Roccastrada. *Rendiconti, 8*(3), 3–29.

Cocozza, T., Lazzarotto, A., & Vai, G.-B. (1974). Flysch e molassa ercinici del Torrente Farma (Toscana). *Bollettino della Società Geologica Italiana, 93*, 115–128.

Cocozza, T., Lazzarotto, A., & Pasini, M. (1975). Segnalazione di una fauna triassica nel conglomerato di Monte Quoio (Verrucano del Torrente Farma – Toscana Meridionale). *Rivista Italiana di Paleontologia e Stratigrafia, 81*(5), 425–436.

Cocozza, T., Costantini, A., Lazzarotto, A., & Sandrelli, F. (1978). Continental Permian in Southern Tuscany. In M. Tongiorgi (Ed.), *Report on the Tuscan Paleozoic basement* (pp. 35–49). Pisa: CNR.

Cocozza, T., Decandia, F. A., Pasini, M., & Vai, G.-B. (1987). The marine Carboniferous sequence in Southern Tuscany: its bearing for Hercynian paleogeography and tectofacies. In H. W. Flügel, F. P. Sassi, & P. Grecula (Eds.), *Pre-Variscan and Variscan events in the Alpine-Mediterranean mountain belts.* IGCP Project No 5, Mineralia Slovaca – Monography (pp. 135–144). Bratislava: Alfa Publ.

Cohen, K. M. and Car, N. (2025): The ICS international chronostratigraphic chart this decade. Episodes, 2025. https://doi.org/10.18814/epiiugs/2025/025001

Compton, R. R. (2017). *Geology in the field*. Earthspun Books.

Conti, P., Costantini, A., Decandia, F. A., Elter, F. M., Gattiglio, M., Lazzarotto, A., . . . Di Pisa, A. (1991). Structural frame of the Tuscan Palaeozoic: A review. *Bollettino della Società Geologica Italiana, 110*, 523–541. https://www.pconti.net/doc/Conti1991.pdf

Conti, P., Cornamusini, G., & Carmignani, L. (2020). An outline of the geology of the Northern Apennines (Italy), with geological map at 1:250.000 scale. *Italian Journal of Geosciences, 139*(2), 149–194. https://doi.org/10.3301/IJG.2019.25

Cornamusini, G., Milaneschi, L., Conti, P., & Liberato, G. P. (2024). Significance of the Calcare Cavernoso: Stratigraphic-structural setting and role as tectonic chaotic unit in the evolution of the Northern Apennines (Italy). *Italian Journal of Geosciences, 143*(2), 1–209. https://doi.org/10.3301/IJG.2024.17

Cosentino, D., Cipollari, P., Marsili, P., & Scrocca, D. (2010). Geology of the central Apennines: A regional review. *Journal of the Virtual Explorer. Electronic Edition, 36*(12). https://doi.org/10.3809/jvirtex.2010.00223

Costantini, A., Gandin, A., Mattias, P. P., Sandrelli, F., & Turi, B. (1980). Un'ipotesi per l'interpretazione paleogeografica della formazione di Tocchi. *Memorie della Società Geologica Italiana, 21*, 203–216.

Costantini, A., Decandia, F. A., Lazzarotto, A., & Sandrelli, F. (1988). L'Unità di Monticiano-Roccastrada fra la Montagnola Senese e il Monte Leoni (Toscana meridionale). *Atti Ticinensi di Scienze della Terra, 31*, 382–420.

Costantini, A., Decandia, F. A., Lazzarotto, A., & Memmi, I. (1991). Evidence of Triassic volcanic activity in the Verrucano Group of Southern Tuscany. *Bollettino della Società Geologica Italiana, 110*, 717–725.

Crasquin, S., & Horne, D. J. (2018). The palaeopsychrosphere in the Devonian. *Lethaia, 51*(4), 547–563. https://doi.org/10.1111/let.12277

Decandia, F. A., Lazzarotto, A., Pandeli, E., Sandrelli, F., & Aldinucci, M. (2004). *The Middle Carboniferous Variscan flysch and post-Variscan siliciclastics.* 32nd International Geological Congress, Florence, Italy, August 20–28, 2004; Field trip guide book – B05. http://www.apuanegeopark.it/documents/Paper_Basement_Paleozoic_Pandeli.pdf

De Franco, R., Petracchini, L., Scrocca, D., Caielli, G., Montegrossi, G., Santilano, A., & Manzella, A. (2019). Synthetic seismic reflection modelling in a supercritical geothermal system: An image of the K-horizon in the Larderello Field (Italy). *Geofluids, 2019*, 1–21. https://doi.org/10.1155/2019/8492453

