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Mid- to late Permian microfloristic evidence in the metamorphic successions of the Northern Apennines: insights for age-constraining and palaeogeographical correlations

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Abstract: Recent studies on the stratigraphy of Paleozoic successions belonging to key sectors of the inner Northern Apennines provide data that can be used to propose new hypotheses about the palaeogeography of the western Mediterranean domain during post-Hercynian times. We report here the first evidence of mid- to late Permian (Guadalupian–Lopingian) palynomorphs in the chronostratigraphically highly debated metamorphic units of the Northern Apennines (the Rio Marina Formation, the Mt Calamita Formation and the Le Cetine Formation) and consider the stratigraphic implications for the Tuscan ‘basement’ to which the study rocks belong and their palaeogeographical attribution considering their Gondwana affinity. These results agree with the interpretation that the Alpine Tethyan sedimentary cycle was already active during the mid- to late Permian.

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The inner Northern Apennines, as a result of their evolution and geological setting, represent a key area for analyses of Paleozoic successions and the overlying units. This region is characterized by exhumed metamorphic units (Brunet *et al.* 2000) consisting of Paleozoic–Mesozoic successions typical of the Adria margin (Vai & Martini 2001). Their exhumation is a consequence of lithospheric and crustal-thinning processes that occurred during extensional tectonics from the Miocene to the present day (Liotta *et al.* 1998; Brogi & Liotta 2008; Barchi 2010), of which the clearest evidence is the opening of the Tyrrhenian Sea (Bartole 1995). For this reason, the Paleozoic successions consist of isolated outcrops presently exposed only in the metamorphic successions of the Tuscan archipelago (i.e. the Middle Tuscan Ridge and Apuan Alps) and Northern Latium (Fig. 1). Nevertheless, their chronostratigraphic characterization can help in reconstructing the stratigraphic evolution of the inner Northern Apennines. Despite their importance, most of the Paleozoic successions have been studied separately, by different researchers, at different times and with different approaches. This has produced discordant results, such as the age of some of the Paleozoic rocks (ranging from Carboniferous to Permian) locally exposed on the island of Elba (e.g. the Rio Marina and Mt Calamita formations, Bagnoli *et al.* 1979) and in southern Tuscany (e.g. the Le Cetine Formation, Bagnoli *et al.* 1978; Coccozza *et al.* 1978). This mismatch has important consequences for the most recent part of the Paleozoic succession because the late Carboniferous is considered to represent the beginning of the segmentation of Gondwana in the north of this region after the Hercynian orogeny (e.g. Stampfli & Borel 2002; von Raumer *et al.* 2002). These Paleozoic sediments are now part of the stacked metamorphic units of the Northern Apennines (Bagnoli *et al.* 1980; Pandeli *et al.* 1988, 1991; Theye *et al.* 1997; Giorgetti *et al.* 1998; Brogi & Giorgetti 2012) and are considered to predate the local Tethyan deposits representing the Alpine cycle (Vai 2001, 2003).

As a consequence of this largely accepted general picture, many stratigraphic and structural studies (e.g. Coccozza *et al.* 1974; Bagnoli *et al.* 1979, 1980; Elter & Pandeli 1991, 2005; Pandeli *et al.* 1994; Vai 2001, 2003; Cassinis 2003; Lazzarotto *et al.* 2003; Aldinucci *et al.* 2008) have tried to correlate the different Paleozoic successions that outcrop patchily throughout the Northern Apennines. The results are unfortunately conditioned by rare and uncertain biostratigraphic attributions, generally based on scattered fossil findings (plant remains, conodonts, brachiopods and forams; e.g. Vai 1972; Coccozza *et al.* 1974; Vai & Francavilla 1974; Ferrari *et al.* 1977; Pasini 1978; Bagnoli *et al.* 1979; Duranti *et al.* 1992; Landi Degl’Innocenti *et al.* 2008). In this context, recent zircon U–Pb geochronology (Musumeci *et al.* 2011; Sirevaag *et al.* 2016; Paoli *et al.* 2016) has been applied to a number of Paleozoic successions and has provided ages between the Permian and late Carboniferous. Nevertheless, the recent discovery of Permian metarhyolite bodies in the Apuan Alps (Vezzoni *et al.* 2018) supplies an input for a new stratigraphic and palaeogeographical reconstruction using a number of alternative approaches.

Palynology represents a successful methodology to obtain biostratigraphic data from low-grade metamorphic units with only scarce preserved biomineralized fossils (e.g. Pflug & Reitz 1992; Machado & Flores 2015). Palynomorphs are organic-walled microfossils and, because they are composed entirely of proteinaceous material, they can be internally re-ordered (but not destroyed) by the processes acting during burial (depending on the depth and duration of burial, the geothermal flux and the geochemistry of the fluids present) and low-grade metamorphism (e.g. Hanel *et al.* 1999; Bernard *et al.* 2007; Traverse 2008; Spina 2015; Ariana-Sab *et al.* 2016; Cornamusini *et al.* 2017; Schito *et al.* 2017, 2019; Riboulleau *et al.* 2018; Spina *et al.* 2018a; Galasso *et al.* 2019). We applied palynology in these stratigraphic studies as a useful tool for the age attribution of the investigated Tuscan low-

