



The mid-Langhian flooding in the eastern Central Paratethys: integrated stratigraphic data from the Transylvanian Basin and SE Carpathian Foredeep

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Abstract

The mid-Langhian (“Badenian”) flooding fully reconnected the semi-isolated Central Paratethys realm with the Mediterranean and, thereby, drastically changed the middle Miocene paleogeography of Central Europe. Due to the scattered stratigraphic record and scarcity of independent age constraints in some areas, the precise age and underlying mechanism are still debated. We present integrated chronostratigraphic data from five sections in the eastern part of the system to reconstruct the flooding event distal from the strait to the Mediterranean. By applying modern Mediterranean biochronology (planktonic foraminifera and calcareous nannofossils), supplemented by an ⁴⁰Ar/³⁹Ar age on a tuff, we demonstrate that the widespread open marine settings in the NW Transylvanian Basin were definitely installed after 14.9 Ma (MMi4d biozone), and in most areas before 14.4 Ma. In the marginal study area in the SE Carpathian Foredeep, fully marine conditions likely set in slightly later (14.6–14.4 Ma). There, short-lived marine incursions into the brackish environment occurred since the latest Burdigalian (“pre-flooding phase”). The new ages overlap with the flooding in the majority of the Central Paratethys (~14.9–14.4 Ma), and with marine overflow into the Black Sea (14.85 Ma). We suggest that the transgression was driven by subsidence of the Pannonian Basin, by creating accommodation space and diminishing barriers between sub-basins, but was likely enhanced by a global sea-level rise. Finally, we speculate that the scarcity of all calcareous material in the SE Carpathian Foredeep before the mid-Langhian flooding might be related to pulses of nutrient-rich brackish and low pH water from the neighboring Black Sea Basin.

Keywords Integrated stratigraphy · Radioisotopic dating · Foraminifera · Nannofossils · Miocene paleogeography

Introduction

During the Miocene, a large part of Central Europe was covered by the Paratethys Sea (Popov et al. 2004; Rögl 1998). In this period, active geodynamic processes, such as the opening of the Pannonian back-arc system, exhumation of the Alps–Carpathians–Dinarides orogens and associated volcanic episodes and opening/closure of sea straits resulted in a dynamic paleogeography (e.g., Fodor et al. 1999; Pécskay et al. 2006; Kováč et al. 2017b; Matenco 2017). As a result, the paleoenvironments of the Central Paratethys basins shifted between open marine, brackish and lacustrine settings (de Leeuw 2011; Piller et al. 2007; Rögl 1998), and its sedimentary record reflects changes in tectonics, eustatics, climate and basin connectivity (Kováč et al. 2017a; Palcu et al. 2015).

Correlating the Paratethys strata to regional and global events, however, is often challenging, because the

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stratigraphic scheme of the Central Paratethys consists of regional stages that are generally based on endemic fossil assemblages and correlations to sequence stratigraphic records (Haq et al. 1987; Hardenbol et al. 1998; Hilgen et al. 2012). Nevertheless, numerous independent age constraints have been produced by integrated stratigraphic methods in the last decade (e.g., Mandić et al. 2012; de Leeuw et al. 2013; Palcu et al. 2015; Sant et al. 2018). These allowed quantitative reconstructions of major tectonic and climatic events affecting the whole area, such as the Badenian Salinity Crisis (de Leeuw et al. 2010), the Badenian-Sarmatian Extinction Event (Harzhauser and Piller 2004; Palcu et al. 2015) and late Miocene installation of Lake Pannon (Filipescu et al. 2011; Magyar et al. 1999; ter Borgh et al. 2013; Vasiliev et al. 2010).

Progress was also made in constraining the timing of the widespread marine “Badenian” transgression (middle Miocene) of the Pannonian Basin and Carpathian Foredeeps (Bukowski et al. 2018; Ćorić et al. 2009; Hohenegger et al. 2009; Kováč et al. 2017a, 2018; Selmečzi et al. 2012). A two-step flood has been proposed (Kováč et al. 2007, 2017a): a first Badenian flooding, occurring in the early Langhian, and second flooding, occurring in the middle Langhian. The first flooding only had a small impact on the Central Paratethys—installing semi-marine environments with occasional marine fauna influxes. The Slovenian corridor probably functioned as the connecting sea strait (Bistricic and Jenko 1985; Sant et al. 2017). Central Europe became covered by a very diverse marine-brackish-lacustrine system (Brzobohatý et al. 2003; Kováč et al. 2003; Mandić et al. 2012; Rijavec 1985; Zuschin et al. 2014) (Fig. 1b).

The second Badenian flooding turned the Central Paratethys into an open marine gulf of the Mediterranean,

by increasing the connection through the Slovenian strait (Brzobohatý et al. 2007; Kováč et al. 2007; Rögl 1998) (Fig. 1a). This event also impacted the Eastern Paratethys, where a short-lived marine episode, the Tarkhanian marine flooding, took place at around 14.85 Ma (Palcu et al. 2017). The onset of the widespread open marine “Badenian Sea”, associated with the second Badenian flooding, is recognized by an abrupt shift to finer-grained sediments enriched in Langhian microfossils, most importantly planktonic foraminifera *Praeorbulina glomerosa* and *Orbulina suturalis* (Kováč et al. 2017a; Sant et al. 2017), and in most regions, nannofossils of the NN5 (Martini 1971) biozone (Papp et al. 1978; Kováč et al. 2007 and references therein) (Table 1). Following the most recent low latitude (Mediterranean) biochronology (Table 1), these Langhian marker fossils suggest a maximum age of ~15.3 Ma for the second Badenian flooding that installed the open marine Badenian Sea (Kováč et al. 2017a; Sant et al. 2017).

The timing and paleoenvironmental changes during the Badenian floodings in the Romanian Carpathian Foredeep and Transylvanian Basin (4 and 6 in Fig. 1) are still unclear. These basins are key areas to reconstruct the effect of these floodings far away from the Slovenian strait and proximal to the Eastern Paratethys (Black Sea basin) strait.

The scarcity of integrated chronostratigraphic data, outdated terminology and/or contradicting interpretations hamper a proper reconstruction of the late Burdigalian–Langhian stratigraphy in the Romanian basins. For instance, the flooding boundary in the Transylvanian Basin is commonly placed at ~16 Ma, whereas directly overlying tuffs (Dej tuff) are dated to be ~14.4 Ma (Szakács et al. 2012; de Leeuw et al. 2013; Beldean et al. 2013). Traditionally,

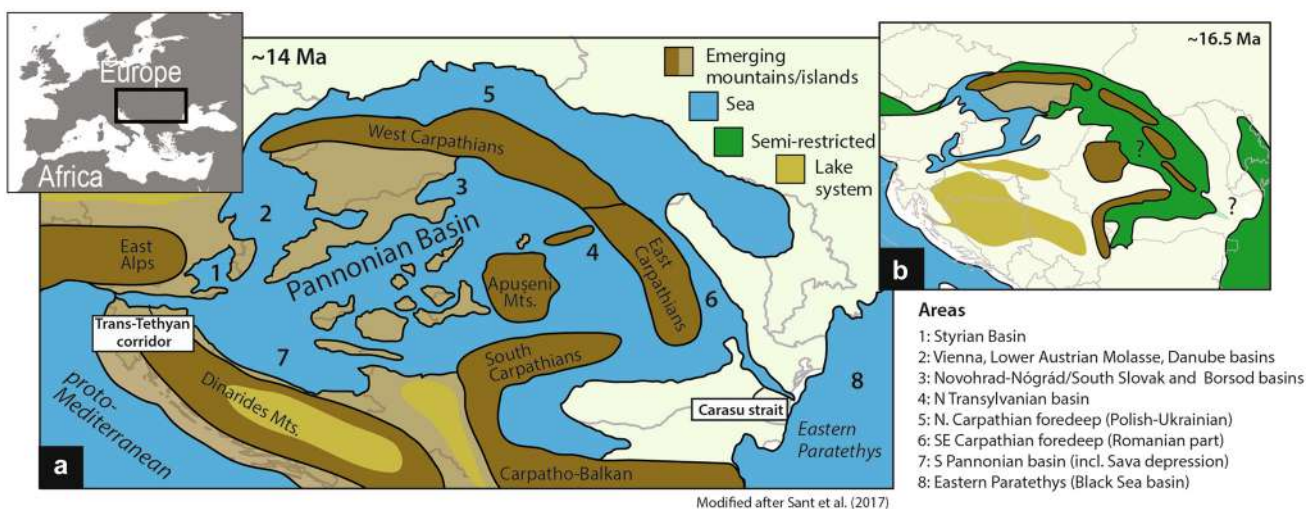


Fig. 1 Simplified facies maps of the Badenian Sea (a) and pre-flooding phase (late Burdigalian) (b). Key areas are numbered. Modified after Sant et al. (2017)

Table 1 Dated planktonic foraminifera and calcareous nannofossils (bold) marker events for the oceanic and Mediterranean biostratigraphic schemes

Oceanic biozones					
Basal marker	Biozone	Author (zone)	Age (Ma)	Author (Age)	
LO <i>S. heteromorphus</i>	CNM8	Backman et al. (2012)	13.53	Backman et al. (2012)	
FO <i>F. robusta</i>	M9b	Wade et al. (2011)	13.13	Lourens et al. (2004)	
FO <i>F. fohsi</i>	M9a	Wade et al. (2011)	13.41	Lourens et al. (2004)	
FO <i>F. "praefohsi"</i>	M8	Wade et al. (2011)	13.77	Lourens et al. (2004)	
FO <i>F. peripheroacuta</i>	M7	Wade et al. (2011)	14.24	Lourens et al. (2004)	
FO <i>O. suturalis</i>	M6	Wade et al. (2011)	15.10	Wade et al. (2011)	
FO <i>D. signus</i>	CNM7	Backman et al. (2012)	15.73	Backman et al. (2012)	
FO <i>P. glomerosa</i> s.s.	M5b	Wade et al. (2011)	16.27	Lourens et al. (2004)	
FO <i>P. sicana</i>	M5a	Wade et al. (2011)	16.38	Wade et al. (2011)	
FO <i>F. birnageae</i>	M4b	Wade et al. (2011)	16.69	Wade et al. (2011)	
LO <i>C. dissimilis</i>	M4a	Wade et al. (2011)	17.54	Lourens et al. (2004)	
FO <i>S. heteromorphus</i>	CNM6	Backman et al. (2012)	17.75	Backman et al. (2012)	
FO <i>T. carinatus</i>	NN3	Martini (1971)	19.18	Pälike et al. (2005)	
LO <i>S. belemnus</i>	NN4	Martini (1971)	17.94	Backman et al. (2012)	
LO <i>H. ampliaperta</i>	NN5	Martini (1971)	14.86	Curry et al. (1995)	
LO <i>S. heteromorphus</i>	NN6	Martini (1971)	13.53	Backman and Raffi (1997)	
Mediterranean biozones					
LO <i>G. peripheroronda</i>	MMi6	Iaccarino AND Salvatorini (1982)	13.52	Mourik et al. (2011)	
LCO <i>S. heteromorphus</i>	MNN6a	Fornaciari et al. (1996) emended Di Stefano et al. (2008)	13.65	Lourens et al. (2004)	
FO <i>G. praemenardii</i>	MMi5c	Iaccarino and Salvatorini (1982)	13.92	Abdul Aziz et al. (2008)	
FCO <i>H. walbersdorfensis</i>	MNN5c	Fornaciari et al. (1996) emended Di Stefano et al. (2008)	14.05	Abdul Aziz et al. (2008)	
LCO <i>H. waltrans</i>	MNN5b	Di Stefano et al. (2008)	14.36	Abdul Aziz et al. (2008)	
FO <i>O. universa</i>	MMi5b	Iaccarino and Salvatorini (1982)	14.36	Abdul Aziz et al. (2008)	
FO <i>O. suturalis</i>	MMi5a	Cita and Premoli Silva (1971–1973)	14.56	Abdul Aziz et al. (2008)	
FO <i>P. gl. circularis</i>	MMi4d	Iaccarino and Salvatorini (1982)	14.89	Abdul Aziz et al. (2008)	
FO <i>P. gl. curva</i>	MMi4c	Di Stefano et al. (2008)	15.34	Turco et al. (2017)	
PE <i>S. heteromorphus</i>	MNN5a	Di Stefano et al. (2008)	15.37	Turco et al. (2017)	
AaB <i>P. siakensis</i>	MMi4b	Di Stefano et al. (2008)	15.48	Turco et al. (2017)	
PB <i>S. heteromorphus</i>	MNN4c	Fornaciari et al. (1996) emended Di Stefano et al. (2008)	15.74	Turco et al. (2017)	
LCO <i>H. ampliaperta</i>	MNN4b	Fornaciari et al. (1996) emended Di Stefano et al. (2008)	16.09	Turco et al. (2017)	
FO <i>G. sicanus</i> (3 apertures)	MMi4a	Iaccarino and Salvatorini (1982) emended Iaccarino et al. (2011)	16.18	Turco et al. (2017)	
LO <i>C. dissimilis</i>	MMi3	Iaccarino and Salvatorini (1982)	17.54	Lourens et al. (2004)	
FCO <i>S. heteromorphus</i>	MNN4a	Di Stefano et al. (2008)	17.99	Foresi et al. (2014)	
LCO <i>S. belemnus</i>	MNN3b	Fornaciari and Rio (1996)	18.44	Foresi et al. (2014)	
FCO <i>S. belemnus</i>	MNN3a	Fornaciari and Rio (1996)	19.05	Foresi et al. (2014)	