Dehm, R. M., Klemm, D.-D., Mueller, C., Wagner, J., & Weber-Diefenbach, K. (1983). Exploration for antimony deposits in

Southern Tuscany, Italy. *Mineralium Deposita, 18*, 423–434. https://doi.org/10.1007/BF00206490

Del Moro, A., Puxeddu, M., di Brozolo, F. R, & Villa, I. M. (1982). Rb-Sr and K-Ar ages on minerals at temperatures of 300–400°C from deep wells in the Larderello geothermal field, Italy. *Contributions to Mineralogy and Petrology, 81*(4), 340–349. https://doi.org/10.1007/BF00371688

Di Sabatino, B., Negretti, G., & Sagri, F. P. (1978). T-P conditions during the Hercynian metamorphism in Tuscany and Latium, and comparison with other regions in the Alpine-Mediterranean area. Continental Permian in Southern Tuscany. In M. Tongiorgi (Ed.), *Report on the Tuscan Paleozoic basement* (pp. 59–66). Pisa: CNR.

Doglioni, C., Moretti, I., & Roure, F. (1991). Basal lithospheric detachment, eastward mantle flow and mediterranean geodynamics: A discussion. *Journal of Geodynamics, 13*(1), 47–65. https://doi.org/10.1016/0264-3707(91)90029-E

Duke, W. L. (1985). Hummocky cross-stratification, tropical hurricanes, and intense winter storms. *Sedimentology, 32*(2), 167–194. https://doi.org/10.1111/j.1365-3091.1985.tb00502.x

Engelbrecht, H. (1997a). From Upper Palaeozoic extensional basin fill to late Alpine low grade metamorphic core complex: Preliminary note on the sedimentary and tectonic development of the Monticiano-Roccastrada-Zone (MRZ; Southern Tuscany, Italy). *Zeitschrift der Deutschen Geologischen Gesellschaft, 148*(3–4), 523–546. https://doi.org/10.1127/zdgg/148/1997/523

Engelbrecht, H. (1997b). *Zur Geologie der Zone von Monticiano-Roccastrada*. Dissertation, LMU Munich.

Engelbrecht, H. (2000). Deposition of tempestites in the eastern Rheic strait: Evidence from the Upper Palaeozoic of Southern Tuscany (Italy). *Facies, 43*, 103–121. https://doi.org/10.1007/BF02536986

Engelbrecht, H. (2002). Notes on the lithofacies and origin of the sediment gravity flows of the Carboniferous Farma Formation (Monticiano-Roccastrada Zone, Southern Tuscany, Italy). *Zentralblatt für Geologie und Paläontologie, Teil 1, 2001*(3–4), 335–347.

Engelbrecht, H. (2008). Carboniferous continental margin deposits in Southern Tuscany, Italy: Results from geological mapping of the geotopes Farma Valley and Sant' Antonio Mine area. *Geological Journal, 43*(2–3), 279–305. https://doi.org/10.1002/gj.1106

Engelbrecht, H. (2019). Revision of geological units of the Carboniferous Farma Basin at the Southern Convergence Front of the West Mediterranean Hercynides. *Kölner Forum für Geologie und Paläontologie, 23*, 89–90.

Engelbrecht, H., Klemm, D.-D., & Pasini, M. (1988). Preliminary notes on the tectonics and lithotypes of the "Verrucano s.l." in the Monticiano area (Southern Tuscany, Italy) and the finding of fusulinids within the M.te Quoio Fm. (Verrucano Group). *Rivista Italiana di Paleontologia e Stratigrafia, 94*(3), 361–382. https://doi.org/10.54103/2039-4942/13147

Ernst, R. E., Rodygin, S., & Grinev, O. M. (2019). Age correlation of large igneous provinces with Devonian biotic crises. *Global and Planetary Change, 185*(7), 103097. https://doi.org/10.1016/j.gloplacha.2019.103097

Festa, A., Dilek, Y., Pini, G. A., Codegone, G., & Ogata, K. (2012). Mechanisms and processes of stratal disruption and mixing in the development of mélanges and broken formations: Redefining and classifying mélanges. *Tectonophysics, 568–569*, 7–24. https://doi.org/10.1016/j.tecto.2012.05.021

Finetti, I. R. (2006). Basic regional crustal setting and superimposed local pluton-intrusion-related tectonics in the Larderello-M. Amiata geothermal province, from integrated CROP seismic data. *Bollettino della Società Geologica Italiana, 125*(1), 117–146.