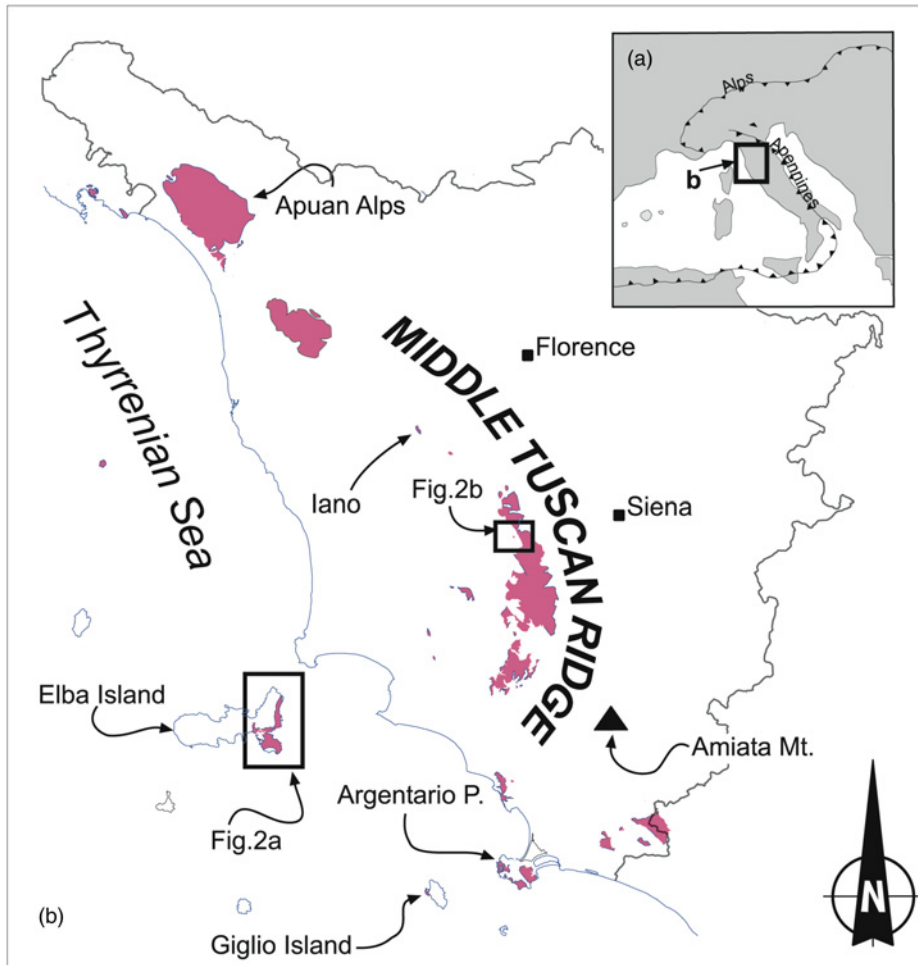


Fig. 1. (a) Geographical location of Tuscany in the framework of the inner Northern Apennines. (b) Distribution of metamorphic units, including Paleozoic rocks (in red). The locations of the studied areas are also shown.

grade metamorphic successions. In the case of the Paleozoic successions of the Northern Apennines, this integrated approach can help to correlate the analysed successions with the well-known, independently dated successions of the western Dolomites (the Southern Alps, Bulla and Tesero sections; Spina *et al.* 2015) and with those belonging to other palaeogeographical domains. The results of this work highlight the underestimation of the occurrence of mid- to late Permian sediments in the Northern Apennines and provide new data to constrain the fragmentation of Gondwana at least since the Permian.

Outline of the evolution of the Northern Apennines and Tyrrhenian Sea

The Northern Apennines and Tyrrhenian Sea (i.e. the inner Northern Apennines) originated from the convergence and collision (Cretaceous–Early Miocene) between the African (Adria microplate) and European (Sardinia–Corsica Massif) plates and the subsequent extensional tectonics (Miocene to Present) affecting the Northern Apennines belt (Brogi & Liotta 2008; Molli 2008; Barchi 2010).

As a consequence of this latter extensional regime, the exposures of Tuscan metamorphic and Paleozoic successions are mainly localized in two different areas: the Tuscan archipelago (Elba Island) and inland along the Middle Tuscan Ridge (Fig. 1). Elba Island is located in the central part of the Tyrrhenian Sea (Fig. 1) and consists of a tectonic pile (Fig. 2a) formed by stacked continental (Tuscan Domain) and oceanic (Ligurian Domain) tectonic units (Bortolotti *et al.* 2001; Bianco *et al.* 2015; Principi *et al.* 2015). The Middle Tuscan Ridge is an arcuate morpho-tectonic feature extending from the Apuan Alps to the Argentario Promontory. In its southern portion it is mainly characterized by Paleozoic–Triassic

metasiliciclastic successions of the Monticiano-Roccastrada Unit (Costantini *et al.* 1988), which experienced HP–LT and later greenschist facies metamorphic conditions (Giorgetti *et al.* 1998; Brunet *et al.* 2000; Brogi & Giorgetti 2012). The Paleozoic successions (Bagnoli *et al.* 1979; Lazzarotto *et al.* 2003) consist of continental to shallow marine formations composed of different lithotypes (from metaconglomerate to phyllite to metacarbonate to radiolarite), organized in locally thick and strongly deformed successions (up to 800 m; Farma Formation; Aldinucci *et al.* 2008). Their interpretation, lateral correlation, age attribution and palaeogeographical setting are still a matter of debate, mainly due to the absence of robust chronological constraints.

Analysed successions

The analyses were carried in key successions of Elba Island and southern Tuscany (Fig. 1). We sampled the Paleozoic successions on Elba Island (the Mt Calamita and Rio Marina formations; Garfagnoli *et al.* 2005; Caggianelli *et al.* 2018; Fig. 2a) at the base of Continental Units 1 and 3 as reported by Bianco *et al.* (2015). In southern Tuscany, we sampled the succession (the Le Cetine Formation; Bagnoli *et al.* 1978) exposed in the Middle Tuscan Ridge (Fig. 2b), structurally located in the Monticiano-Roccastrada Unit (Costantini *et al.* 1988), which is part of a duplex structure system in which the Triassic and Paleozoic rocks are repeated. Key samples were used for palynological analyses (see Table 1).

Materials and methods

Twenty-two samples of black phyllite (five samples from the Mt Calamita Formation, nine from the Rio Marina Formation and eight from the Le Cetine Formation) were used in the palynological

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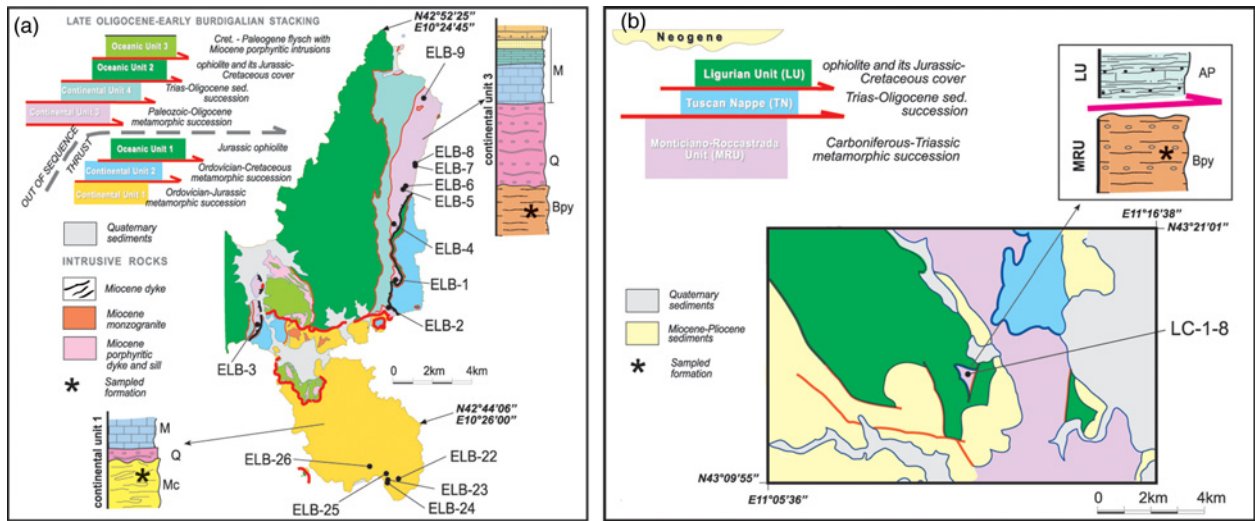