Mediterranean calcareous plankton assemblages differ from the Atlantic low-latitude ones from the middle Miocene onwards, so their bio-events have significantly different ages (see also Di Stefano et al. 2008, 2015; Foresi et al. 2011; Iaccarino et al. 2011; Turco et al. 2002)

Calcareous nannoplankton marker species are marked bold

FO first occurrence, FCO first common occurrence, LO last occurrence, LCO last common, occurrence, PB paracme beginning, PE paracme end, AaB acme a base

sediments positioned below the open marine Langhian strata are assigned to the Burdigalian stage, but in some cases part of these underlying sediments turned out to

have a younger age corresponding to the early Langhian (MăruŃeanu 1999; Melinte-Dobrinescu and Stoica 2014; Orbocea et al. 1986a; Rabagia and MaŃenco 1999; Sant et al.

2017). Paleontological studies commonly use the regional Paratethys taxonomy, which is difficult to compare to the Mediterranean and/or low latitude taxonomy. In addition, the use of ambiguous formation names and/or regional fossil assemblages may result in a variable and contradictory correlations (Melinte-Dobrinescu and Stoica 2014; Săndulescu 1988; Mărunțeanu 1999; Beldean et al. 2012; de Leeuw et al. 2013; Székely et al. 2016). All of the factors above complicate correlation of the stratigraphy from different regions to each other, and to the global time scale (Paučá 1976; Krézsek and Bally 2006 and references therein; Beldean et al. 2013; Gozhyk et al. 2015; Holcová et al. 2018; Kováč et al. 2018) hampering paleogeographic and paleoenvironmental reconstructions of the whole Central Paratethys system.

Here, we reconstruct the timing and environmental changes of the Langhian (Badenian) transgression in the Transylvanian Basin and SE Carpathian Foredeep by integrating magnetostratigraphic and radioisotopic data with biostratigraphic results. We apply the most recent Mediterranean biochronology for calcareous nannofossils (MNN-zones of Fornaciari and Rio 1996; Di Stefano et al. 2008; Iaccarino et al. 2011) and planktonic foraminifera (MMi-zones of Di Stefano et al. 2008; Iaccarino et al. 2007, 2011) for dating (Table 1). This approach usually results in younger ages compared to studies using oceanic biozones and biochronology (Sant et al. 2017). Furthermore, we will investigate the role that marine straits, eustatic fluctuations and tectonic changes play in the distribution of the marine system by integrating the new data with (bio)stratigraphic and paleogeographic results from other Paratethys localities.

Geological background

Basins and mountains of the Pannonian-Carpathian region

The Pannonian back-arc basin formed in response to the rapid rollback of a Carpathian slab attached to ‘stable Europe’ during the Miocene (~18–9 Ma) (e.g., Fodor et al. 1999; Horvath et al. 2006; Matenco and Radivojević 2012). In the Pannonian Basin, a chain of (half-)grabens formed during a diachronous rifting phase with a peak in the middle Miocene (Balázs et al. 2016; Nagymarosy and Hámor 2012; Tari et al. 1992). In the eastern part (Tisza-Dacia terranes), the extensional process lasted until the late Miocene (Balázs et al. 2016).

The Transylvanian Basin was a satellite basin of the Pannonian basin since the Miocene (4 in Fig. 1) and is at present surrounded by the Apuseni mountains and E, SE and S Carpathians. The majority of the Miocene sediments is situated on top of the Tisza-Dacia terrane (Schmid et al.

2008; Ustaszewski et al. 2008). In the lower Miocene, it was a flexural foredeep basin created by the emplacement of the Pienides thrust nappes over the northern edge of the Transylvanian embayment (Krézsek and Bally 2006; Tischler et al. 2008). In contrast to the Pannonian Basin, the majority of middle-late Miocene subsidence in the Transylvanian Basin was most likely driven by a deep (thermal) lithospheric mechanism, and only a fraction of the sediment infill is associated with early middle Miocene crustal faults (Tiliță et al. 2013 and references therein). The eastern flank of the Transylvanian Basin is covered by a thick pile of Miocene-Pliocene volcanics (Mason et al. 1998; Szabó et al. 1992). At present, the basin is the most important province for gas production in Romania and its surroundings (Ciulavu et al. 2000).

The East Carpathian Foredeep circumvents the eastern and south-eastern flanks of the E and SE Carpathians (6 in Fig. 1). Neritic (relatively deep) marine deposition took place in the ESE foredeep close to the Carpathian thrust front (Tărăpoancă 2003). Lower to middle Miocene sediments accumulated in shallow-marine to fluvio-deltaic settings in piggyback basins (Săndulescu et al. 1995), or in grabens between gradually uplifting Carpathian blocks (Matenco et al. 2010; Mirăuța 1967; Olteanu 1951). At present, most lower-middle Miocene sediments are incorporated into thrust and fold belts of the external nappes. A total shortening of ~160–220 km was accommodated in the East, SE and South Carpathians by thick- and thin-skinned nappe stacking during Cretaceous-Miocene moments of subduction (Morley 2003; Roure et al. 1993). Thermochronological studies infer that most of the topographic relief of the E and SE Carpathians formed during Miocene times (~15–11 Ma), whereas the bulk exhumation in the South Carpathians occurred during the late Cretaceous-Paleogene (Matenco 2017; Merten 2011). The Apuseni mountains were most likely almost completely exhumed since Oligocene times (e.g., Merten et al. 2011; Kounov and Schmid 2013; Reiser et al. 2017). Eventual Miocene uplift must have been below 1 km (Matenco 2017).

Connectivity between the Transylvanian Basin, Carpathian Foredeep and Pannonian Basin is uncertain due to the removal of Miocene sediments during the uplift of the Carpathians, the burial of lower-middle Miocene strata under volcanic rocks, and a general lack of age constraints. Connections between the Transylvanian Basin and the SE Carpathian Foredeep likely existed at least until Langhian times (Matenco 2017; Merten 2011; Tărăpoancă 2003), because patches of such Langhian sediments are nowadays found in the Carpathian mountains (Codarcea et al. 1968).

Early middle Miocene stratigraphy of the Romanian Paratethys

In the regional Central Paratethys time scale, the lower Miocene is subdivided into the Egerian, Eggenburgian, Ottnangian and Karpatian stages (Cicha and Rögl 2003; Hilgen et al. 2012; Papp and Cicha 1978; Piller et al. 2007). The middle Miocene is divided into the Badenian and Sarmatian stages (Papp and Cicha 1978; Papp and Steininger 1974) (Fig. 2). Local stages are used in many Romanian maps and old literature. This can be confusing since the terms Chat-tian-Burdigalian (ch-bu), Helvetian (he) and Tortonian (to) are roughly referring to the late Oligocene-early Miocene, Burdigalian–earliest Langhian, and early middle Miocene time intervals, respectively (e.g., Ilie 1951; Marinescu and Peltz 1967). In this manuscript, we only use the term Burdigalian (20.4–16.1 Ma) *sensu* Hilgen et al. (2012).

Stratigraphy of the Romanian East Carpathian Foredeep

A significant portion of the lower-middle Miocene strata of the Romanian East Carpathian Foredeep is incorporated into the Tarcău, Vrancea and Subcarpathian Carpathian nappes (so-called external nappes) (Juravle 2010; Mărunțeanu 1999; Săndulescu et al. 1980). The large variety in sedimentary settings resulted in a wide array of formations with different thicknesses in different areas. Figure 2 depicts the generalized stratigraphy for two areas in the Romanian East Carpathian Foredeep, namely for the most external nappe (Subcarpathian), and for a more restricted piggyback basin in the SE Carpathian bend area. Detailed stratigraphic background data can be found in Supplementary Data A and Sant et al. (2019).

In general, the lower Miocene stratigraphy begins with an evaporitic package followed by sands, silts and marls. The subsequent grey schlier formation is dominated by marls with silt and sand interbeds and a few tuff horizons (e.g., Ioniță 1962; Marunteanu 1999). In most areas, a significant evaporitic interval (10–25 m thick; Perchiu gypsum) is noted in the basal part (Mărunțeanu 1999), while gypsum levels are common in the uppermost grey schlier Fm as well (Mărunțeanu 1999; Săndulescu et al. 1980). In the SE Carpathian bend area, the more or less time equivalent strata are termed the Doftana Fm (Săndulescu et al. 1995; Stefanescu et al. 2006) (Fig. 2). The marginal areas of the SE Carpathian Foredeep, such as the SE Carpathian Bend area are represented mostly by shallow marine-to-brackish environments with tiny and fragile marine microfossils and occurrences of ostracods and characean algae indicating fluctuating salinities. Orbocea et al. (1986b) suggest salinities comparable to the present-day Black Sea (~18 PSU).

The lower parts of the Grey Schlier and Doftana Fms (i.e. the pre-flooding sediments) are attributed to the NN4

calcareous nannofossils biozone (Martini 1971) corresponding to the lower Miocene portion of the uppermost CNM5 and CNM6 zones (Backman et al. 2012) (Table 1). Their upperparts are characterized by scarce *P. glomerosa* and/or *O. suturalis* and nannofossils of the NN5 Biozone, testifying of a middle Miocene (Langhian) age for their upper parts (Mărunțeanu 1999; Orbocea et al. 1986a; Săndulescu et al. 1995; Ștefănescu and Mărunțeanu 1980).