Fulignati, P. (2018). Hydrothermal fluid evolution in the "Botro ai Marmi" quartz-monzonitic intrusion, Campiglia Marittima, Tuscany, Italy. Evidence from a fluid inclusion investigation. *Mineralogical Magazine, 82*(5), 1169–1185. https://doi.org/10.1180/mgm.2018.116

Gandin, A., Giamello, M., Guasparri, G., Mugnaini, S., & Sabatini, G. (2000). The calcare cavernoso of the Montagnola Senese (Siena, Italy): Mineralogical-petrographic and petrogenetic features. *Mineralogica et Petrographica Acta, 43*, 271–289.

Giuntoli, F., & Viola, G. (2021). Cyclic brittle-ductile oscillations recorded in exhumed high-pressure continental units: a record of deep episodic tremor and slow slip events in the Northern Apennines. *Geochemistry, Geophysics, Geosystems, 22*(9), e2021GC009805. https://doi.org/10.1029/2021GC009805

Goscombe, B. D., Passchier, C. W., & Hand, M. (2004). Boudinage classification: End-member boudin types and modified boudin structures. *Journal of Structural Geology, 26*(4), 739–763. https://doi.org/10.1016/j.jsg.2003.08.015

Greenhalgh, S. R., McBride, J. H., Bartley, J. M., Keach, R. W., II, Britt, B. B., & Kowallis, B. J. (2015). Along-strike variability of thrust fault vergence. *Interpretation (Tulsa), 3*(3), SX1–SX12. https://doi.org/10.1190/INT-2014-0182.1

Häder, D.-P., Williamson, C. E., Wängberg, S.-A., Rautio, M., Rose, K. C., Gao, K., . . . Worrest, R. (2015). Effects of UV radiation on aquatic ecosystems and interactions with other environmental factors. *Photochemical & Photobiological Sciences : Official Journal of the European Photochemistry Association and the European Society for Photobiology, 14*(1), 108–126. https://doi.org/10.1039/c4pp90035a

Kabanov, P., Hauck, T. E., Gouwy, S. A., Grasby, S. E., & Van der Boon, A. (2023). Oceanic anoxic events, photic-zone euxinia, and controversy of sea-level fluctuations during the Middle-Late Devonian. *Earth-Science Reviews, 241*, 104415. https://doi.org/10.1016/j.earscirev.2023.104415

Kaiser, S. I., Aretz, M., & Becker, R. T. (2015). The global Hangenberg Crisis (Devonian-Carboniferous transition): Review of a first-order mass extinction. *Special Publication – Geological Society of London, 423*(1), 387–437. https://doi.org/10.1144/SP423.9

Kidder, D. L., & Worsley, T. R. (2010). Phanerozoic Large Igneous Provinces (LIPs), HEATT (Haline Euxinic Acidic Thermal Transgression) episodes, and mass extinctions. *Palaeogeography, Palaeoclimatology, Palaeoecology, 295*(1–2), 162–191. https://doi.org/10.1016/j.palaeo.2010.05.036

Kligfield, R. (1979). The Northern Apennines as a collisional orogen. *American Journal of Science, 279*(6), 676–691. https://doi.org/10.2475/ajs.279.6.676

Klingsohr, S. (1984). *Geologische Kartierung im Rücken von Monticiano und erzpetrographische Untersuchungen an Kernproben der Eisensulfidlagerstätte Campiano, Südtoskana, Italien*. Diploma thesis, LMU Munich.

Lazzarotto, A., & Moretti, A. (1973). Nuove considerazioni sul sondaggio del Belagaio in relazione ai risultati di recenti ricerche sulla dorsale Monticiano-Roccastrada nella Toscana meridionale. *Bollettino della Società Geologica Italiana, 94*, 151–166.

Lazzarotto, A., Aldinucci, M., Cirilli, S., Costantini, A., Decandia, F. A., Pandeli, E., . . . Spina, A. (2003). Stratigraphic correlation of the Upper Palaezoic-Triassic successions in southern Tuscany, Italy. *Bollettino della Società Geologica Italiana, 2*(2), 25–35. Retrieved from https://www.researchgate.net/publication/267443373

Le Breton, E., Handy, M. R., Molli, G., & Ustaszewski, K. (2017). Post-20 Ma motion of the Adriatic Plate: New constraints from surrounding orogens and implications for crust-mantle decoupling. *Tectonics, 36*(12), 3135–3154. https://doi.org/10.1002/2016TC004443

Liotta, D. (1991). The Arbia-Val Marecchia line, Northern Apennines. *Eclogae Geologicae Helvetiae, 84*(2), 413–430. https://doi.org/10.5169/seals-166782

Liotta, D. (2002). Stratigraphic and structural outline of the Montagnola Senese area (Southern Tuscany). *Bollettino della Società Geologica Italiana, 1*, 705–713.