Fig. 2. Detail of the sampled area with the local tectono-stratigraphic setting and locations of the collected samples. (a) Eastern Elba Island (redrawn from Bianco *et al.* 2015). Continental unit 3: Bpy, black phyllite (Rio Marina Formation); Q, quartzite and phyllite (Verrucano Formation); M, Mesozoic–Tertiary metasediments. Continental unit 1: Mc, mica schist and black phyllite (Mt Calamita Formation); Q, quartzite and phyllite (Verrucano Formation); M, Mesozoic dolostone. (b) Central Middle Tuscan Ridge: Ligurian Unit (LU): AP, shale and carbonates; Monticiano-Roccastrada Unit (MRU): Bpy, metaconglomerate and black shale (Le Cetine Formation). See main text for more detail.

studies. All the samples (*c.* 20 g each) were treated by standard palynological acid maceration (with 37% HCl and 50% HF), density separation of the organic matter (using a ZnCl₂ solution) and filtration of the organic-rich residue at 10 μm. As a result of the high degree of thermal alteration, the organic residue was treated with Schultz solution. Light microscope observations were performed on palynological slides using a Leica DM1000 microscope using the differential interference contrast technique in transmitted light. Images were captured using the camera on the digital microscope and successively corrected for contrast and brightness using the open-source Gimp software. The palynological slides were stored at the Sedimentary Organic Matter Laboratory of the Department of Physics and Geology, University of Perugia, Italy.

Elba Island: the Mt Calamita Formation

The Mt Calamita Formation is part of Continental Unit 1 in Bianco *et al.* (2015), referred as the Porto Azzurro Unit in Garfagnoli *et al.* (2005). The Mt Calamita Formation represents the deepest part of a tectonic unit. It was involved in the collisional stage during the Oligocene–Miocene and then experienced HT–LP metamorphism related to the emplacement and cooling of the Porto Azzurro monzogranite during the Late Miocene–Early Pliocene (e.g.

Duranti *et al.* 1992; Dini *et al.* 2002; Caggianelli *et al.* 2018). According to Principi *et al.* (2015), the maximum thickness of the Mt Calamita Formation is *c.* 800 m and it can be divided in several sub-units on the basis of the main lithological differences. The whole Paleozoic succession is overlain by a sedimentary cover represented by Triassic metasiliciclastic (Verrucano Group, referred to as the Quarziti di Barabarca Formation in Garfagnoli *et al.* 2005) and Jurassic metacarbonate successions (Fig. 2a). The Mt Calamita Formation is mainly composed of mica schist, with levels of phyllite, metasandstone and quartzite (Fig. 3a, b; Garfagnoli *et al.* 2005). Metre to decimetre thick metabasite lenses, formed by amphibole–plagioclase layers, are locally embedded within the metasedimentary succession parallel to the main schistosity (Barberi *et al.* 1967; Caggianelli *et al.* 2018). The whole succession was intruded by decimetre to metre thick leucogranite dykes, principally on the eastern side of the promontory. As a result of the lack of biomineralized fossils, the Mt Calamita Formation has been uncertainly attributed to: (1) the pre-Carboniferous (Garfagnoli *et al.* 2005); (2) the early Carboniferous (Musumeci *et al.* 2011); (3) a generic ?Precambrian–Paleozoic age (Principi *et al.* 2015); or (4) deposition from at least the early Cambrian (Sirevaag *et al.* 2016).

Samples were collected in decimetre- to decameter-sized bodies of black, graphite-rich phyllite (samples ELB 22–26; Fig. 2a and

Table 1. Analysed samples and related geographical coordinates

Le Cetine Formation	Latitude	Longitude	Rio Marina Formation	Latitude	Longitude
LC-1	43°13'42.89"N	11°10'1.23"E	ELB-1	42°46'53.19"N	10°25'3.73"E
LC-2	43°13'42.89"N	11°10'1.23"E	ELB-2	42°46'29.60"N	10°24'50.01"E
LC-3	43°13'42.89"N	11°10'1.23"E	ELB-3	42°45'59.00"N	10°20'41.36"E
LC-4	43°13'43.26"N	11°10'0.89"E	ELB-4	42°47'58.62"N	10°25'1.47"E
LC-5	43°13'43.07"N	11°9'59.43"E	ELB-5	42°48'48.73"N	10°25'12.05"E
LC-6	43°13'42.37"N	11°9'58.03"E	ELB-6	42°48'48.72"N	10°25'22.61"E
LC-7	43°13'42.37"N	11°9'58.03"E	ELB-7	42°49'21.48"N	10°25'50.07"E
LC-8	43°13'42.84"N	11°9'58.29"E	ELB-8	42°49'21.48"N	10°25'50.07"E
			ELB-9	42°51'2.94"N	10°26'5.74"E
Calamita Complex	Latitude	Longitude	ELB-24	42°42'57.78"N	10°24'56.62"E
ELB-22	42°42'49.81"N	10°25'3.55"E	ELB-25	42°42'58.27"N	10°24'56.70"E
ELB-23	42°42'56.18"N	10°24'57.68"E	ELB-26	42°43'4.11"N	10°24'10.89"E