The base of the Badenian flooding is clearly visible by a shift to marls with a rich Langhian planktonic foraminiferal assemblage (*Globigerina* Marls). Above, the marls are intercalated with ash layers of the Slănic Tuff, that range from 0.2 to 3 m in thickness (rarely up to ~100 m) (Mărunțeanu 1999; Săndulescu et al. 1995). The Slănic tuff is chemically correlated to the Dej tuff in the Transylvanian Basin and the Novoselits-Danilov tuff in the Ukrainian Carpathian Foredeep (Pishvanova 1969) but was never dated by isotopic methods. The overlying Upper Salt Formation, with its base at the Langhian-Serravallian boundary, testifies of the Badenian Salinity Crisis (de Leeuw et al. 2010; Túnyi et al. 2005) and correlates to the Transylvanian salts. It is characterized by evaporitic lithologies in various forms, including salt breccia within marls, sands, halite among black shales and packages of gypsum (Peryt 2006).

Stratigraphy of the N Transylvanian Basin

The lower Miocene marine (to semi-restricted) sediments in the Transylvanian Basin thicken northwards towards the Pienides nappes (Ciulavu et al. 2000; Krézsek and Bally 2006; Tischler 2005). The southern Transylvanian Basin was dominated by marginal marine and continental sediments (Ilie 1955; Krézsek et al. 2010). The lower Miocene succession is mainly represented by the Vima, Coruș, Chechiș and Hida formations having highly diachronous boundaries and reflecting deep to shallow basinal settings (Filipescu 2011; Krézsek and Bally 2006) (Fig. 2). In the last decade, a series of high-resolution biostratigraphic studies has been carried out (Beldean et al. 2010, 2012, 2013; de Leeuw et al. 2013; Filipescu and Beldean 2007; Filipescu and Silye 2008; Székely et al. 2016, 2017). More detailed stratigraphic background data is found in Supplementary Data A and Sant et al. (2019).

The pre-flooding sediments of the Hida Fm (Hofmann 1879; Koch 1900) represent a relatively deep phase of the pre-Badenian basin that is followed by a gradual shallowing from slope to proximal delta fan environment. Siliciclastic deposits dominate and range from turbidites yielding diverse planktonic and benthic foraminifera to coarse-grained delta fans in the top with agglutinated foraminifera and small planktonic foraminifera (Beldean et al. 2012; Iva et al. 1971; Popescu and Brotea 1994; Székely et al. 2016). An Aquitanian–Burdigalian age was

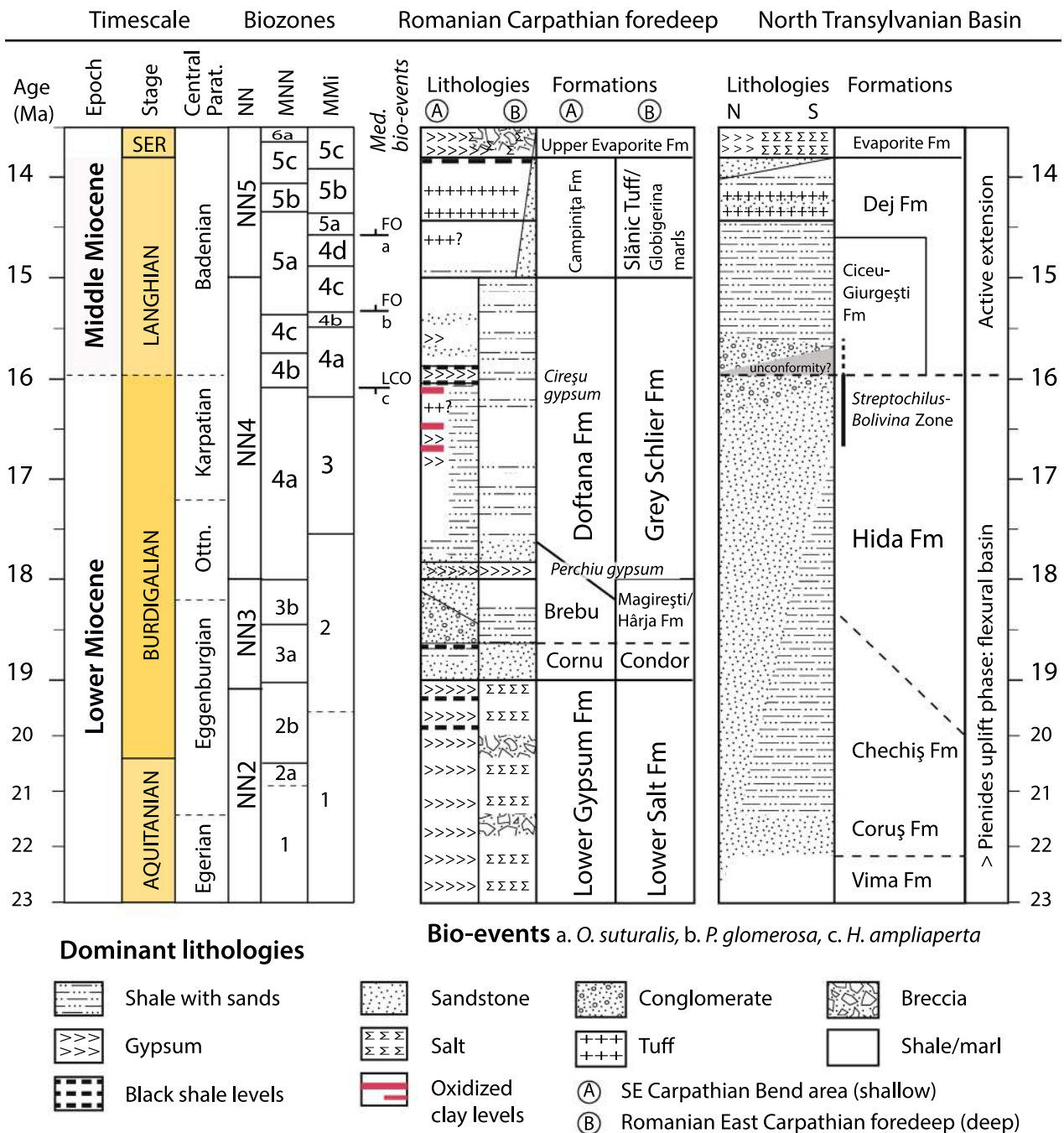


Fig. 2 An updated compilation of lower-middle Miocene stratigraphy and common formation names for the Romanian East Carpathian Foredeep (after Ioniță 1962; Măruțeanu 1999; Sandulescu et al. 1980; Sandulescu et al. 1995) and north Transylvanian Basin (Krészek and Bally 2006 and references therein). Note that the most recent low latitude (Mediterranean) biochronology is used to construct the age framework. The global and regional stages in the timescale are taken after Hilgen et al. (2012). The Mediterranean biozones and Atlantic NN-biozones are displayed for reference (see Table 1 for all references). The lower Miocene Fms in the Transylvanian Basin are diachronous and poorly dated, therefore, formation boundaries cannot be drawn. The Dej and Slănic tuffs have many synonyms,

a.o., Perșani tuff in the SE Transylvanian Basin (Rado et al. 1980), Racos tuff in the E Transylvanian Basin, Nereju tuff in the SE Carpathian Foredeep (Ioniță 1962), Govora tuff in the South Carpathians, and Novoselits-Danilov tuff in the Ukrainian Foredeep (Pishvanova 1969). The Evaporite Fm (Ciupagea et al. 1970; Dragoș 1969; Paucă 1976; Stoica and Gherasie 1981) is also known as the Ocna Dejului Fm in the NW Transylvanian Basin (Mészáros 1991). In the easternmost Foredeep the lowermost Grey Schlier Fm is known as the Teșcani Fm (Măruțeanu, 1999) but in other literature and/or regions it may refer to other intervals (Mirăuța 1967; Sandulescu et al. 1980, 1995). The term “Poiana marls” is used for the uppermost Magiurești Fm (Măruțeanu 1999; Olteanu 1951)

determined for the Hida Fm based on calcareous nannofossils marking the NN2–NN4 biozonal interval (Table 1; Fig. 2) (e.g., Mészáros 1991; Beldean et al. 2010, 2012). The uppermost Hida Fm is in many locations characterized by shallow coastal facies (e.g., Beldean et al. 2013). An unconformity is suggested at the top of the Hida Fm in the majority of the basins, which is most pronounced in the central and southern part (Krézsek and Bally 2006).

In some sections in the N Transylvanian Basin, an interval with a bloom in small-sized (63 μm fraction) biserial foraminifer *Streptochilus* was recorded in the top of the Hida Fm just below the Dej Fm (Beldean et al. 2013). Beldean et al. (2010) proposed to name it the *Streptochilus-Bolivina* Abundance Zone. Similar blooms occurred in the Atlantic and Indian oceans between 19 and 17 Ma, and are attested to the upwelling of cold nutrient-rich water or eutrophic conditions (Smart and Thomas 2006). Beldean et al. (2013) proposed that the *Streptochilus-Bolivina* Abundance Zone could have a similar age.

The Dej Formation is associated with a new transgression and the instalment of the widespread Badenian Sea (Kováč et al. 2017a; Sant et al. 2017). The basin axis migrated southwards, so the deposits cover a large area (Filipescu 1996; Krézsek et al. 2010; Popescu 1970). It is represented by a ~50–150 m thick carbonatic and siliciclastic package bearing a rich foraminiferal assemblage with *Praeorbulina glomerosa* at the base (Popescu 1970), followed by the appearance of *Orbulina suturalis*. These typical middle Miocene (Langhian) foraminiferas first occur in the Mediterranean at 15.34 Ma (*P. gl. curva*; *P. gl. glomerosa* is dated at 15.15 Ma) and 14.56 Ma, respectively, following the most recent Mediterranean biochronology (Table 1). At the western basin margin, bioclastic limestones are common (Filipescu and Gîrbacea 1997; Paucă 1951). The middle part of the Dej Fm contains tuffs, tuffites and volcanoclastic material of the Dej tuff complex, derived from a phase of explosive volcanism in the intra-Carpathians (Moisescu and Popescu 1967; Seghedi et al. 2004; Seghedi and Szakacs 1991). Its thickness varies between 1 and 120 m (Szakács et al. 2012). The tuff serves as an important marker of the basal middle Miocene succession because it is well visible in the field and in seismic profiles. Various authors dated the Dej tuff by $^{40}\text{Ar}/^{39}\text{Ar}$, K–Ar, zircon fission track and (U–Th)/He methods, resulting in a wide age range between ~16 and 13 Ma (Nicolescu and Mârza 2010; Seghedi and Szakacs 1991; Szakács et al. 2012). The most recent and most consistent ages are known from $^{40}\text{Ar}/^{39}\text{Ar}$ dating, providing ages of 14.38 ± 0.06 and 14.37 ± 0.06 Ma in the NW Transylvanian Basin (de Leeuw et al. 2013). The uppermost sediments of the Dej Fm are usually marine marls that were deposited deeper than 100 m (Filipescu 2011).

The Dej Fm is overlain by salt and gypsum that can be traced all over the Transylvanian basin (e.g., Ciupagea et al.