Liotta, D., Brogi, A., Ruggieri, G., & Zucchi, M. (2021). Fossil vs. active geothermal systems: A field and laboratory method to disclose the relationships between geothermal fluid flow and geological structures at depth. *Energies, 14*(4), 933. https://doi.org/10.3390/en14040933

Lo Bue, R., Faccenda, M., & Yang, J. (2021). The role of Adria Plate lithospheric structures on the recent dynamics of the Central Mediterranean Region. *Journal of Geophysical Research – Solid Earth, 126*(10), 2021JB022377. https://doi.org/10.1029/2021JB022377

Lo Pò, D., Braga, R., Massonne, H.-J., Molli, G., Montanini, A., & Theye, T. (2015). Fluid-induced breakdown of monazite in medium-grade metasedimentary rocks of the Pontremoli basement (Northern Apennines, Italy). *Journal of Metamorphic Geology, 34*(1), 63–84. https://doi.org/10.1111/jmg.12171

Lo Pò, D., Braga, R., Massonne, H.-J., Molli, G., Montanini, A., & Bargossi, G. M. (2018). High-pressure tectono-metamorphic evolution of mylonites from the Variscan basement of the Northern Apennines, Italy. *Journal of Metamorphic Geology, 36*(1), 23–39. https://doi.org/10.1111/jmg.12281

Loreto, M. F., Zitellini, N., Ranero, C. R., Palmiotto, C., & Prada, M. (2021). Extensional tectonics during the Tyrrhenian backarc basin formation and a new morpho-tectonic map. *Basin Research, 33*(1), 138–158. https://doi.org/10.1111/bre.12458

Lu, J., Zhang, P., Dal Corso, J., Yang, M., Wignall, P. B., Greene, S. E., . . . Hilton, J. (2021). Volcanically driven lacustrine ecosystem changes during the Carnian Pluvial Episode (Late Triassic). *Proceedings of the National Academy of Sciences of the United States of America, 118*(40), e2109895118. https://doi.org/10.1073/pnas.2109895118

Lucente, F. P., & Speranza, F. (2001). Belt bending driven by lateral bending of subducting lithospheric slab: Geophysical evidences from the northern Apennines (Italy). *Tectonophysics, 337*(1–2), 53–64. https://doi.org/10.1016/S0040-1951(00)00286-9

Lucente, F. P., Margheriti, L., Piromallo, C., & Barruol, G. (2006). Seismic anisotropy reveals the long route of the slab through the western-central Mediterranean mantle. *Earth and Planetary Science Letters, 241*(3–4), 517–529. https://doi.org/10.1016/j.epsl.2005.10.041

Malusà, M. G., Faccenna, C., Baldwin, S. L., Fitzgerald, P. G., Rossetti, F., Balestrieri, M. L., . . . Piromallo, C. (2015). Contrasting styles of (U)HP rock exhumation along the Cenozoic Adria-Europe plate boundary (Western Alps, Calabria, Corsica). *Geochemistry, Geophysics, Geosystems, 16*(6), 1786–1824. https://doi.org/10.1002/2015GC005767

Marini, F., Pandeli, E., Tongiorgi, M., Pecchioni, E., & Orti, L. (2020). The Carboniferous-mid Permian successions of the Northern Apennines: New data from the Pisani Mts. inlier (Tuscany, Italy). *Italian Journal of Geosciences, 139*(2), 212–232. https://doi.org/10.3301/IJG.2019.27

Marshall, J. E. A., Lakin, J., Troth, I., & Wallace-Johnson, S. M. (2020). UV-B radiation was the Devonian-Carboniferous boundary terrestrial extinction kill mechanism. *Science Advances, 6*(22), eaba0768. https://doi.org/10.1126/sciadv.aba0768

Martini, I. P., Sagri, M., & Colella, A. (2001). Neogene–Quaternary basins of the inner Apennines and Calabrian arc. In G. B. Vai & I. P. Martini (Eds.), *Anatomy of an orogen: the Apennines and adjacent Mediterranean basins* (pp. 375–399). Dordrecht: Springer; https://doi.org/10.1007/978-94-015-9829-3_22

Martini, R., Gandin, A., & Zaninetti, L. (1989). Sedimentology, stratigraphy and micropaleontology of the Triassic evaporitic sequence in the subsurface of Boccheggiano and in some outcrops of Southern Tuscany (Italy). *Rivista Italiana di Paleontologia e Stratigrafia, 95*(1), 3–28. https://doi.org/10.13130/2039-4942/10632

Miall, A. D. (2010). *The geology of stratigraphic sequences*. Berlin: Springer; https://doi.org/10.1007/978-3-642-05027-5

Molli, G. (2008). Northern Apennine-Corsica orogenic system: An updated overview. *Special Publication – Geological Society of London, 298*(1), 413–442. https://doi.org/10.1144/SP298.19