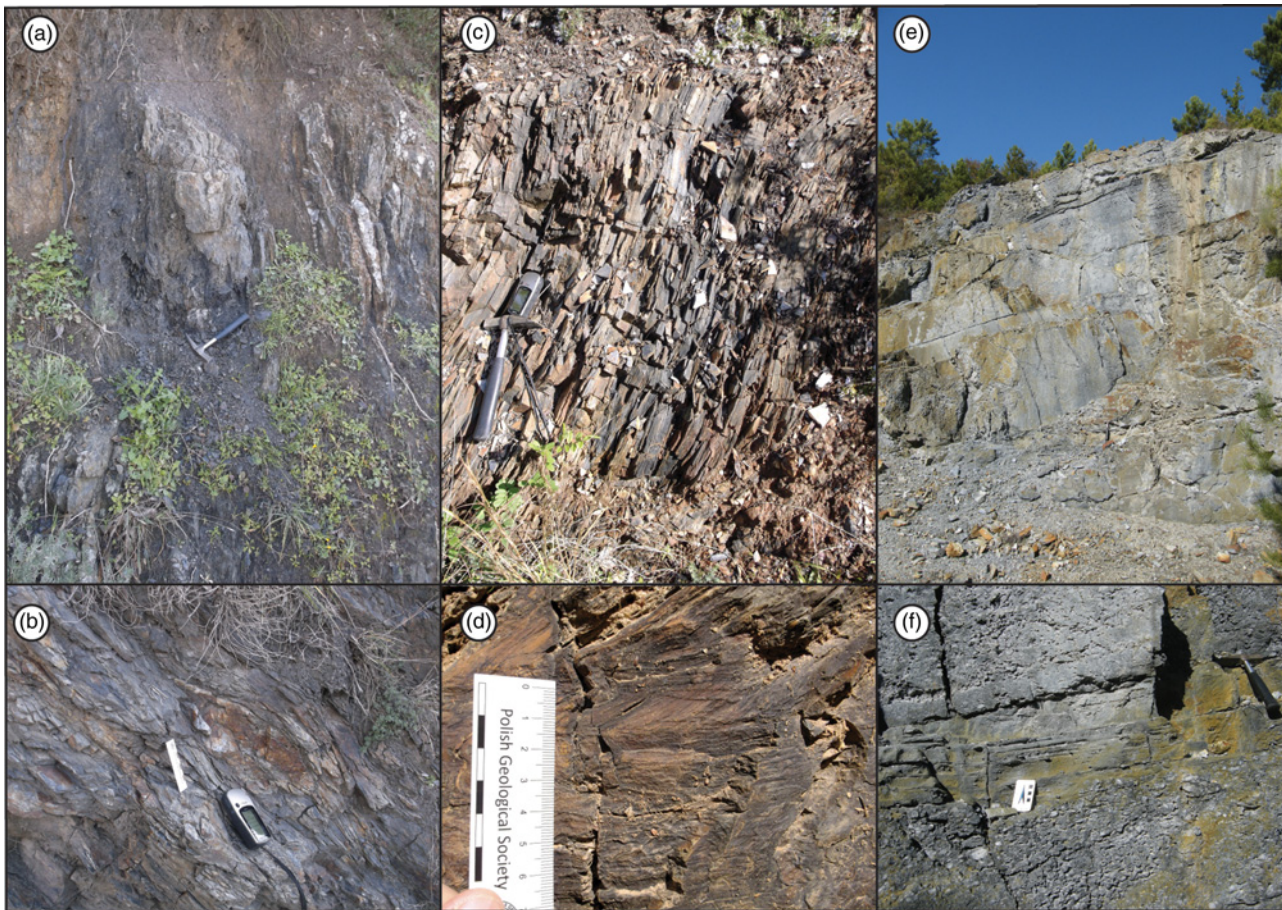


Fig. 3. Outcrops of the analysed Paleozoic formations from southern Tuscany. (a) Vertical alternation of black phyllite and metasandstone from the Mt Calamita Formation (ELB-23); (b) detail of the deformed and folded black phyllite in the Mt Calamita Formation (ELB-23); (c) well-stratified black phyllite from the Rio Marina Formation (ELB-1); (d) example of cross-stratification inside the Rio Marina Formation (ELB-4); (e) partial view of the quarry containing the Le Cetine Formation (samples LC-1 and LC-3 derived from the lower portion); and (f) example of fining-upwards succession (channel-fill) in the Le Cetine Formation (LC-6).

Fig. 3a, b; Table 1). These samples yielded a palynofacies consisting mainly of abundant inertinite (i.e. opaque phytoclasts consisting of black equidimensional and lath-shaped particles composed of charcoal and biochemically oxidized wood). Vitrinite (mainly consisting of partially oxidized wood) fragments and orange to dark brown, rarely corroded, palynomorphs also occur.

The microfloristic assemblage (Plate 1, Appendix A) is not well diversified and consists of spores of *Thymospora opaqua*, *Densoisporites* sp. cf. *playfordii*, *Cyclogranisporites* sp., *Kraeuselisporites* sp. and *Lundbladispota* sp., bisaccate pollen grains of *Klausipollenites schaubergerii* and *Alisporites* spp. The *incertae sedis* organic microfossil *Reduviasporonites chalcatus* is abundantly present in all the processed samples.

Elba Island: the Rio Marina Formation

The Rio Marina Formation is part of Continental Unit 3 in Bianco *et al.* (2015) and is related to the Monticiano-Roccastrada Unit in Bortolotti *et al.* (2001). It crops out mainly on the eastern side of Elba Island (Fig. 2a). It is c. 250 m thick and consists of black to grey phyllite and metasandstone, as well as lenses of metaconglomerate and coarse-grained metasandstone (Fig. 3c, d; Bortolotti *et al.* 2001). The arenaceous layers show planar lamination, cross-bedding and cross-lamination, with a weakly recrystallized matrix defined by a quartzose sericitic assemblage (Bagnoli *et al.* 1979). Bagnoli *et al.* (1979) attributed the deposit to a prograding delta depositional environment. The Rio Marina Formation is unconformably overlain by the Verrucano Group (Principi *et al.* 2015).

The fossil content is characterized by scattered pelecypods, brachiopods, crinoids, echinoids, cephalopods, forams and vegetal remains (Bodechtel 1964; Kahler & Kahler 1969; Coccozza *et al.* 1974; Vai 1978; Bagnoli *et al.* 1979). In detail, Bodechtel (1964), from a single sample uncertainly attributed to the Rio Marina Formation, recorded fusulinids as *Parafusulina* sp., probably belonging to the Artinskian *Praeparafusulina lutugini*–*Parafusulina pseudojaponica* zones (Kahler & Kahler 1969). Coccozza *et al.* (1974) and Vai (1978) extended the age of the Rio Marina Formation to the late Pennsylvanian on the basis of plant remains, uncertainly attributed to *Calamites* sp. cf. *suckovi*.

Sirevaag *et al.* (2016) measured U–Pb and Pb–Pb isotopic ratios in the detrital zircon of the Rio Marina Formation, with an age ranging between c. 2050 and 265 Ma. They gave a minimum age of c. 280 Ma to this formation on the basis of a small peak (about 5%) of the youngest detrital zircon.

Nine samples were collected from the Rio Marina Formation (Fig. 2a; Table 1). Only the phyllitic and siltitic lithotypes were sampled because these were richer in organic matter. Phytoclasts were found in all samples as inertinite and vitrinite. A few acritarchs (*incertae sedis* marine phytoplankton) and sporomorphs were found in five samples (ELB 2–6).