1970; Krézsek and Bally 2006). This phase reflects the Badenian Salinity crisis, a well-documented event in the Central Paratethys, of which the onset was radioisotopically dated at 13.8 Ma (Báldi et al. 2017; de Leeuw et al. 2010; Hryniv et al. 2007; Peryt 2000; Tünyi et al. 2005).

Methods

Sections and sampling

The Badenian transgression was studied in two sections in the Carpathian bend area in the Romanian Carpathian Foredeep (Campinița (CA) and Brebu (BR) sections) and in three sections in the northern Transylvanian Basin (Valea Dracului (DV), Ciceu-Giurgesti (CG) and Cepari (CP) sections) (Figs. 3, 4). They are described in detail in Sant et al. (2019). Besides the flooding boundary, the Campinița (155 m) and Brebu sections (~400 m) also cover a large pre-flooding interval.

Samples for micropaleontological and paleomagnetic analyses were taken with a resolution of ~1–2 m from the Valea Dracului, Ciceu-Giurgesti, Cepari and Campinița sections. The resolution is lower in intervals with coarse-grained material, such as sands and volcanic material. In addition, a detailed biostratigraphic sampling (25 cm resolution) was carried out along the flooding boundary of the CA and BR sections, labeled as CX and BX. A qualitative analysis of planktonic foraminiferal assemblages was carried out on the > 125 μm fraction of 88 washed residue samples focusing on biostratigraphic marker species at the Department of Chemistry, Life Sciences and Environmental Sustainability of the University of Parma. The samples containing benthic foraminifera (> 125 μm size fraction) were qualitatively examined and representative taxa were mounted on micropaleontology slides for reference. A total of 81 calcareous nannofossil smear slides were prepared and analyzed with a Zeiss Axioscope microscope under magnification 1000 \times at the BIOMLG Department of the University of Catania. The full description of the procedures are found in Sant et al. (2019).

A battery-powered electric drill cooled with water was used to take the total amount of 139 standard oriented paleomagnetic samples. Three tuff levels were sampled (> 1 kg per sample) for $^{40}\text{Ar}/^{39}\text{Ar}$ analyses; at 6 m in the Cepari section, at ~145 m in the Campinița section, and from the middle part of the Brebu section (marked by asterisk; Figs. 3, 4).

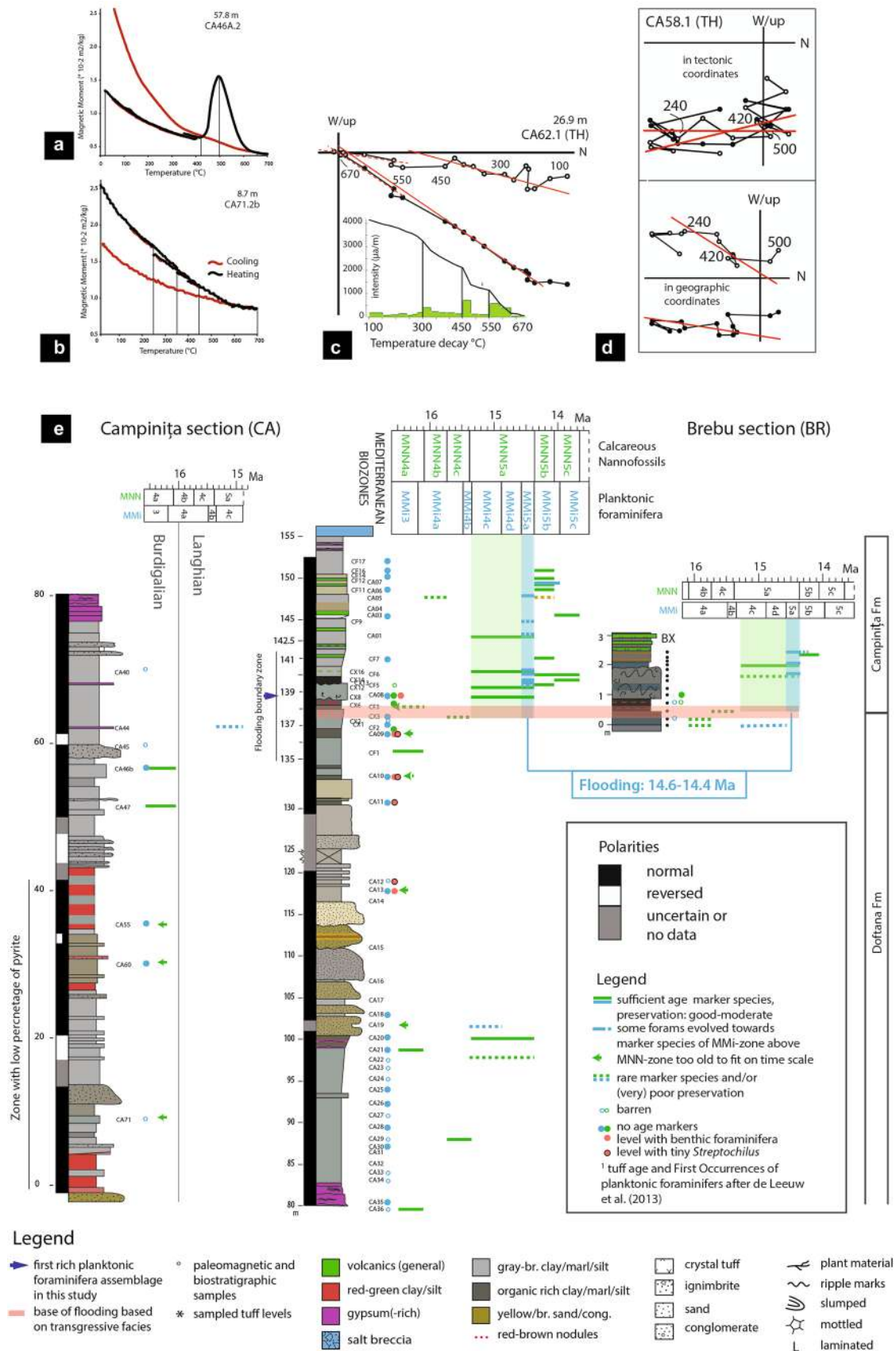


Fig. 3 Overview of paleomagnetic and biostratigraphic results of the sections in the SE Carpathian Foredeep (CA, BR). Demagnetization diagrams in W/up orientation (Zijderveld 1967) (**a**, **b**) and Curie diagrams (**c**, **d**). Red (dashed) lines indicate various resolved paleomagnetic vectors. *th* thermal demagnetization, *af* alternating field demagnetization. **a** Irreversible behavior between 250 and 450 °C points to thermal breakdown of an iron sulphide. **b** The large peak in magnetization around 430 °C reflects pyrite-to-magnetite conversion. Sample codes and levels are displayed in the upper right corners. **c** Includes a decay diagram with the green blocks indicating the total decay in intensity per temperature step. **d** The diagram in geographic coordinates yields a feasible paleomagnetic direction, while the one in tectonic coordinates does not make sense. **e** Synthesis of calcareous nannofossil (green), planktonic foraminifer (blue) and magnetostratigraphic results. The shifts to open marine conditions are marked by horizontal red bars. Vertical semi-transparent green and blue boxes mark the age ranges for the flooding boundaries based on nannofossils and foraminifera, respectively. Ages in green and blue reflect the maximum ages for the onset of the open marine transgression bases on calcareous nannofossils and foraminifera, respectively. MNN and MMi biozones after Di Stefano et al. (2008), Fornaciari and Rio (1996), Iaccarino et al. (2011)

Paleomagnetism and rock magnetism

Prior to demagnetization, magnetic susceptibility per mass was measured using an Agico MFK1 Kappabridge (field strength 200 A/m; frequency 976 Hz). In addition, small amounts (~30 mg) of representative samples were used for thermomagnetic experiments on a modified horizontal translation type Curie balance (noise level $5 \times 10^{-9} \text{ Am}^2$). All experiments were performed in paleomagnetic laboratory Fort Hoofddijk at Utrecht University. Stepwise thermal demagnetization (TH) was applied on 143 specimens (CA 83; CG 21; CP 15; DV 24) in a magnetically shielded furnace with temperature increments of 20–40 °C. Further demagnetization of a sample was stopped when a clear random intensity increase was observed, indicating the built-up of a new magnetic mineral masking the original NRM. This usually occurred after 390 °C. To prevent this chemical reaction during thermal demagnetization, a dual TH-AF method was tested on a small selection of samples (20 samples). A sample was first stepwise thermally demagnetized and measured up to 350 °C and subsequently demagnetized by AF demagnetization.

Alternating field demagnetization (AF) was executed on 33 specimens using a laboratory-built automated measuring device applying sixteen 2–20 mT increments up to 100 mT by means of an AF-coil interfaced with the magnetometer (Mullender et al. 2016). After both magnetic cleaning methods (AF and TH), the natural remanent magnetization (NRM) was measured on a 2G Enterprises DC SQUID magnetometer (noise level $3 \times 10^{-12} \text{ Am}^2$). The per component demagnetization scheme of Dankers and Zijderveld (1981) was used to avoid measuring a large GRM during AF demagnetization. The paleomagnetic

results were analyzed in Zijderveld plots and quantified using the least square method on at least four consecutive demagnetization steps (Kirschvink 1980; Zijderveld 1967) via the open access online platform Paleomagnetism.org (Koymans et al. 2016). The paleomagnetic results are described in detail in Supplementary Material B.

Radioisotopic dating

The bulk tuff samples of Cepari (CPA), Campinița and the Brebu section were processed at the Mineral Separation facility of the Faculty of Earth and Life Sciences of VU University in Amsterdam. In several steps, the tuffs were crushed into blocks of ~1 cm³, disintegrated in a diluted calgon solution with a Robot Coupe blixer 4 v.v. and ultrasonic bath, wet sieved and dry sieved until a fraction of 90–500 µm was left. At this point, microscope analyses revealed that only sample CPA (with generally angular grains) contained high-quality biotite grains suitable for radio-isotopic dating. Biotite grains were separated using heavy liquids (diiodomethane; $\rho > 3.00 \text{ g/cm}^3$). During the final hand-picking step, the thickest, most angular hexagonal biotite minerals of the size fraction 500–400 µm and the ones without inclusions, as visible under an optical microscope, were selected for radioisotopic dating.

Mineral separates were packed in a 6 mm ID AI vial together with Fish Canyon Tuff sanidine (FCs) standards. Samples and standards were irradiated at the Oregon State University TRIGA reactor in the cadmium shielded CLICIT facility for 12 h.

⁴⁰Ar/³⁹Ar analyses were carried out at the geochronology laboratory of the VU University on an ARGUS VI+, which is a high sensitivity, low-resolution multi-collector noble gas mass spectrometer with an internal volume of 710 ml. The mass spectrometer is equipped with four Faraday cups at the H2, H1, AX and L1 positions and two compact discrete dynodes (CDDs) at positions L2 and L3. The system is equipped with a 10^{12} Ohm amplifier on H2 and 10^{13} Ohm amplifiers on H1, AX and L1 cups. The resolution of the system is ~200 and, therefore, does not resolve hydrocarbon or chlorine interferences. The ARGUS VI+ has an NP10 getter and ion gauge on the source of the mass spectrometer. The NP10 getter is run cold and the ion gauge is turned off during analyses, because of its pumping capacity for argon. Samples were run on H2–L2 collectors. Sixteen single grain fusions were performed on sample CPA. Data reduction is done in ArArCalc (Koppers 2002). Ages are calculated with decay constants after Min et al. (2000) and $28.201 \pm 0.022 \text{ Ma}$ for FCs (Kuiper et al. 2008). The atmospheric air value of 298.56 from Lee et al. (2006) is used. All relevant procedures and analytical data for age calculations can be found online (Supplementary Data C).