Molli, G., Brogi, A., Caggianelli, A., Capezzuoli, E., Liotta, D., Spina, A., & Zibra, I. (2020). Late Palaeozoic tectonics in Central Mediterranean: A reappraisal. *Swiss Journal of Geosciences, 113*(1), 23. https://doi.org/10.1186/s00015-020-00375-1

Montanari, D., Ruggieri, G., Bonini, M., & Balestrieri, M. L. (2023). First application of low temperature thermochronology as a tool for geothermal exploration: A promising, preliminary test from the Larderello-Travale geothermal field (Italy). *Geothermics, 107*, 102603. https://doi.org/10.1016/j.geothermics.2022.102603

Müller, C., & Klemm, D.-D. (1989). *Sb and Hg in lower Paleozoic schists and blackschists – the basement of Sardinia and North Tuscany and their relationship to the epigenetic Plio-Pleistocene Sb-deposits in South Tuscany*. Abstracts 79th Annual Meeting of the Geologische Vereinigung (mineral deposits), Leoben, Austria, 15.–18.02.1989. Oxford: Blackwell Sci. Publ.

Murdock, D. J. E., & Smith, M. P. (2021). *Panderodus* from the Waukesha Lagerstätte of Wisconsin, USA: A primitive macrophagous vertebrate predator. *Papers in Palaeontology, 7*(4), 1977–1993. https://doi.org/10.1002/spp2.1389

Musumeci, G., Mazzarini, F., Tiepolo, M., & Di Vincenzo, G. (2011). U-Pb and ^{40}Ar-^{39}Ar geochronology of Palaeozoic units in the northern Apennines: Determining protolith age and alpine evolution using the Calamita Schist and Ortano Porphyroid. *Geological Journal, 46*(4), 288–310. https://doi.org/10.1002/gj.1266

Nicollet, K., Muñoz, F., Savard, G., Montanari, D., d'Ambrosio, M., Saccorotti, G., … Lupi, M. (2023). *Seismic methods to investigate the interplay between magmatism and tectonics in the Northern Apennines hinterland, Tuscan Magmatic Province, Italy*. EGU General Assembly 2023, Vienna, EGU23-12965. https://doi.org/10.5194/egusphere-egu23-12965

Novarese, V. (1946). Trivellazione nella Valle della Farma (Catena Metallifera Toscana). *Bollettino del R. Ufficio Geologico d'Italia, 69*, 8–19.

Pandeli, E., Sandrelli, F., & Aldinucci, M. (2004). The geology of Iano: an introductory summary. In E. Pandeli, F. A. Decandia, & M. Tongiorgi (Eds.), *The Paleozoic basement through the 500 Ma history of the Northern Apennine* (pp. 20–23). Florence, Italy: 32nd International Geological Congress, Vol. 1, PR01 to B15; field trip guide book B05. http://www.apuanegeopark.it/documents/Paper_Basement_Paleozoic_Pandeli.pdf

Paoli, G., Stokke, H., Rocchi, S., Sirevaag, H., Ksienzyk, A., Jacobs, J., & Košler, J. (2016). Basement provenance revealed by U-Pb detrital zircon ages: A tale of African and European heritage in Tuscany, Italy. *Lithos, 277*, 376–387. https://doi.org/10.1016/j.lithos.2016.11.017

Papeschi, S. (2015). Tectonometamorphic evolution of the Massa Unit in the Punta Bianca area (Northern Apennines, Italy). MSc thesis, Pisa. https://www.researchgate.net/publication/334560093

Papeschi, S., Musumeci, G., Massonne, H. J., Mazzarini, F., Ryan, E., & Viola, G. (2020). High-pressure (P = 1.5–1.8 GPa) blueschist from Elba: Implications for underthrusting and exhumation of continental units in the Northern Apennines. *Journal of Metamorphic Geology, 38*(5), 495–525. https://doi.org/10.1111/jmg.12530

Papeschi, S., Pontesilli, A., Romano, C., Rossetti, F., & Theye, T. (2022). Alpine subduction zone metamorphism in the Palaeozoic successions of the Monti Romani (Northern Apennines, Italy). *Journal of Metamorphic Geology, 40*(5), 919–953. https://doi.org/10.1111/jmg.12650

Pascucci, I., Martini, I. P., Sagri, M., & Sandrelli, F. (2007). Effects of transverse structural lineaments on the Neogene-Quaternary basins of Tuscany (inner Northern Apennines, Italy). In G. Nichols, E. Williams, & C. Paola (Eds.), *Sedimentary processes, environments and basins: a tribute to Peter Friend* (pp. 155–182). Kingston upon Thames, UK: Blackwell; https://doi.org/10.1002/9781444304411.ch8

Pasini, M. (1978a). *Edmondia scalaris* (M'Coy, 1844) and *Phestia* Cf. *attenuata* (Fleming, 1828) nella Formazione di Carpineta (Paleozoico Superiore della Toscana Meridionale). *Rivista Italiana di Paleontologia e Stratigrafia, 84*, 849–864.