The microflora content (Plate 1) is characterized by monolete and trilete spores as *Densoisporites playfordii*, *T. opaqua*, *Calamospora* sp., *Cyclogranisporites* sp., *Kraeuselisporites* sp., *Laevigatosporites* sp., ?*Lundbladispota* sp. Bisaccate pollen grains, such as *Lueckisporites virkkiae* and *Alisporites nuthallensis*, were also recorded. Monosaccates, such as *Potoniesporites* sp., very rarely

Permian microflora in Tuscan successions

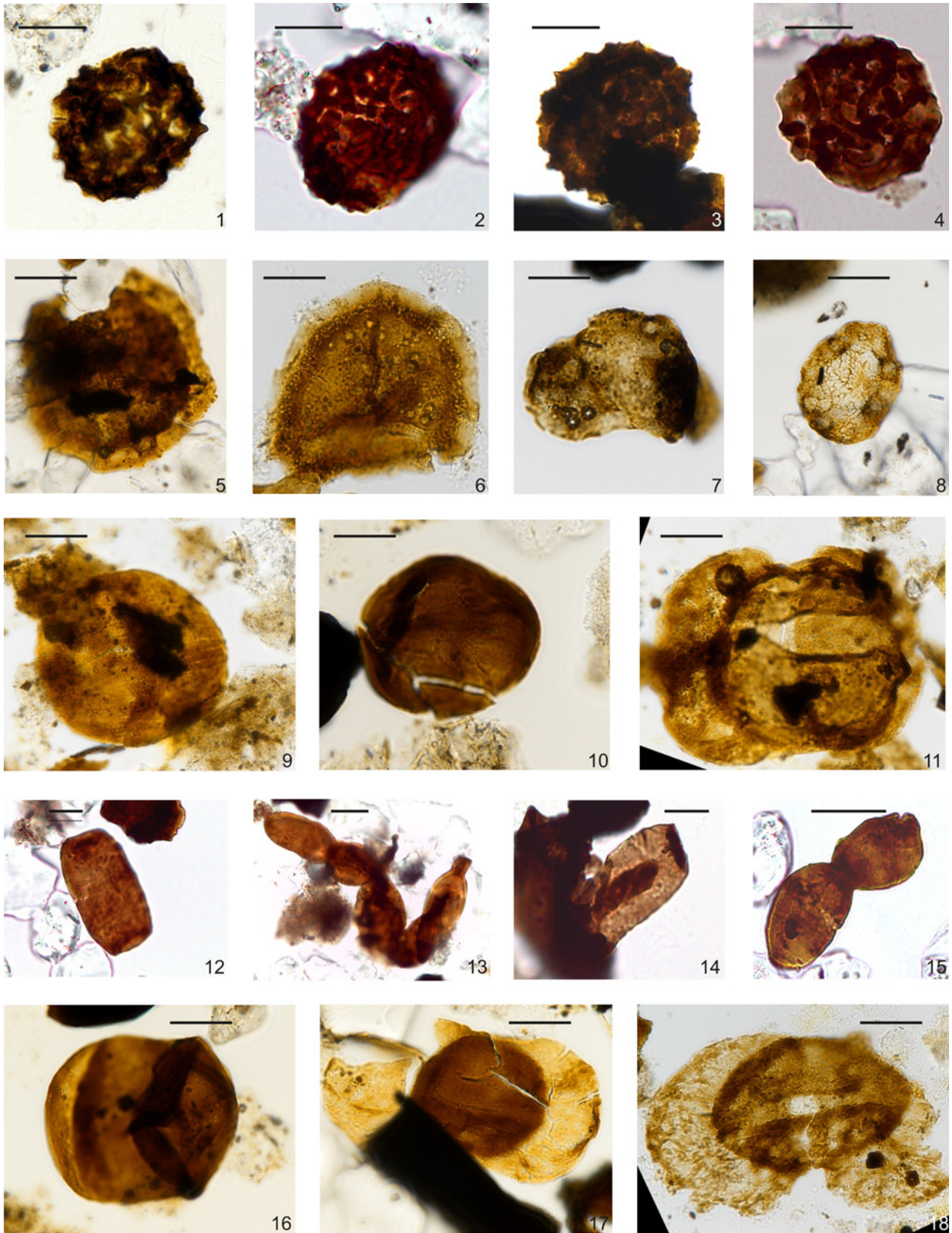


Plate 1. Sporomorphs recorded from the studied successions (scale bar = 10 µm). 1–4, *Thymospora opaqua* Singh 1964 (1, 2, Mt Calamita Formation, slide ELB-22; 3, 4, Rio Marina Formation, slide ELB-2); 5, *Kraeuselisporites* sp. (Rio Marina Formation, slide ELB-2); 6, *Densoisporites playfordii* Balme 1970 (Rio Marina Formation, slide ELB-2); 7, *Klausipollenites schaubergeri* (Potonié and Klaus) Jansonius 1962 (Mt Calamita Formation, slide ELB-23); 8, *Florinites* sp. (Rio Marina Formation, slide ELB-3); 9, 10, *Vittatina costabilis* Wilson 1962 (Rio Marina Formation, slide ELB-2); 11, *Lueckisporites virkkiae* Potonié and Klaus 1954 (Rio Marina Formation, slide ELB-3); 12–15, *Reduviasporonites chalastus* (Foster) Elsik 1999 (Le Cetine Formation: 12, slide LC-3; 13, slide LC-4; 14, 15, slide LC-5); 16, *Alisporites* sp. (Rio Marina Formation, slide ELB-2); 17, *Taeniaesporites* sp. (Mt Calamita Formation, slide ELB-23); and 18, *Corisaccites alutas* Venkatachala and Kar 1966 (Rio Marina Formation, slide ELB-3).

occurred, in addition to monosulcates (e.g. *Cycadopites* sp.) and polyplacates (e.g. *Vittatina costabilis*). *R. chalastus* was abundant in all the processed samples.

Middle Tuscan Ridge: the Le Cetine Formation

This succession is located west of the Montagnola Senese (Fig. 2b; Table 1) and is exposed in an abandoned quarry (Fig. 3e; Menchetti & Batoni 2015). It consists of black phyllite, metasandstone and metaconglomerate (Fig. 3e, f), the latter forming thick levels interpreted as channel-fill deposits (Bagnoli *et al.* 1978). This formation is exposed for c. 70 m. As a result of the limited outcrop, this unit has been variously correlated with other Paleozoic formations, such as: (1) the Poggio al Carpino Formation (mid- to late Permian; Lazzarotto *et al.* 2003) by Cocozza *et al.* (1978); (2) the Rio Marina Formation; (3) the Iano Paleozoic succession (late Carboniferous–early Permian; Bagnoli *et al.* 1978, 1979); and (4) the Verrucano Group (Triassic; Puxeddu *et al.* 1979). Costantini *et al.* (2009) refer the Le Cetine Formation to the late Carboniferous.

Eight samples were collected from the black phyllite intercalated between the sandstones in this stratigraphic unit (Figs. 2b and 3e, f; Table 1). No sporomorph was recorded, nevertheless three samples (LC 3–5) yielded a palynoassemblage dominated by *R. chalastus* (Plate 1).