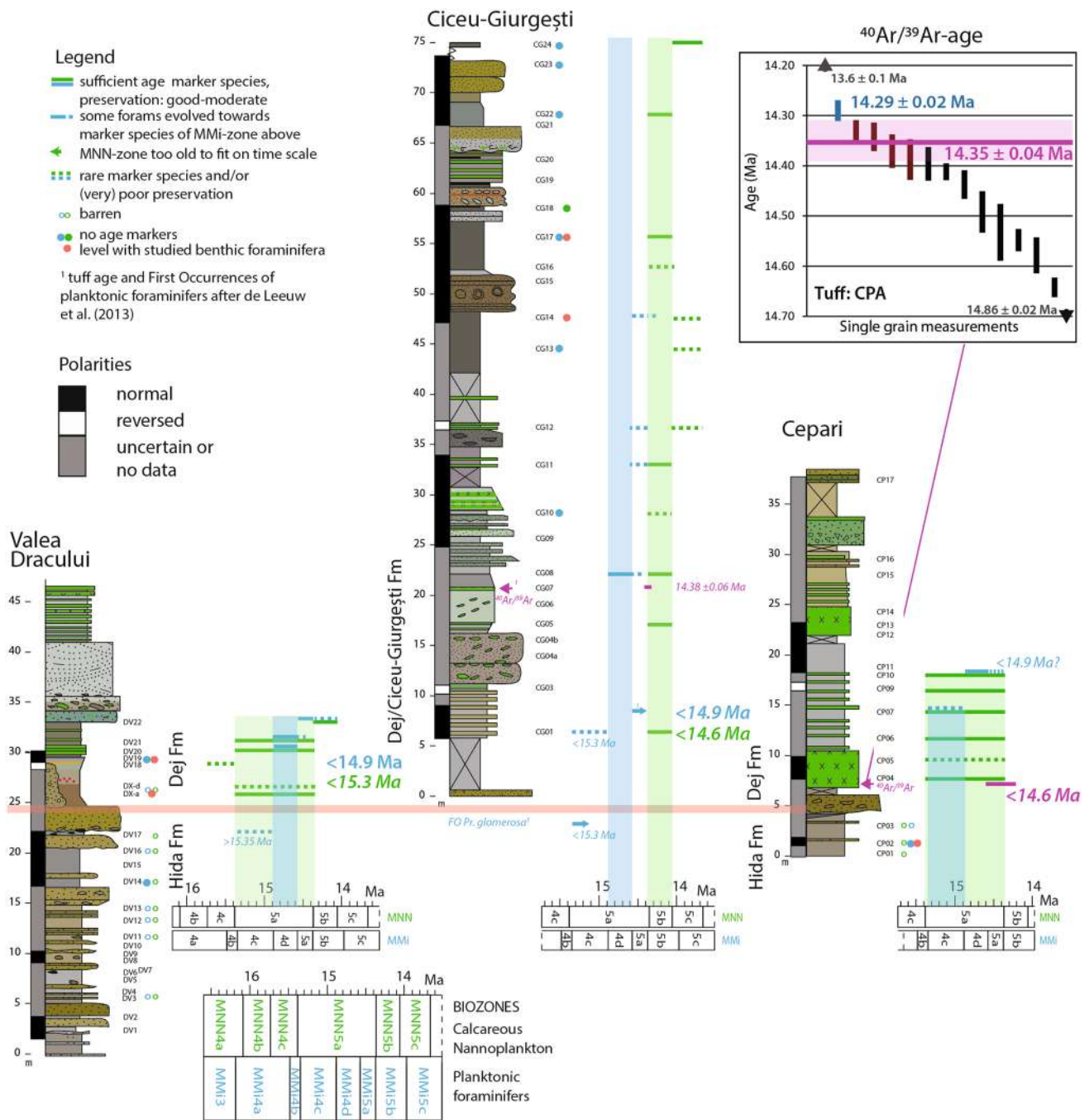


Fig. 4 Synthesis of calcareous nannofossil (green), planktonic foraminifer (blue), magnetostratigraphic results and $^{40}\text{Ar}/^{39}\text{Ar}$ dated tuff levels (purple) of three sections in the N Transylvanian Basin (DV, CG, CP). MNN and MMi biozones after Di Stefano et al. (2008), Fornaciari and Rio (1996), Iaccarino et al. (2011). The $^{40}\text{Ar}/^{39}\text{Ar}$ ages in CG and DV are after de Leeuw et al. (2013). The shift to open marine conditions is marked by horizontal red bars. Vertical semi-transparent green and blue boxes mark the age ranges for the flooding boundaries based on nannofossils and foraminifera, respectively.

Ages in green and blue reflect the maximum ages for the onset of the open marine transgression based on calcareous nannofossils and foraminifera, respectively. See the legend and text for more details. $^{40}\text{Ar}/^{39}\text{Ar}$ results for all measured biotite grains of sample CPA (6 m, section CP). The youngest reliable age is marked in blue. The red zone shows the mean age based on the four grains marked in red. The different approaches for age interpretation are described in “Integrated (chrono)stratigraphy and environments” of the main text

Results

Paleomagnetic results

In all studied sections, AF demagnetization resulted in stable vectors at least in the range 20–32 mT, with exceptions up to 60 or 80 mT. Thermomagnetic experiments (Curie Balance) together with thermal demagnetization results show that two groups of samples exist in the sections (Fig. 3a–d).

The majority of the NRMs from all samples belong to the first group and are stable within the temperature range 180 °C to max. 390 °C. Above this temperature, pyrite converts into magnetite and the original signal is masked by a random magnetic vector visible during thermal demagnetization and in thermomagnetic experiments, complicating interpretation of the paleomagnetic signal (Fig. 3a). The combined TH-AF method could resolve the magnetic vector only in some limited cases (Supplementary Data B). The presence of pyrite (FeS₂) throughout the majority of the sections suggests that sub-oxic to anoxic bottom conditions occurred frequently in the N Transylvanian Basin and SE Carpathian foredeep during the studied period (Berner 1970, 1984; Rickard and Luther 2007). In contrast, the interval with mottled gray, green and red beds in the lowermost part of CA (0–40 m) lacks pyrite and comprises the second group of samples. Most samples carry two or three higher temperature components with the same polarity (Fig. 3b). Their magnetic components are generally stable in the temperature ranges: 180–300 °C or 180–450 °C, > 300–550 °C, and > 550–660 °C, respectively, reflecting the magnetic carriers magnetite (max. 550 °C) and hematite (up to ~ 660 °C) (Fig. 3c). In addition, the presence of an iron sulphide is suggested by irreversible behavior in the range 250–420 °C in the thermomagnetic runs. Better oxygenated conditions and/or a higher terrestrial influx is indicated by the frequent occurrence of magnetite in the interval 100–135 m in the Campinița section.

All sections are dominated by normal polarity (Figs. 3e, 4). The long pre-flooding interval in Campinița contains several short reversed horizons in the basal part (18–60 m). DV has one reversed sample just above the flooding boundary, and CP contains one ~ 10 m above the transgressive surface. However, caution is needed for the use of these data for magnetostratigraphy, because the scatter in paleomagnetic directions in all sections is high and indications for partial remagnetization exists. The mean inclination of the Campinița section is abnormally low (~ 3°), which might reflect the partial overprint of reversed directions and/or a normal field overprinting an older reversed field. Sample CA58.1 has a reversed vector between 240 and 420 °C when plotted in geographic coordinates, while

in tectonic coordinates a realistic magnetic direction cannot be recognized (Fig. 3d). This implies a reversed overprint. For CG and DV, the mean directions in geographic coordinates are closer to the present-day field direction than those in tectonic coordinates, suggesting that part of the NRM might be overprinted by a recent magnetic field (Supplementary Data B).

Radioisotopic results

All ⁴⁰Ar/³⁹Ar measurements with a high percentage of radiogenic ⁴⁰Ar* (> 85%) and a low value for ³⁷Ar_{Ca} (< 2.5 fA; and, therefore, Ca-poor minerals) are considered reliable. This is valid for thirteen out of sixteen single biotite grain measurements of tuff level CPA in the Cepari section (Fig. 4). The ages of these grains range between 14.29 ± 0.02 and 14.86 ± 0.02 Ma. Due to the volcanoclastic nature of the tuff, reworked, detrital grains might be present in the sample. These components are older than the eruption age, so the oldest grains likely do not represent the eruption age. Following this reasoning, we could state that the youngest reliable grain with an age of 14.29 ± 0.02 Ma is the best measure of the eruption age. Alternatively, we could assume that some ⁴⁰Ar loss occurred and that the eruption age is approximated best by a statistically sound mean value (data points are included until MSWD > *T* test statistic using a 2σ confidence limit). This approach is similar to that of U/Pb studies where the youngest grain(s) might represent Pb loss and the oldest grain(s) xenocrystic components (e.g., Sahy et al. 2015). In this case level, CPA has an age of 14.35 ± 0.04 Ma. If we simply consider the full age range of all grains, the tuff is younger than ~ 14.6 Ma.

⁴⁰Ar/³⁹Ar-dating on biotite is regularly reported to be less precise than dating on K-feldspar, because the presence of some extraneous ⁴⁰Ar in biotite grains will result in overestimated age values (e.g., Hora et al. 2010). Another phenomenon which can lead to an overestimated age of platy minerals such as biotite is loss of ³⁹Ar during recoil (Onstott et al. 1995; Turner and Cadogan 1974). However, for our study this effect is expected to be very small since the net loss of ³⁹Ar is assumed to be trivially small for relatively large grains (e.g., Hall 2014). Our measured age range in the Cepari tuff sample is overlapping with the ⁴⁰Ar/³⁹Ar-ages for the base of the Dej tuff in different sections in the NW Transylvanian Basin as presented by de Leeuw et al. (2013), being 14.37 ± 0.06 and 14.36 ± 0.06 Ma and based on K-feldspars. This suggest that extraneous ⁴⁰Ar and recoil only affected the age determinations of our biotite sample to a small extent. An overview of the analytical data and procedures is found in Supplementary Material C.

Integrated (chrono)stratigraphy and environments

Flooding boundary characteristics

The flooding boundary in the SE Carpathian Foredeep is in both sections (CA, BR) marked by a thin (<0.5 m) anoxic (potentially sapropelic) interval characterized by black clays with jarosite and secondary iron oxide crusts that probably are the result of the pyrite oxidation (Fig. 3e). It is followed by silt/fine sands with water escape structures and marls with tuffs. Benthic foraminifera below the anoxic interval indicates some oxygen restriction, whereas the ones just above suggest a shift to a more oxic, shallow marine environment with deviating salinity (Sant et al. 2019).