Pasini, M. (1978b). Further paleontological records from the Upper Paleozoic outcrops of the Farma Valley (Southern Tuscany). In M. Tongiorgi (Ed.), *Report on the Tuscan Paleozoic basement* (pp. 71–75). Pisa: CNR, Rapporto interno del Sottopr. Energia Geotermica, Prog. Fin. Energetica.

Pasini, M. (1979a). Gli Archeodiscidi dell'affioramento di Poggio alle Pigne nella Valle del Torrente Farma (Toscana Meridionale). *Memorie della Società Geologica Italiana, 20*, 343–346.

Pasini, M. (1979b). I fusulinidi della valle del Torrente Farma (Toscana meridionale). *Memorie della Società Geologica Italiana, 20*, 323–342.

Pasini, M. (1991). Residual evidences of Permian carbonate platform within the Apennine sequences (Italy). *Bollettino della Società Geologica Italiana, 110*, 843–848.

Pasini, M., & Vai, G.-B. (1997). Review and updating of the Moscovian to Artinskian marine rocks in peninsula Italy. *Geodiversitas, 19*(2), 187–191. Retrieved from https://sciencepress.mnhn.fr/sites/default/files/articles/pdf/g1997n2a2.pdf

Pasini, M., & Winkler Prins, C. F. (1981). Carboniferous brachiopods from the locality Poggio alle Pigne in the Farma Valley (Southern Tuscany, Italy). *Rivista Italiana di Paleontologia e Stratigrafia, 86*(3), 459–468.

Perrone, V., Martín-Algarra, A., Critelli, S., Decandia, F. A., d'Errico, M., Estevez, A., . . . Zaghloul, M. M. (2006). 'Verrucano' and 'Pseudoverrucano' in the Central-Western Mediterranean Alpine chains: Palaeogeographical evolution and geodynamic significance. *Special Publication – Geological Society of London, 262*(1), 1–43. https://doi.org/10.1144/GSL.SP.2006.262.01.01

Pieruccioni, D., Galanti, Y., Biagioni, C., d'Orazio, M., Molli, G., & Vezzoni, S. (2020). Fornovolasco (Alpi Apuane, Tuscany, Italy): Type locality for the Permian felsic magmatism in the Northern Apennines. *International Journal of Earth Sciences, 109*(6), 2133–2134. https://doi.org/10.1007/s00531-020-01887-9

Pisarzowska, A., Rakociński, M., Marynowski, L., Szczerba, M., Thoby, M., Paszkowski, M., . . . Gereke, M. (2020). Large environmental disturbances caused by magmatic activity during the late Devonian Hangenberg Crisis. *Global and Planetary Change, 190*, 103155. https://doi.org/10.1016/j.gloplacha.2020.103155

Platt, J. P., Behr, W. M., & Cooper, J. (2015). Metamorphic core complexes: Windows into the mechanics and rheology of the crust. *Journal of the Geological Society, 172*(1), 9–27. https://doi.org/10.1144/jgs2014-036

Poli, G., & Peccerillo, A. (2016). The Upper Miocene magmatism of the Island of Elba (Central Italy): Compositional characteristics, petrogenesis and implications for the origin of the Tuscany Magmatic Province. *Mineralogy and Petrology, 110*(4), 421–445. https://doi.org/10.1007/s00710-016-0426-6

Pomar, L., Morsilli, M., Hallock, P., & Bádenas, B. (2012). Internal waves, an under-explored source of turbulence events in the sedimentary record. *Earth-Science Reviews, 111*(1–2), 56–81. https://doi.org/10.1016/j.earscirev.2011.12.005

Racki, G. (2020). A volcanic scenario for the Frasnian-Famennian major biotic crisis and other Late Devonian global changes: More answers than questions? *Global and Planetary Change, 189*, 103174. https://doi.org/10.1016/j.gloplacha.2020.103174

Ramsay, J. G., & Huber, M. I. (1987). *The techniques of modern structural geology: folds and fractures* (Vol. 2). London: Academic Press.

Redini, R. (1942). Rinvenimento di Antracolitico nel gruppo Monticiano-Roccastrada. *Bollettino della Società Geologica Italiana, 60*, 331–334.