Discussion

The microfloristic content of the Rio Marina and Mt Calamita formations allowed us to assign a mid- to late Permian age on the basis of similar morphological characteristics documented in the Southern Alps and other key areas of Laurasia and northern Gondwana. The assemblages contain important index species used in worldwide correlation, such as the bisaccate taeniate *L. virkkiae* known as a Guadalupian–Lopingian marker fossil. Its first occurrence was documented in the Russian Kazanian (i.e. Wordian) type sequence (Utting *et al.* 1997). It was documented in an assemblage with *K. schaubergerii*, *Taeniaesporites pellucidus*, *D. playfordii* and *R. chalastus* in the late Permian of the Southern Alps (Visscher & Brugman 1988; Massari *et al.* 1994; Cirilli *et al.* 1998). Spina *et al.* (2015) recorded a similar assemblage in the Tesero and Bulla sections from the upper Bellerophon Formation along the *praeparvus* conodont biozone of Changxingian age (Farabegoli *et al.* 2007). The late Permian of Spain is also marked by the occurrence of *L. virkkiae* (Doubinger *et al.* 1990). Taeniate and polyplacate pollens such as *L. virkkiae* and *V. costabilis* occurred in the Germanic Basin in the lower part of the Lower Buntsandstein, which is considered as Permian (Utting & Piasecki 1995; Utting *et al.* 2004). Similarities to the Tuscan palynological associations have also been noted in the Lopingian microflora from other northern European regions (e.g. the UK, Clarke 1965; Ireland, Visscher 1971; Hungary, Haas *et al.* 1988; the Barents Sea, Mangerud 1994; and Greenland: Twitchett *et al.* 2001). The palynological assemblages yielded from the Flowerpot Formation of Wordian age in Oklahoma (USA) (Chudinov 1965) contain abundant *L. virkkiae*. *L. virkkiae*, *K. schaubergerii* and *R. chalastus* were also found in the Meishan section (China), ratified as the Global Stratotype Section and Point for the basal boundary of the Triassic System (Yin *et al.* 2001), in the *Vittatina–Protohaploxylinus* Assemblage, attributed to the Changxingian–lower Induan on the basis of the conodont zones (Ouyang & Utting 1990). In Pakistan, the Chhidru Formation, considered to be Changxingian on the basis of conodonts (Wardlaw & Pogue 1995), yielded a palynological assemblage characterized by *L. virkkiae*, *D. playfordii* and *R. chalastus* (Balme 1970). Similar microflora have been recognized in other northern Gondwana regions (e.g. Oman, Saudi Arabia, southwestern Turkey and Iran), as well as

L. virkkiae, *R. chalastus* and *T. opaqua*, which characterize the OSPZ5 and OSPZ6 biozones of the Guadalupian (e.g. Stephenson *et al.* 2003; Stolle *et al.* 2011; Spina *et al.* 2018b). Regarding the finding and chronostratigraphic value of *R. chalastus*, which characterizes the entire palynoassemblage of the Le Cetine Formation and is abundantly present in both the Mt Calamita and Rio Marina formations, it is remarkable that its origin was alternatively interpreted as fungal, algal or animal (e.g. Balme 1970; Eshet *et al.* 1995; Foster *et al.* 2002; Spina *et al.* 2015; Rampino & Eshet 2018). The consideration of *R. chalastus* as a chronostratigraphic marker for the worldwide correlation of the Permian–Triassic boundary, as a reflection of environmental change across this boundary (Eshet *et al.* 1995; Visscher *et al.* 1996; Steiner *et al.* 2003), has been discussed by several researchers (e.g. Foster *et al.* 2002; Utting *et al.* 2004; Spina *et al.* 2015). In particular, Foster *et al.* (2002) showed that *R. chalastus* proliferated outside the hypothesized time of the mass extinction event, ranging from Wordian to Induan, thus making it clear that *R. chalastus* cannot be considered a key marker for the Permian period. The finding of such chronostratigraphic elements provides robust constraints for a clear chronological definition of these highly debated key formations of the Tuscan Paleozoic succession, in contrast with the different and various correlations proposed in the past.

The best example of the difficulties in the chronological attributions derives from the Mt Calamita Formation and its black phyllite, analysed in this work and previously described by Garfagnoli *et al.* (2005) as Lithofacies C. This was correlated, on the basis of their lithological similarities and the occurrence of black quartzite, with the Silurian–Early Devonian units of Tuscany, such as the *Orthoceras*-bearing dolostone and lydite of the Apuan Alps (Conti *et al.* 1991) and the Risanguigno Formation (Costantini *et al.* 1988). In addition, Garfagnoli *et al.* (2005) highlighted that similar lithotypes rest on the Ordovician Porphyroids and Porphyritic Schists cropping out in the east of Elba Island (Duranti *et al.* 1992; Bortolotti *et al.* 2001). By contrast, based on U/Pb dating on detrital zircons, the Mt Calamita Formation was considered as early Carboniferous by Musumeci *et al.* (2011), whereas Sirevaag *et al.* (2016) indicated a maximum depositional age in the Cambrian (c. 520 Ma; Fig. 4).

The chronological attribution of the Rio Marina Formation has been widely debated in the geological literature and, based on lithological correlation and different and scattered fossil findings, spans from the late Carboniferous to the early Permian (Vai 1978). The Early Permian (c. 280 Ma) is assumed to be the maximum depositional age by Sirevaag *et al.* (2016) and Paoli *et al.* (2017) on the basis of U–Pb radiometric ages on three detrital zircons (Fig. 4). The microfloristic evidence presented here does not contradict the radiometric data from detrital zircons, allowing us to relate the Rio Marina Formation to a younger age, such as the mid- to late Permian.

The previous stratigraphic correlations for the Le Cetine Formation were conditioned by the uncertainty in the chronostratigraphic attribution, which ranges from the late Carboniferous to Triassic (Cocozza *et al.* 1974; Bagnoli *et al.* 1978, 1979; Puxeddu *et al.* 1979; Costantini *et al.* 2009). After this palynological study, and considering the lithological characteristics, we propose that the Le Cetine Formation is also correlated to the Permian Arenarie di Poggio al Carpino Formation (Lazzarotto *et al.* 2003), in agreement with Cocozza *et al.* (1974). The correlation based on palynology also allows a wider relationship with a number of Paleozoic successions in southern Tuscany, characterized by similar lithotypes and chronologically attributed to the same time span. We refer to the sedimentary succession drilled at depth by several boreholes in the geothermal area of southern Tuscany (the Amiata Mt area, Bagnore 20 and PC26, Bagnoli *et al.* 1980; PC30, Pandeli *et al.* 1988; Brogi 2008), generally characterized by metapsammite and metapelite,