In addition, tiny (<125 µm) *Streptochilus/Bolivina* foraminifera were found in a few samples in the 20 m-interval below the main flooding boundary in Campinița (CA12, CA11, CA10 and CA9B, ~118–134 m). Similar foraminifera are known from the pre-flooding strata of the N Transylvanian Basin (Beldean et al. 2010, 2013), but it is the first time that these tiny foraminifera are documented from a similar stratigraphic position in the Romanian Carpathian Foredeep. *Streptochilus/Bolivina* foraminifera have a tychoepelagic lifestyle, meaning that they can be both benthic and planktonic (Darling et al. 2009; Smart and Thomas 2007). This attribute makes them very adaptable to new circumstances and suitable to invade a new environment, such as one with highly fluctuating nutrient conditions (Beldean et al. 2010). They are also commonly associated with oxygen minimum zones (Bernhard and Sen Gupta 1999; Nigam et al. 2007; Smart and Thomas 2007).

Apart from the appearance of a rich Langhian foraminiferal assemblage, the flooding boundaries in the Transylvanian sections do not have much in common. At Valea Dracului, the flooding base comprises brown, green-gray and purple-gray clays followed by marl with tuffs. A distinct level (1 cm) with rounded grains with iron oxides occurs within the clays. Rare and recrystallized foraminifera suggest shallow-marine waters. In the Cepari section, the flooding boundary is characterized by gray clays cut by a poorly sorted sand with sand intraclasts. Abundant large Miocene benthic foraminifera and ostracods indicative for a shallow shelf with a high organic load are present below the boundary.

Timing of the flooding in the SE Carpathian Foredeep

Only samples with moderate-good biostratigraphic preservation are taken into account for the chronostratigraphy (Sant et al. 2019). The pre-flooding (Doftana Fm) sediments have a generally poorly preserved fossil content with few age markers. Nevertheless, the biostratigraphic data suggest that the uppermost part of the Doftana Fm is Langhian in

age, which is in agreement with (Orbocea et al. 1986b), and that marine incursion occurred before 14.6 Ma (Fig. 3e). The interval between 98 and 102 m contains marker fossils attributed to the Langhian Biozone MNN5a and most likely MMi4c (*P. glomerosa glomerosa*), indicating a period with a marine incursion (Sant et al. 2019). Following the MNN4a attribution of sample CA36, the base of the Langhian stage is positioned above the gypsum interval around level 80 m, while an even lower position ~60 m cannot be excluded because of the uncertain MMi4c attribution of sample CA44. The dominantly normal polarity for the Langhian part of the section (60/80–153 m) indicates that the majority of the paleomagnetic signal is overprinted because it does not correspond to the Global Polarity Time Scale (Hilgen et al. 2012). The normal polarity with some reversed intervals (interval 0–58 m) might correspond to the late Burdigalian Chron C5Cn.

The biostratigraphic age data of the transgressive sequences in Campinița and Brebu in the SE Carpathian Foredeep are in good agreement with each other (Fig. 3e). In both sections, the nannofossil and foraminifera data at the flooding boundary overlap, and show a distinct shift in age (and preservation) compared to the samples below the boundary. The flooding correlates to the MNN5a (15.3–14.4 Ma) and MMi5a (14.6–14.4 Ma) biozones, indicating that the open marine conditions in this isolated part of the Carpathian Foredeep region installed most probably after 14.6 Ma.

Timing of the flooding in the N Transylvanian Basin

All analyzed samples from the pre-flooding strata (Hida Fm) were barren in planktonic foraminifera and calcareous nannofossils. In Valea Dracului, the MNN5a zone overlaps with the MMi4d zone, narrowing the onset of open marine conditions down to 14.9–14.6 Ma (Fig. 4). One good quality reversed polarity sample just below the Dej tuff and appearance of *O. suturalis* could potentially be correlated to Chron C5ADr (~14.78–14.61 Ma) (Hilgen et al. 2012).

The biostratigraphic data at the base of the Ciceu-Giurguş section are scarce and the interpreted biozones are conflicting, with planktonic foraminifera indicating older ages (14.9–14.4 Ma) than the nannofossils (<14.4 Ma). The older age determination based on planktonic foraminifera could be due to the uncertain classification of the *Praeorbulina/Orbulina* group because of the poor preservation of the fossils (Sant et al. 2019). De Leeuw et al. (2013) described the finding of *P. glomerosa* (sensu Blow) at ~5 m below the base of our section, and the FO of *O. suturalis* at ~8 m stratigraphic height. This suggests (using the Mediterranean biochronology) an age range of 15.3–14.6 Ma for the basal part (<8 m). Their $^{40}\text{Ar}/^{39}\text{Ar}$ age around 20 m is in agreement with the nannofossil data from this study

(MNN5b) and their study (de Leeuw et al. 2013). Like in the Valea Dracului section, a reversed sample is present just below the appearance of Dej tuff elements, which could match with Chron C5ADr.

In Cepari, the age range of the CPA tuff (14.6–14.3 Ma) is overlapping with the calcareous nannofossil MNN5a Sub-zone, while the scarce foraminifera samples might suggest a slightly older age (Fig. 4). Note that the poor preservation of *Praeorbulina glomerosa* and perhaps *Orbulina suturalis* could hamper the recognition of *Orbulina suturalis* in sample C7B. No age markers were encountered in the basal part below the transgression.

Summarizing, the widespread open marine Badenian Sea in the N Transylvanian Basin most probably installed in the period 14.9–14.4 Ma (MMi4d), relying mostly on data from

the DV and CP sections. Scarce or poorly preserved planktonic foraminifera in the clastic base of the transgressive sequences in the CG and DV sections hint at the presence of the MMi4c Biozone (15.3–14.9 Ma), showing that episodic influxes of water with Mediterranean foraminifera in the study areas occurred the earliest by 15.3 Ma, but that a deeper and fully marine environment throughout the basin was most probably installed later. Our proposed age range for the transgression fits much better with the most recent radio-isotopic ages between 14.6 and 14.2 Ma for the overlying Dej tuffs than the ages provided by previous studies based on open ocean (Atlantic) biochronologies (~16.4 Ma) (Beldean et al. 2013; de Leeuw et al. 2013; Krézsek and Filipescu 2005).

No consensus exists about the salinity in the N Transylvanian Basin. Although older studies inferred a fluvio-deltaic setting for the majority of the Hida Fm (Popescu 1975), more recent biostratigraphic and sedimentological studies propose almost normal levels of salinity for most

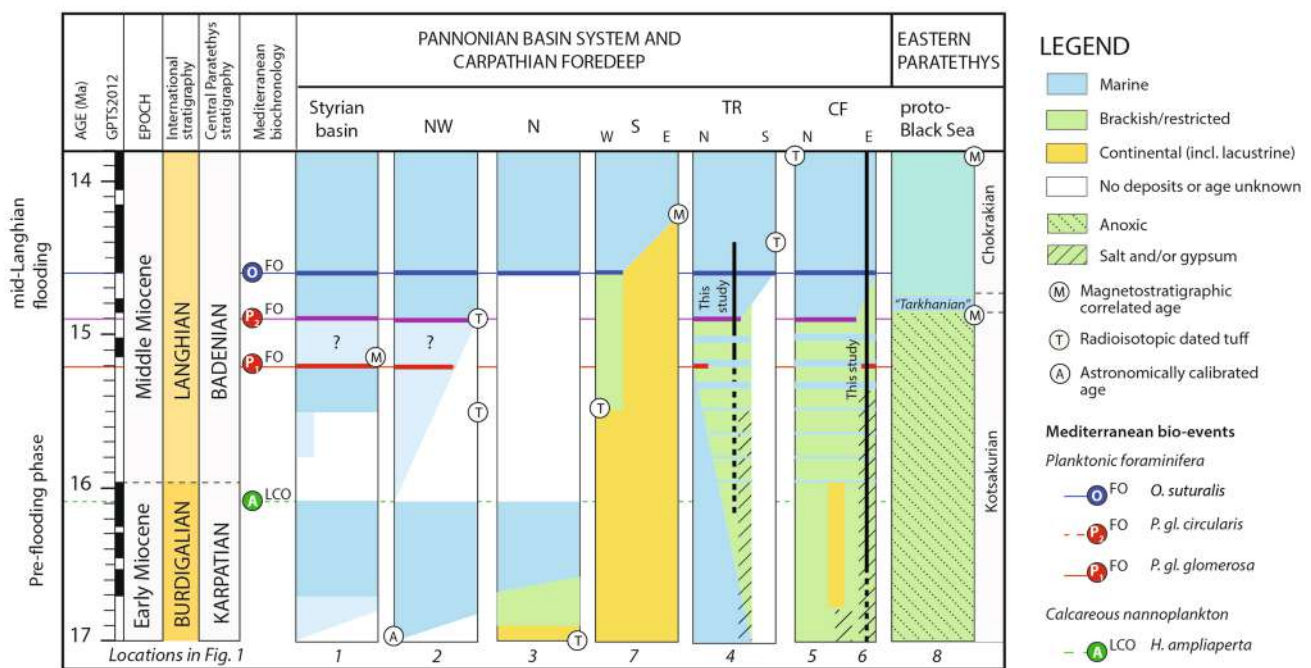


Fig. 5 Compilation of paleoenvironmental and age data for key areas in Central Europe. Biochronology follows the Mediterranean. Time-scale (GPTS) after Hilgen et al. (2012). Note that the FO of *Praeorbulina* in the Central Paratethys refers to the FO of *P. gl. glomerosa*, occurring always together with *P. gl. curva* (Rögl et al. 2002b). The Slovenian strait to the Mediterranean was most likely open since ~17 Ma because the onset of the late Burdigalian sea is astronomically correlated to be ~17 Ma in the NW Pannonian region (Korneuburg Basin; Zuschin et al. 2014). The main literature used for the compilation of the basins: Styrian Basin (Kollmann 1965; Auer 1996; Rögl et al. 2002b; Hohenegger et al. 2009; Spezzaferri et al. 2009); NW—Austrian Molasse (Rögl et al. 2003; Ćorić and Rögl 2004; Ćorić et al. 2004) and Vienna basin—(Andreyeva-Grigorovich et al.

2001; Rögl et al. 2002a; Kováč et al. 2004; Harzhauser et al. 2018); N—Nograd/South Slovakian—(Vass 2002b; Pálffy et al. 2007; Nagymarosy and Hámor 2012; Selmečzi et al. 2012; Püspöki et al. 2017); S—North Croatian basins (Baldi 2006; Ćorić et al. 2009; Mandić et al. 2012; Kovačić et al. 2016; Pavelić and Kovačić 2018), northern Bosnia (Mandić et al., 2019b; Pezelj et al., 2013) and Dinaride Lake System (de Leeuw et al. 2012); TR—Transylvanian Basin (see caption Fig. 2); CF—Polish-Ukrainian (Andreyeva-Grigorovič et al. 1997; Oszczykpo and Oszczykpo-Clowes 2011; Oszczykpo et al. 2014, 2016) and Romanian Carpathian Foredeep (see caption Fig. 2); proto-Black Sea (Gontsharova 1991; Palcu et al. 2017, 2019). See Sant et al. (2017) and Sant (2018) for more background information

of the pre-flooding succession in the NW Transylvanian basin (Beldean et al. 2010, 2012, 2013; Székely and Filipescu 2016). In the period right before the Badenian flooding, the sea became shallow and restricted because of the prograding delta front, which is especially noted in the NE Transylvanian Basin (Beldean et al. 2013).