Redini, R. (1958). Su varie questioni geologico-paleontologiche della Catena Metallifera Toscana. Sull' età Neotriassica della fauna del M. Rotondo, del M. Pisano e della fauna di Poggio Troncone, nelle Alpi Apuane. *Bollettino del Servizio Geologico d'Italia, 79*(3–5), 593–744.

Roche, V., Sternai, P., Guillou-Frottier, L., Menant, A., Jolivet, L., Bouchot, V., & Gerya, T. (2018). Emplacement of metamorphic core complexes and associated geothermal systems controlled by slab dynamics. *Earth and Planetary Science Letters, 498*, 322–333. https://doi.org/10.1016/j.epsl.2018.06.043

Rossetti, F., Balsamo, F., Villa, I. M., Bouybaouenne, M., Faccenna, C., & Funiciello, R. (2008). Pliocene–Pleistocene HT-LP metamorphism during multiple granitic intrusions in the southern branch of the Larderello geothermal field (southern Tuscany, Italy). *Journal of the Geological Society, 165*(1), 247–262. https://doi.org/10.1144/0016-76492006-132

Ryan, E., Papeschi, S., Viola, G., Musumeci, G., Mazzarini, F., Torgersen, E., … Ganerød, M. (2021). Syn-orogenic exhumation of high-P units by upward extrusion in an accretionary wedge: insights from the Eastern Elba Nappe Stack (Northern Apennines, Italy). *Tectonics, 40*(5), e2020TC006348. https://doi.org/10.1029/2020TC006348

Sammuri, P., & Pantaloni, M. (2018). Mineral resources mapping in the 18th century (Tuscany, Italy). *De Re Metallica, 30*, 37–48.

Sammuri, P., & Scapigliati, W. (2022). La miniera di caolino del Monte Alto a Piloni di Torniella (Roccastrada, Grosseto): geologia e storia industriale dell'area estrattiva. *Atti della Società*

Toscana di Scienze Naturali – Memorie, A 129, 65–80. https://doi.org/10.2424/ASTSN.M.2022.07

Servizio Geologico d'Italia (2004). *Geological map of Italy 1:250.000, Regione Toscana*. Università di Siena, Dipartimento di Scienze della Terra, Centro di Geotecnologie.

Signorini, R. (1947). Cenni preliminari su un rilevamento nella Val di Merse. *Bollettino della Società Geologica Italiana, 4*, 433–446.

Signorini, R. (1967). *Note illustrative della Carta Geologica D'Italia alla scala 1: 100.000, F 120*. Siena: Servizio Geologica d'Italia.

Sillitoe, R. H., & Brogi, A. (2021). Geothermal systems in the Northern Apennines, Italy: Modern analogues of Carlin-style gold deposits. *Economic Geology and the Bulletin of the Society of Economic Geologists, 116*(7), 1491–1501. https://doi.org/10.5382/econgeo.4883

Sirevaag, H., Jacobs, J., Ksienzyk, A., Rocchi, S., Paoli, G., Jørgensen, H., & Košler, J. (2016). From Gondwana to Europe: The journey of Elba Island (Italy) as recorded by U-Pb detrital zircon ages of Paleozoic metasedimentary rocks. *Gondwana Research, 38*, 273–288. https://doi.org/10.1016/j.gr.2015.12.006

Spina, A., Cirilli, S., Decandia, F. A., & Lazzarotto, A. (2001). *Palynological data from the Poggio al Carpino sandstones Fm. (Southern Tuscany, Italy)*. Proceedings of the International Congress on "Stratigraphic and Structural Evolution of the Late Carboniferous to Triassic Continental and Marine Successions in Tuscany (Italy). Regional Reports and General Correlation", 30.04.–07.05.2001 (pp. 65–66). Siena, Italy.

Spina, A., Aldinucci, M., Brogi, A., Capezzuoli, E., Cirilli, S., & Liotta, D. (2021). *Permian sporomorphs from upper Palaeozoic succession of Southern Tuscany (Italy): new constraints for the stratigraphy and palaeogeographic setting of the Tuscan Domain*. EGU General Assembly, 19–30 Apr. 2021. https://doi.org/10.5194/egusphere-egu21-2112

Stampfli, G. M. (2005). Plate tectonics of the Apulia-Adria microcontinents. In I. R. Finetti (Ed.), *Atlases in geosciences 1. CROP Project – deep seismic exploration of the Central Mediterranean and Italy* (pp. 747–766).

Stea, B. (1971). Mineralizzazione ad antimonio. In La Toscana meridionale. *Rendiconti della Società Italiana di Mineralogia e Petrologia, 27*, 422–442.