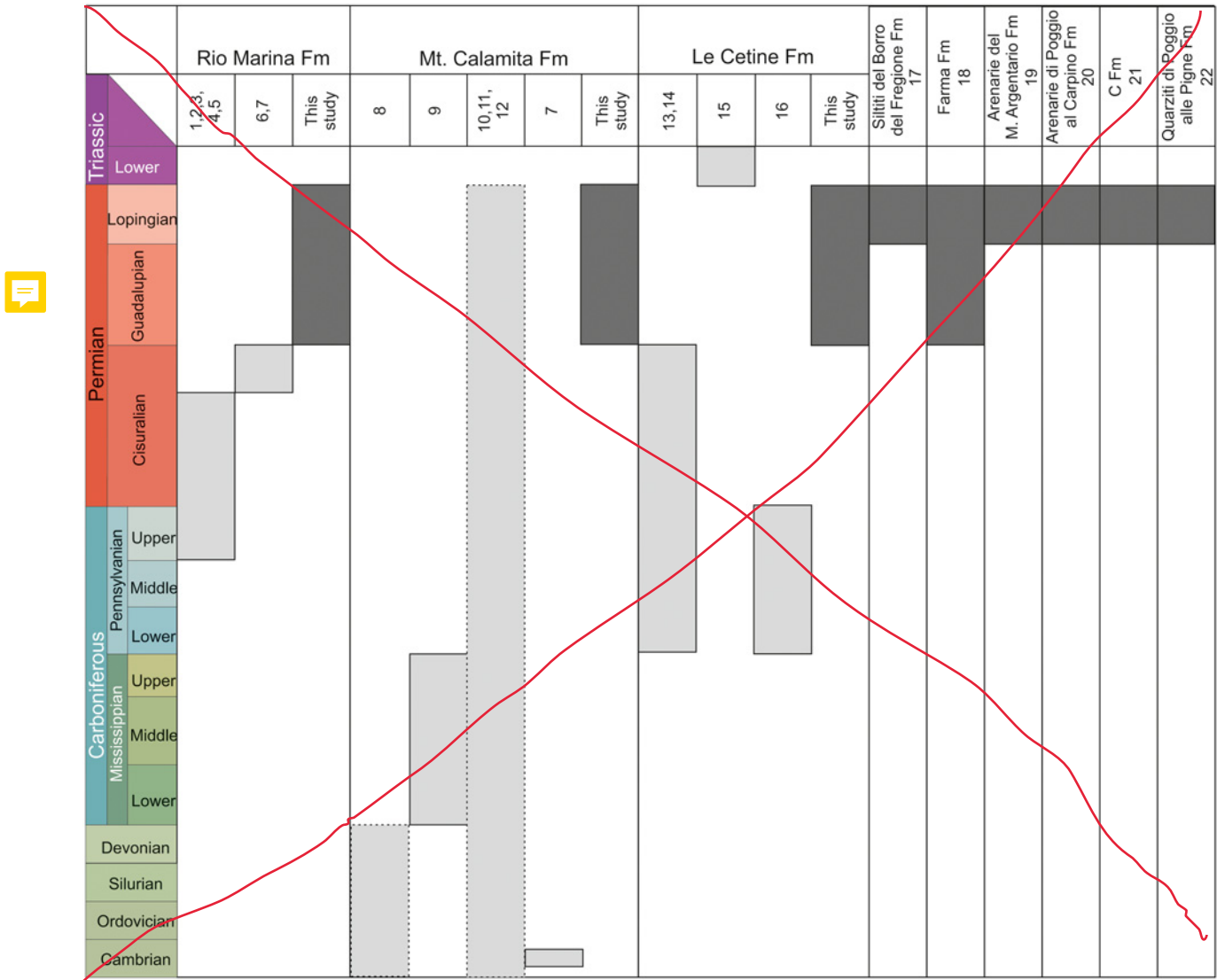


Fig. 4. Stratigraphic correlation chart for the different Palaeozoic formations of southern Tuscany. Numbers indicate references of previous attributions: 1, Bodechtel (1964); 2, Coccozza *et al.* (1974); 3, Vai (1978); 4, Pandeli & Puxeddu (1990); 5, Bortolotti *et al.* (2001); 6, Paoli *et al.* (2016); 7, Sirevaag *et al.* (2016); 8, Garfagnoli *et al.* (2005); 9, Musumeci *et al.* (2011); 10, Principi *et al.* (2015); 11, Barberi *et al.* (1967); 12, Pandeli *et al.* (1994); 13, Bagnoli *et al.* (1978); 14, Bagnoli *et al.* (1979); 15, Puxeddu *et al.* (1979); 16, Costantini *et al.* (2009); 17, Pandeli (1998); 18, Aldinucci *et al.* (2008); 19, Cirilli *et al.* (2002); 20, Lazzarotto *et al.* (2003); 21, Pandeli & Pasini (1990); 22, Aldinucci *et al.* (2001).

enriched in organic materials and subdivided into three parts (the A Formation, B Formation and C Formation) by Pandeli *et al.* (1988). The C Formation is attributed to the late Permian by the occurrence of fusulinids (Pandeli & Pasini 1990). Similar considerations and correlations can also be performed with other formations already assigned to the late Permian by their stratigraphic, compositional features and fossil content (i.e. the Quarziti di Poggio alle Pigne Formation and the Siltiti del Borro del Fregione Formation of Pandeli (1998) and Aldinucci *et al.* (2001)), respectively). Consequently, the Tuscan mid- to late Permian formations might be framed in a common depositional unit (Fig. 4).

This is not the first finding of sporomorphs in Paleozoic Tuscan successions. Other formations cropping out in the Middle Tuscan Ridge (e.g. the Farma Formation, Aldinucci *et al.* 2008; the Arenarie del Mt Argentario Formation, Cirilli *et al.* 2002; and the Arenarie di Poggio al Carpino Formation, Lazzarotto *et al.* 2003) showed a similar microfloristic content, implying a wider correlation that we now renew and base on a more robust, biostratigraphic dataset. Conversely, the different correlations proposed previously (Fig. 4) were interpreted as deriving from different palaeodepositional marine environments (e.g. from open

sea flysch deposits for the Farma Formation to epicontinental/prograding delta/shallow marine for the Arenarie del Monte Argentario Formation and the Arenarie di Poggio al Carpino Formation), providing evidence for the presence of a wide and articulated palaeoenvironmental setting developed after the Hercynian orogeny. Independent depositional environments are related to SW–NE-oriented epicontinental basins developed in tectonic depressions where fan deltas settled up, and turbidity flows occurred. A similar scenario agrees with pull-apart basins in extensional (transtensional?) settings, as reconstructed for the Permian evolution of Corsica and Provence (Rau 1990; Cortesogno *et al.* 1998). In this framework, and according to Cirilli *et al.* (2002), the Permian sediments of the Northern Apennines could be interpreted as the first products of the marine transgression marking the beginning of the Alpine Tethyan cycle.