The mid-Langhian marine flooding in the Paratethys

Temporal and spatial extent

The timing of the installment of the Badenian Sea in the study areas in the N Transylvanian Basin (< 14.9–14.4 Ma) fits very well with the timing in the neighboring Polish-Ukrainian Foredeep (4–5 in Fig. 5). Like in the Transylvanian Basin, several marine ingressions marked by scarce Mediterranean fossils are known from the late Burdigalian onwards, but most biostratigraphic data with *P. glomerosa circularis* and/or *O. suturalis* point to a definite transgression of the sea after 14.9 Ma (Andrejeva-Grigorovič et al. 1997; Andrejeva-Grigorovich et al. 2008; Oszczypko et al. 2006, 2016; Oszczypko and Oszczypko-Clowes 2011).

The environmental evolution of the restricted Carpathian Bend area of the East Carpathian foredeep (i.e. Campinița section) followed a similar pattern with several marine incursions before the mid-Langhian, and the installment of the permanently open marine Badenian sea after 14.6 Ma (*O. suturalis*) (Fig. 5). The same is valid for most marginal/shallow marine areas, including the NE Vienna basin (2 in Fig. 5) (Harzhauser et al. 2018), SE Transylvanian Basin (Ilie 1951) and South Carpathian Foredeep (Iorgulescu 1951). The basal flooding strata in the North Pannonian Basin (Hámmor 1985; Kováč et al. 2007; Selmeczi et al. 2012; Vass 2002a) and North Croatian basins (Ćorić et al. 2009) (3 and 7 in Fig. 5) are uniformly associated with *O. suturalis* (< 14.6 Ma). The transgression reached the S Pannonian Basin relatively late, namely between 14.6 and 14.2 Ma (Ćorić et al. 2009; Mandić et al. 2019a, b; Pezelj et al. 2013).

The timing and mode of the onset of the marine flooding in the western areas (1–2 in Fig. 5) is not completely resolved. In the Styrian Basin (1 in Fig. 5) and Upper Austrian Molasse basin, a basal flooding with scarce *P. gl. glomerosa*, *P. glomerosa curva* and/or *H. waltrans* was reported, implying a maximum age of 15.5 Ma (Ćorić and Rögl 2004; Hohenegger et al. 2009; Rögl et al. 2002b; Spezzaferri et al. 2009). However, it is uncertain whether these basal sediments reflect the onset of the long-lasting early Badenian Sea or only reflect a short marine phase.

Summarizing, the widespread “mid-Langhian” flooding in the Central Paratethys occurred in most regions after

~ 15 Ma. This coincides with the “Tarkhanian” marine phase in the Eastern Paratethys (Euxinian Basin) that was magnetostratigraphically dated at 14.85 Ma (Palcu et al. 2019). This phase was caused by marine influx from the East Carpathian Foredeep (Central Paratethys) to the proto-Black Sea basin (Eastern Paratethys) via the Carasu strait (Chiriac 1969) (Figs. 5, 6).

Outflow of nutrient-rich brackish water from Eastern Paratethys (similar to present day Black Sea outflow and temporary nannoplankton blooms in the Marmara Sea) may be related to the temporarily blooms of the opportunistic *Streptochilus* foraminifera just below the flooding boundary in the studied areas of the East Carpathian Foredeep and Transylvanian Basin (Beldean et al. 2010, 2013). The temporary low oxic setting and anoxic levels in the Campinița and Brebu sections might be related to the low-density brackish outflow plume of Black Sea water (Liu et al. 2017; Palcu et al. 2019). The suggestion by Beldean et al. (2010) that the *Streptochilus* blooms might have a similar age range as the blooms in the Atlantic and Indian Oceans (~ 19–17 Ma) is erroneous. The tiny *Streptochilus* in the SE Carpathian Foredeep are assigned to the Langhian stage (< 16 Ma), and in the Transylvanian Basin (DV section) they occur up to ~ 5 m below the first level with Langhian fossils (Beldean et al. 2013) (Fig. 4). The *Streptochilus* blooms could serve as ecological markers for the initial phase (or phases) of the marine transgression in the eastern part of the Central Paratethys.

Our new ages can be used to improve paleogeographic reconstructions (Fig. 6a). Around ~ 14.2 Ma the Badenian Sea had spread out to the majority of the basins in Central Europe. It significantly expanded north-eastward in the Polish-Ukrainian Foredeep (at least 70 km), and branched out towards the Republic of Moldova (Goncharova and Il'ina 1997; Oszczypko and Oszczypko-Clowes 2011). The S Transylvanian Basin was most-probably well-connected with the S and SE Carpathian Foredeep (Iorgulescu 1951), and also to the southern Pannonian Basin via a corridor south of the Apuseni Mountains (Pop 1951; Roșca 1951). The connection between the NW–SE oriented Focsani depression in the East Carpathian Foredeep and the SW Transylvanian Basin most likely persisted in the Langhian (Mirăuța 1967; Tărăpoancă 2003). See Supplementary Material D for background information on the facies map in Fig. 6a.

Drivers for the mid-Langhian flooding

The mid-Langhian flooding took place in a dynamic period with large scale tectonic events and global sea-level fluctuations and is probably the result of a complex interplay between those processes (Balázs et al. 2016; John et al. 2011; Kováč et al. 2017b, 2018; Zachos et al. 2001). The geodynamic processes (i.e. active rifting and uplift) controlled to

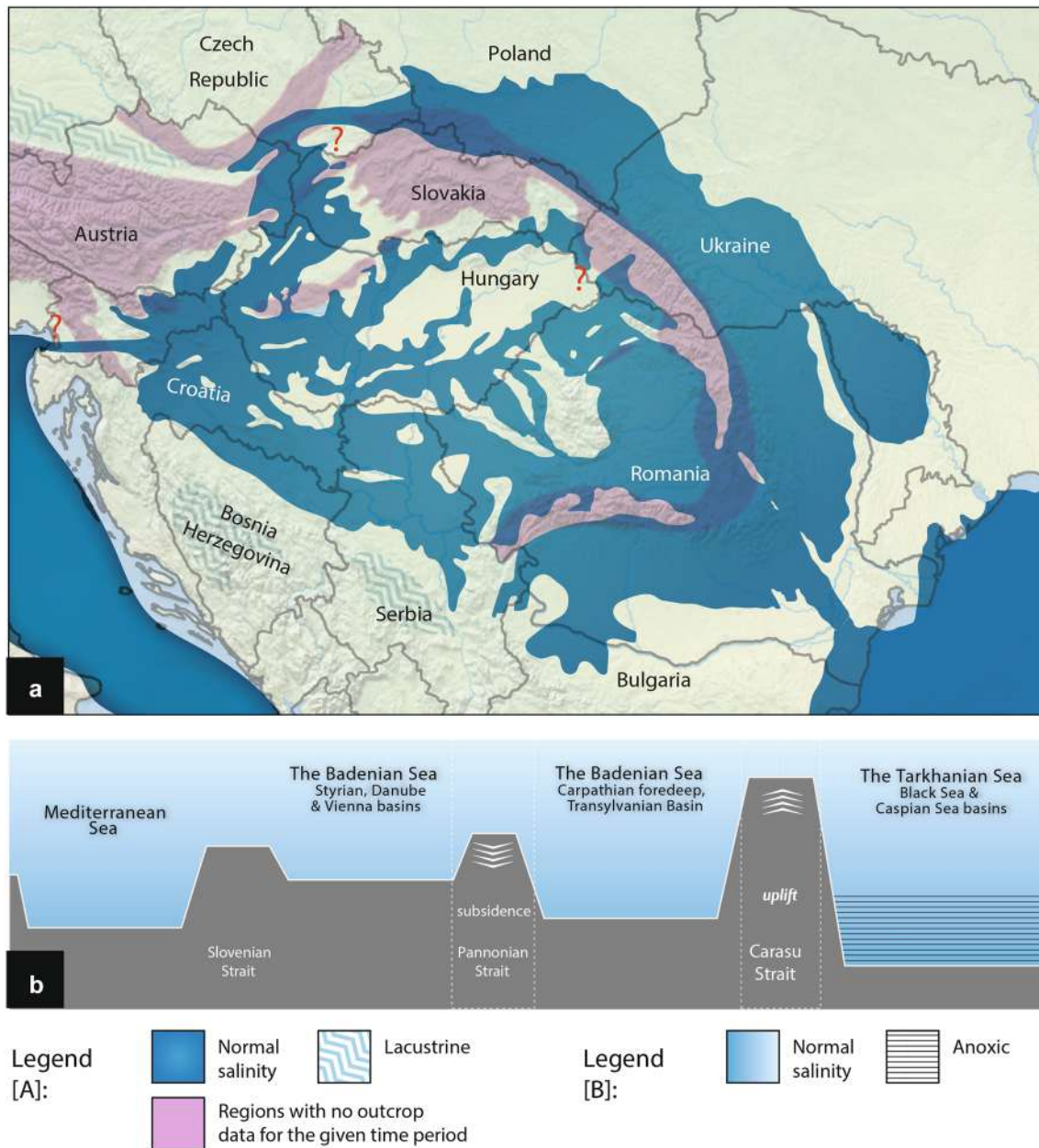


Fig. 6 **a** Map showing the (minimal) outline of the open marine Badenian Sea (~14 Ma) based on present-day outcrops supplemented by stratigraphic data from boreholes and seismic profiles, which were mostly used in areas with a large post-Middle Miocene cover and/or Miocene sediments in folded nappes (e.g., Carpathian fold and thrust belt). Some paleoconnections are inferred based on structural data

about the timing of uplifted terranes. Question marks denote uncertain areas, usually as a result of a lack of age data. Note: these are no palinspastic reconstructions; see Kováč et al. (2017a, b) for palinspastic maps. Background information with references for each area are depicted in Supplementary Material D. **b** Simplified gateway configuration for the Badenian Sea

a large extent the accommodation space, sedimentation rates and opening or closure of connections between the different basins (Báldi et al. 2002; Horvath et al. 2006; Matenco and Radivojević 2012; Palcu et al. 2017). As a result, the expression of the flooding is slightly different in each area. For example, the Ugljevik region in the SE Pannonian basin was flooded after 14.2 Ma (Mandic et al. 2019b; Pezelj

et al. 2013), which suggests that in this marginal area the space and/or connection to the sea was mainly controlled by tectonics.

A combined record of eustatics in the west Atlantic and Pacific oceans suggests a sea-level rise of ~30 m around 14.7 Ma (right after the sea-level drop associated with Mi3a isotope event), followed by a slow rise until ~14.2 Ma (John

et al. 2011; Kominz et al. 2008). This eustatic phase overlaps largely with the timing of the expansion of the normal saline Badenian Sea which we here constrained to the age range 14.9–14.4 Ma. Thus, it is possible that the mid-Langhian flooding was triggered by this global sea-level rise. The rise is in agreement with the fact that the fully marine water transgressed far over the platforms next to the foredeeps, rapidly transformed lakes into marine basins (Kovačić et al. 2016; Krézsek and Bally 2006; Pavelić and Kovačić 2018) and spilled over into the Eastern Paratethys via the Carasu strait (Goncharova and Il'ina 1997; Holcová et al. 2018; Palcu et al. 2019) (Fig. 6b).

