

# The end of the Great Khersonian Drying of Eurasia: Magnetostratigraphic dating of the Maeotian transgression in the Eastern Paratethys

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## Abstract

Central Eurasia underwent significant palaeoclimatic and palaeogeographic transformations during the middle to late Miocene. The open marine ecosystems of the Langhian and Serravallian seas progressively collapsed and were replaced in the Tortonian by large endorheic lakes. These lakes experienced major fluctuations in water level, directly reflecting the palaeoclimatic conditions of the region. An extreme lowstand of the Eastern Paratethys lake (−300 m) during the regional Khersonian stage reveals a period of intensely dry conditions in Central Eurasia causing a fragmentation of the Paratethys region. This period of “Great Drying” ended by a climate change towards more humid conditions at the base of the Maeotian stage, resulting in a large transgressive event that reconnected most of the Paratethyan basins. The absence of a robust time frame for the Khersonian–Maeotian interval hampers a direct correlation with the global records and complicates a thorough understanding of the underlying mechanisms. Here we present a new chronostratigraphic framework for the Khersonian and Maeotian deposits of the Dacian Basin of Romania, based on integrated magneto-biostratigraphic studies on long and continuous sedimentary successions. We show the dry climate conditions in the Khersonian start at 8.6–8.4 Ma. The Khersonian/Maeotian transition is dated at 7.65–7.5 Ma, several million years younger than previous estimates. The Maeotian transgression occurs later (7.5–7.4 Ma) in more marginal and shallower basins, in agreement with the time transgressive character of the flooding. In addition, we date a sudden water level drop of the Eastern Paratethys lake, the Intra-Maeotian Event (IME), at 6.9 Ma, and hypothesize that this corresponds to a reconnection phase with the Aegean basin of the Mediterranean. Finally, we discuss the potential mechanisms explaining the particularities of the Maeotian transgression and conclude that the low salinity and the seemingly “marine influxes” most likely correspond to episodes of intrabasinal mixing in a gradual and pulsating transgressive setting.

## KEYWORDS

climate change, Khersonian/Maeotian, late Miocene, magnetostratigraphy, Messinian, palaeogeography, Paratethys, Stratigraphy

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## 1 | INTRODUCTION

A major part of the Earth's continental surface is represented by endorheic (landlocked) and arheic (no apparent drainage) areas (Hostetler, 1995). The formation and evolution of endorheic basins are associated with interplays between tectonics and climate. These basins are generally found in intracontinental settings and commonly relate to arid environments; they are presently collecting only about 2% of global river runoff (Garcia-Castellanos, Verges, Gaspar-Escribano, & Cloetingh, 2003). Endorheic basins and their lakes can be highly sensitive to variations in climate and adverse anthropogenic activities, such as overexploitation of water resources that affect the water inflow/outflow balance (e.g., Yapiyev, Sagintayev, Inglezakis, Samarkhanov, & Verhoef, 2017). A positive water balance will expand the lakes until equilibrium is found, while a negative balance will reduce and potentially fragment the lakes.

Half of the endorheic surface on the planet and most of the world's terminal lakes are presently located in Central Eurasia, where they prove to be excellent recorders of hydrological changes (e.g., Yapiyev et al., 2017). The Caspian Sea, world's largest endorheic lake has exhibited sea-level variations of ~4 m in the last 100 years, while the Aral Lake has lost 90% of its size in the last 50 years (Firoozfar, Bromhead, Dykes, & Neshaei, 2012; Kroonenberg, Badyukova, Storms, Ignatov, & Kasimov, 2000; Kroonenberg, Rusakov, & Svitoch, 1997). These basins are remnants of an ancient epicontinental sea called Paratethys (Laskarev, 1924). During the Oligocene to middle Miocene, the Paratethys Sea was connected to the global ocean via a series of interconnected basins with shallow sills (Harzhauser, Piller, & Steininger, 2002; Palcu, Golovina, Vernyhorova, Popov, & Krijgsman, 2017). Tectonic uplift of the Alpine-Carpathian mountain system was largely responsible for the progressive restriction and complete isolation of the Paratethys domain during the middle-late Miocene (ter Borgh et al., 2013; Kováč et al., 2007, 2017). In the Tortonian, Paratethys fragmented into two large endorheic lakes, Lake Pannon in central Europe (Magyar, Geary, & Müller, 1999; Magyar et al. 2013) and the Eastern Paratethys lake system that extended from Romania in the west to Turkmenistan in the east (Popov et al., 2004, 2006). The Eastern Paratethys Lake experienced major fluctuations in water level, reflecting the palaeoclimatic changes that occurred in Eurasia.