Stephan, S., Kroner, U., Romer, R. L., & Rösel, D. (2019). From a bipartite Gondwana shelf to an arcuate Variscan belt: The early Paleozoic evolution of northern Peri-Gondwana. *Earth-Science Reviews, 192*, 491–512. https://doi.org/10.1016/j.earscirev.2019.03.012

Tanelli, G. (1983). Mineralizzazione metallifere e minerogenesi della Toscana. *Memorie della Società Geologica Italiana, 25*, 91–109.

Tavarnelli, E. (1996). The effects of pre-existing normal faults on thrust ramp development: An example from the northern Apennines, Italy. *Geologische Rundschau, 85*(2), 363–371. https://doi.org/10.1007/BF02422241

Tavarnelli, E., Mazzarini, F., Scialoja, E., & Isola, I. (2021). Deformation history of a foredeep basin during the incorporation of its deposits within an advancing orogenic wedge: The case of the Oligocene-Early Miocene Macigno Costiero Formation, southern Tuscany, northern Apennines, Italy. *Journal of Structural Geology, 147*, 104347. https://doi.org/10.1016/j.jsg.2021.104347

Terrill, D. F., Jarochowska, E., Henderson, C. M., Shirley, B., & Bremer, O. (2022). Sr/Ca and Ba/Ca ratios support trophic partitioning within a Silurian conodont community from Gotland, Sweden. *Paleobiology, 48*(4), 601–621. https://doi.org/10.1017/pab.2022.9

Tirel, C., Gautier, P., van Hinsbergen, D. J. J., & Wortel, M. J. R. (2009). Sequential development of interfering metamorphic core complexes: Numerical experiments and comparison with the Cyclades, Greece. *Special Publication – Geological Society of London, 311*(1), 257–292. https://doi.org/10.1144/SP311.10

Vai, G. B. (1991). Palaeozoic strike-slip rift pulses and palaeogeography in the circum-Mediterranean Tethyan realm. *Palaeogeography, Palaeoclimatology, Palaeoecology, 87*(1–4), 223–252. https://doi.org/10.1016/0031-0182(91)90137-G

Vai, G. B. (2001). Structure and stratigraphy: an overview. In G. B. Vai & I. P. Martini (Eds.), *Anatomy of an orogen: the Apennines and adjacent Mediterranean basins* (pp. 15–31). Dordrecht: Springer; https://doi.org/10.1007/978-94-015-9829-3_3

Van der Meer, D. G., Stap, L. B., Scotese, C. R., Mills, B. J. W., Sluijs, A., & van Hinsbergen, D. J. J. (2025). Phanerozoic orbital-scale glacio-eustatic variability. *Earth and Planetary Science Letters, 667*, 119526. https://doi.org/10.1016/j.epsl.2025.119526

Vannoli, P., Martinelli, G., & Valensise, G. (2021). The seismotectonic significance of geofluids in Italy. *Frontiers in Earth Science (Lausanne), 9*, 579390. https://doi.org/10.3389/feart.2021.579390

Vezzoni, S., Biagioni, C., d'Orazio, M., Pieruccioni, D., Galanti, Y., Petrelli, M., & Molli, G. (2018). Evidence of Permian magmatism in the Alpi Apuane metamorphic complex (Northern Apennines, Italy): New hints for the geological evolution of the basement of the Adria plate. *Lithos, 318–319*, 104–123. https://doi.org/10.1016/j.lithos.2018.08.003

Vignaroli, G., Faccenna, C., Rossetti, F., & Jolivet, L. (2009). Insights from the Apennines metamorphic complexes and their bearing on the kinematic evolution of the orogen. *Special Publication – Geological Society of London, 311*, 235–256. https://doi.org/10.1144/SP311.9

Viola, G., Torgersen, E., Mazzarini, F., Musumeci, G., van der Lelij, R., Shoenenberger, J., & Garofalo, P. S. (2018). New constraints on the evolution of the Inner Northern Apennines by K-Ar dating of Late Miocene-Early Pliocene compression of the Island of Elba, Italy. *Tectonics, 37*(9), 3229–3243. https://doi.org/10.1029/2018TC005182

Viti, M., Mantovani, E., Babbucci, D., Tamburelli, C., Cenni, N., Baglione, M., & d'Intinosante, V. (2015). Belt-parallel shortening in the Northern Apennines and seismotectonic implications. *International Journal of Geosciences, 6*(8), 938–961. https://doi.org/10.4236/ijg.2015.68075

Walker, R. G. (1984). *Facies models*. Hamilton, Canada: Geological Association of Canada.

Zuza, A. V., & Cao, W. (2023). Metamorphic core complex dichotomy in the North American Cordillera explained by buoyant upwelling in variably thick crust. *GSA Today, 33*(3–4), 4–11. https://doi.org/10.1130/GSATG548A.1

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