The eustatic sea-level rise might have overcome a barrier in the region of the Trans-Hungarian Shear Zone that separated the restricted environments in the east from the (semi-restricted) marine environments in the west (Fig. 6b), allowing fully anti-estuarine circulation and oxidation of all marine areas in the Central Paratethys (Kováč et al. 2017a). Tectonic subsidence or rotations might have assisted in removing the barrier, but since the shift from pre-flooding to fully marine Badenian conditions is very sharp in most records it seemed to occur relatively fast and it is most likely that the sea level rise was the main trigger for the onset. The early Badenian sea was sustained in most places at least up to 13.8 Ma suggesting that by that time the connection to the Mediterranean had widened or deepened to retain the marine environments even after the global sea level dropped again.

The pre-flooding phase

With the “pre-flooding phase” we refer to the period of variable lacustrine, brackish and marine conditions in Central Europe between the opening of the Slovenian strait and the mid-Langhian flooding (~ 17–15 Ma) (Fig. 5). Kováč et al. (2018) defined most of this period as a relative sea-level low-stand (~ 16.5–15.5 Ma). The paleoenvironmental evolution of this phase is poorly understood compared to that of the widespread Badenian Sea but it deserves some attention.

It is important to stress that in many Paratethys basins all pre-flooding sediments are commonly attributed to the upper Burdigalian, based on the assumption that the base of the marine Badenian succession has the same age as the Langhian Stage of the Mediterranean (~ 16 Ma). However, the much younger age for the marine flooding in many locations implies that a large part of the pre-flooding strata could actually be early Langhian in age. This is in agreement with the data from the Campinița section in the Romanian Carpathian Foredeep (this study) and stratigraphic data from other studies (Andreyeva-Grigorovich et al. 2001; Kovačić et al. 2016).

A gradual or abrupt ‘brackification’ towards the east during the pre-flooding phase is indicated by a general decrease

in the amount of marine nannofossils and microfossils, with the richest assemblages nearby the Slovenian strait (Bartol 2009; Hohenegger et al. 2009) and poorest in the East Carpathian Foredeep (Fig. 7; Suppl. D). A snapshot during the pre-flooding phase (at ~ 16.5 Ma) is depicted in Fig. 7 to exemplify a scenario with a decrease in salinity from the Slovenian strait towards the east. In this period, the north Pannonian basin was marine (Pálffy et al. 2007), while the majority of the East Carpathian Foredeep was most probably semi-restricted marine (e.g., Mirauta 1967; Rusu et al. 1996; Oszczykko et al. 2016). Connections between the Carpathian Foredeep, N Transylvanian Basin and NE Pannonian Basin most probably still existed over the rising East Carpathians (Mirăuța 1967). The North Carpathian Foredeep was likely still isolated from the Vienna Basin and NW Carpathian Foredeep (SW Czech Republic) by the uplift of the Cieszyn-Slavkov Paleo Ridge (SE Czech Republic–SW Poland) (Oszczykko 1997; Oszczykko and Oszczykko-Clowes 2011). In the presented example, one large jump from marine to hyposaline conditions occurs at the location of one dominant gateway in the Pannonian Basin, potentially at the Mid-Hungarian Shear Zone (Fig. 7b). The boundary between the areas with normal and low salinities would require the presence of a paleo-strait limiting water exchange between both regions.

In some areas in the SE Carpathian Foredeep, calcareous material is completely absent (Andreyeva-Grigorovič et al. 1997). If marine calcareous fossils occur, they are confined to narrow horizons representing short marine influxes, such as in the middle part of the Campinița section and in other parts of the East Carpathian Foredeep (e.g., Mirauta 1967; Andreyeva-Grigorovich et al. 2008), where they are known to occur during late Burdigalian–early Langhian times. It suggests that the environments were hostile to life or poorly suited for the preservation of shells. It also implies that a fragile connection existed between the East Carpathian Foredeep and the rest of the Central Paratethys Sea (Pauca 1976).

No direct sedimentary or biostratigraphic proof exists for a sea strait between the Central Paratethys and the Eastern Paratethys (proto-Black Sea Basin) during the pre-flooding phase (Chiriac 1969; Goncharova and Il'ina 1997; Gontsharova 1991). However, new modeling data suggest that the Eastern Paratethys had a positive water budget during the pre-flooding period of the Central Paratethys (Simon et al. 2018). Moreover, tectonic data indicate that the NW–SE striking normal fault system cross-cutting the Romanian East Carpathian Foredeep and Carasu strait area could have already existed since the late Burdigalian (Tărăpoancă 2003). Therefore, it is possible that some spill-over from the brackish and anoxic proto-Black Sea (Mai-kop Series) to the neighboring East Carpathian Foredeep occurred (Palcu et al. 2019). At that time, the Black Sea was notorious for its severe anoxic paleoenvironments with

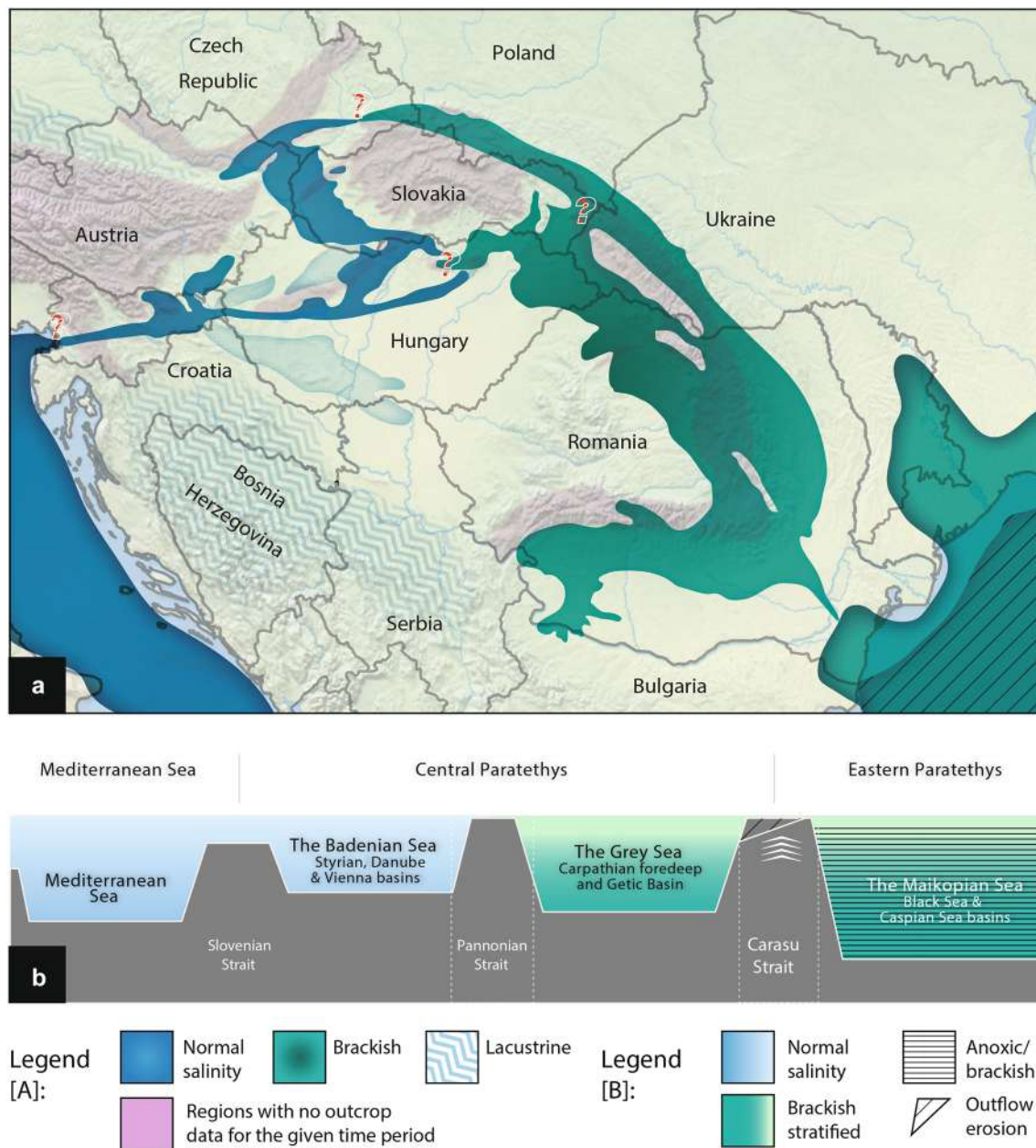


Fig. 7 **a** Map showing a paleogeographic scenario with the outline of the marine, brackish and freshwater basins for one snapshot during the pre-flooding phase (~16.5 Ma). The semi-transparent marine bays in present-day Hungary are uncertain. See caption of Fig. 6 and

Supplementary Material D for background information. **b** Simplified gateway configuration for this scenario of the pre-flooding phase. All background information on the map is found in Supplementary Material D

salinity stratification and lack of vertical circulation (Hudson et al. 2008; Katz et al. 2000; Sachsenhofer et al. 2017), so a spillover would be one way to explain the hostile environments in the foredeep. In this case, a freshwater plume would have existed in the SE Carpathian Foredeep that faded out towards the areas more distant of the Carasu strait. Another way to explain this would be a very large input of freshwater from the mountains and surrounding plains creating an anti-estuarine circulation with bottom anoxia (Kováč et al. 2018).

Conclusions

Our integrated magneto-biostratigraphic and $^{40}\text{Ar}/^{39}\text{Ar}$ data from the Transylvanian Basin and East Carpathian Foredeep provide new insights into the age and paleoenvironments of the pre-flooding and mid-Langhian marine flooding phases in Central Europe.

By applying the most recent Mediterranean biochronology, we show that marine incursions into the dominantly

brackish basins of Central Europe (traced by horizons with preserved marine micro and/or nannofossils) occurred at least since the latest Burdigalian. Such short marine influxes have been reported throughout the Ukrainian–Romanian Foredeep areas in the late Burdigalian–early Langhian period (pre-flooding period). A progressive eastward restriction in the marine incursions in the Central Paratethys realm during the pre-flooding phase (~17–15 Ma) seems to go hand-in-hand with a decline in the presence of marine Mediterranean micro- and nannofossils. The apparently lower salinity in the eastern part of the Central Paratethys might have been enhanced by the presence of one or several sills blocking the inflow of marine water into these areas. We speculate that the striking scarcity of calcareous material of all sorts in the East Carpathian Foredeep might have been enhanced by pulses of nutrient-rich brackish water from the neighboring Eastern Paratethys via the Carasu strait region.

The widespread open marine settings of the Badenian Sea in the NW Transylvanian Basin were installed between 14.9 Ma (MMi4d biozone) and 14.4 Ma. In the marginal study area in the SE Carpathian Foredeep, fully marine conditions set in slightly later than in the NW Transylvanian Basin (14.6–14.4 Ma versus 14.9 Ma).

In the majority of the formerly brackish and freshwater basins in the north, central and southern Central Paratethys, open marine conditions were also introduced between ~14.9 and 14.4 Ma. The significant expansion of the Central Paratethys sea suggests that, besides the geodynamically controlled subsidence in the Pannonian Basin, global sea-level rise might have enhanced the change in paleogeography and paleoenvironments. This is in good agreement with the incursion of Central Paratethys open marine water into the restricted Eastern Paratethys at 14.85 Ma (regional Tarkhanian Stage).

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