An important reverberation of the microfloristic findings reported here and their related age concerns the correlation with the Laurasian plate, to which the Tuscan metasedimentary successions were previously and lithostratigraphically correlated. These successions, together with the coeval formations cropping out in Sardinia (e.g. Pandeli *et al.* 1994), were considered to be deposited along the

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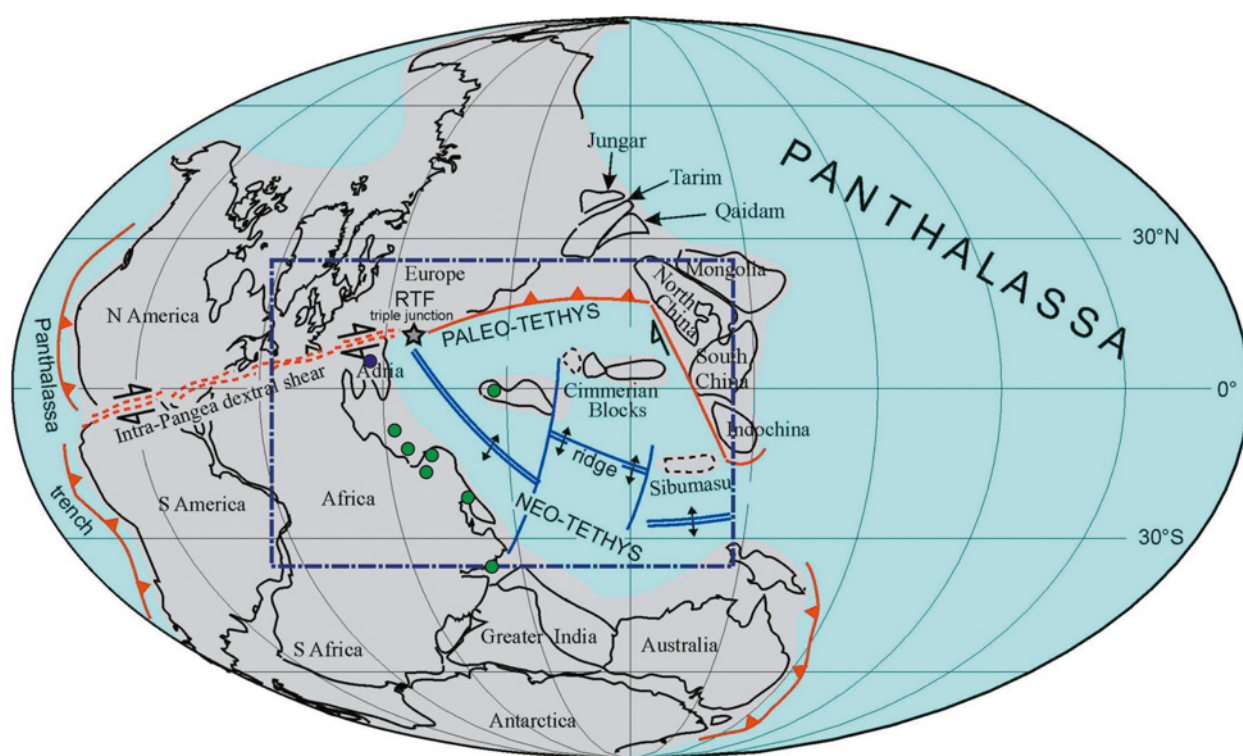


Fig. 5. Occurrence of *Thymospora opaqua* in the northern Gondwana region and in the present study in the Guadalupian–Lopingian. See text for references. Palaeogeographical map at the time of the Permian–Triassic boundary (c. 251 Ma) from Berra & Angiolini (2014).

southern part of the Hercynian chain (the Sardinia–Corsica block; Elter & Pandeli 2005). However, after this new chronostratigraphic framework, a microfloristic comparison is no longer possible as a result of the sedimentary gap in Sardinia between the mid- to late Permian (post-Sakharmanian) acidic volcanism and the reddish clastic sequence of the Early Triassic (Pittau et al. 2008). Consequently, the mid- to late Permian Tuscan deposits cannot be related to Laurasia and must be referred to a different palaeogeographical setting/terrene.

A possible new scenario is suggested by the presence of the typical Gondwana element *T. opaqua* (mid-Permian), together with cosmopolitan taxa such as *L. virkkiae* and *K. schaubergerii* (Stephenson 2018), documented herein for the first time. These findings account for a Gondwana affinity for the studied metasedimentary sequences.

Until now, *T. opaqua* showed a limited palaeogeographical distribution in the Guadalupian and/or Lopingian across most of the margin of northern Gondwana (e.g. Oman and Saudi Arabia, Stephenson et al. 2003; the Zagros Basin in Iran, Spina et al. 2018b; the Alborz Basin, Châteauneuf & Stampfli 1978; southwestern Turkey and northern Iraq, Stolle 2007; Stolle et al. 2011; and Pakistan, Jan et al. 2009; Fig. 5). Insights into the presence of Gondwana-related sediments in the Tuscan Paleozoic succession were also highlighted by Sirevaag et al. (2016) based on zircon of possible northern Gondwana origin observed in the Rio Marina Formation, but considered as probably recycled through erosion from the European Hun terranes rather than sourced from Africa.

Conclusions

Our results introduce new constraints to better define the age of the Paleozoic units in the inner Northern Apennines and suggest a new palaeogeographical and tectonic scenario.

(1) The Rio Marina Formation and the Le Cetine Formation can be referred to the mid- to late Permian, similar to the upper, younger,

part of the Mt Calamita Formation, the deepest outcropping structural unit of the Northern Apennines.

(2) A number of Paleozoic formations of the inner Northern Apennines can be correlated and attributed to Guadalupian–Lopingian age. This is an important constraint and sheds light on the geodynamic evolution of the pre-Triassic deposits.

(3) Taking into account previously published work and the results of this study, the Mt Calamita Formation includes rocks ranging from pre-Cambrian to mid- to late Permian, making this formation worthy of further detailed study.

(4) The commonly proposed lithostratigraphic correlation between the Paleozoic Tuscan and Sardinian successions has to be reconsidered because the mid- to late Permian (post-Sakharmanian) sedimentary gap in the Sardinia successions is coincident with the deposition of the Tuscan succession.

(5) The first occurrence of Gondwana-related sporomorphs suggests an innovative palaeogeographical scenario of late Paleozoic successions from the northern Gondwana margin.

(6) Correlation with other previously studied mid- to late Permian successions leads us to consider that the fragmentation of the northern margin of Gondwana was the result of several transtensional (pull-apart) basins, in which different marine to continental, laterally related depositional environments led to the sedimentation of these Tuscan mid- to late Permian formations.

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Appendix A: list of taxa

?Lundbladispora sp.Alisporites nuthallensis

Alisporites sp.

Calamospora sp.

Corisaccites alutas Venkatachala and Kar 1966

Cycadopites sp.

Cyclogranisporites sp.

Densoisporites playfordii Balme 1970

Densoisporites sp. cf. playfordii Balme 1970

Florinites sp.

Klausipollenites schaubegeri (Potonié and Klaus) Jansonius 1962

Kraeuselisporites sp.

Laevigatosporites sp.

Lueckisporites virkkiae Potonié and Klaus 1954

Lundbladispora sp.

Potonieisporites sp.

Reduviasporonites chalastus (Foster) Elsik 1999

Taeniaesporites sp.

Thymospora opaqua Singh 1964

Vittatina costabilis Wilson 1962

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