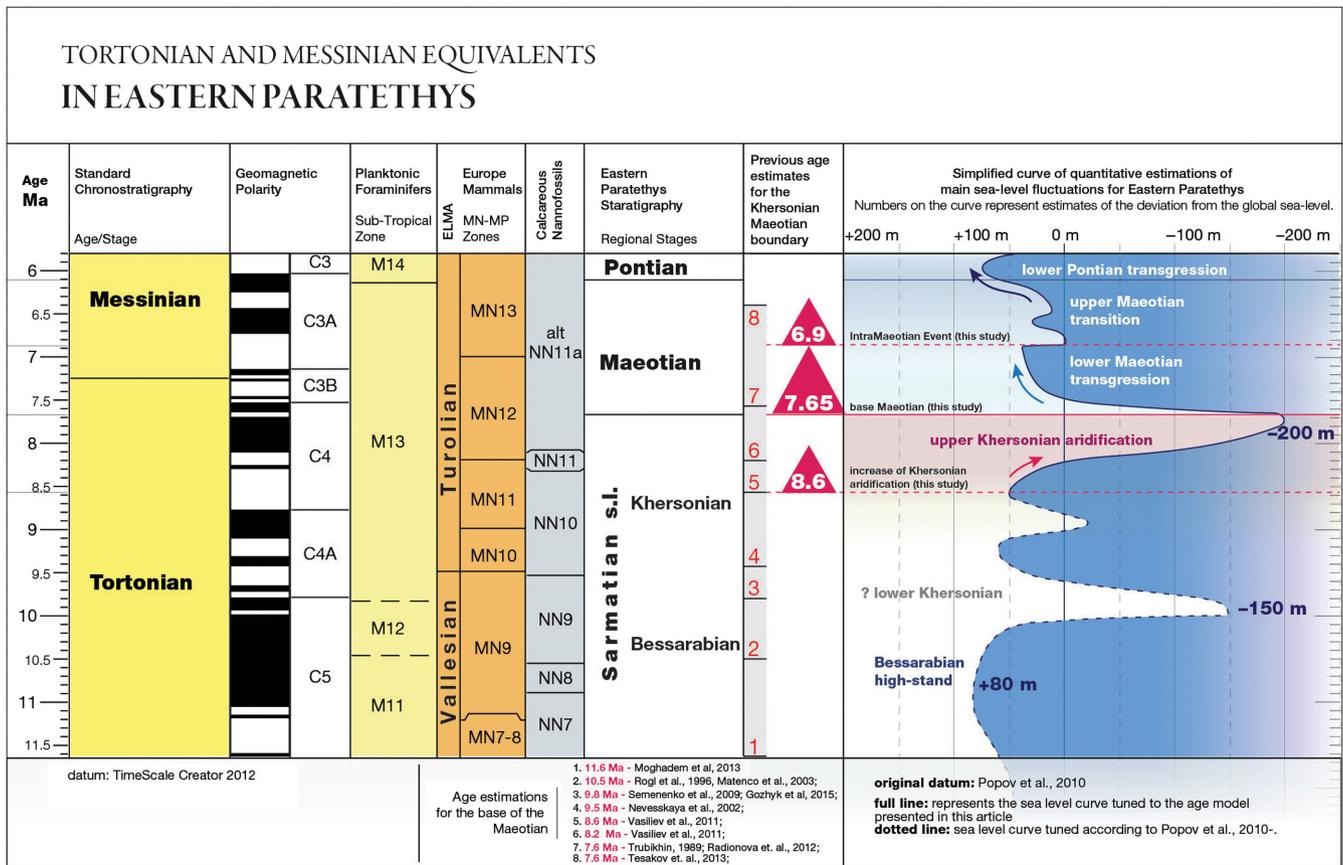
In the late Miocene, open steppe landscapes developed in Central Europe, which gradually migrated eastwards (Ivanov et al., 2011; Stekina, 1979). During the regional Khersonian substage (~9–8 Myr ago), these landscapes became increasingly xerophytic, suggesting phases of declining precipitation in the Paratethyan realm (Ivanov et al., 2011; Ivanov & Koleva-Rekalova, 1999; Stuchlik,

### Highlights

- A change to a positive water budget in the endorheic Paratethys ends a great drying phase in Central Eurasia during the late Miocene.
- We establish a magneto-biostratigraphic time frame for the middle Miocene deposits of the Eastern Paratethys based on three sections in the Dacian Basin (Romania).
- We date the end of the drying and the subsequent flooding, corresponding to the regional Khersonian/Maeotian boundary in the end of the Tortonian at an age of 7.65 Ma.
- The IntraMaeotian event reconnects Paratethys with the Mediterranean at an age of 6.9 Ma.
- Major sedimentary disturbances in Paratethys are linked with the IntraMaeotian Event and do not correspond to the Messinian Salinity Crisis.

Ivanov, & Palamarev, 1999; Ivanov et al., 2002; Böhme, Ilg, & Winklhofer, 2008; Böhme, Winklhofer, & Ilg, 2011). A dry climatic belt was established north-northeast of the Carpathians, gradually spreading to the south, towards the Euxinic Basin, and later occupying the southern parts of the Dacian Basin and the Pannonian Basin (Ivanov et al., 2011). Paratethys suffered from this dry period, which caused a major regression that split the basin into several smaller subbasins (Kojumdgieva, 1983; Popov, Antipov, Zastrozhnov, Kurina, & Pinchuk, 2010). Some of these subbasins became freshwater systems (Dacian Basin, Kuban Basin), while others became aberrantly brackish (Black Sea region) containing specific halite impoverished brine sediments produced by chemical precipitation of aragonite muds in warm, shallow-water conditions (Koleva-Rekalova, 1994; Ivanov & Koleva-Rekalova, 1999; Ivanov et al., 2002, Ivanov, Ashraf, & Mosbrugger, 2007).

The Maeotian transgression (Rögl, 1998), reconnected the Paratethys basins again putting an end to the late Miocene dry and endorheic phase in Central Eurasia. The age of the Maeotian transgression is seriously disputed with ~5 Myr age discrepancies between the different models (11.6 Ma—Moghadam, 2013; 10.5 Ma—Rögl, 1998; Maţenco, Bertotti, Cloetingh, & Dinu, 2003; 9.8 Ma—Semenenko, Andreeva-Grigorovich, Maslun, & Lulieva, 2009; Gozhyk, Semenenko, Andreeva-Grigorovich, & Maslun, 2015; 9.5 Ma—Nevevskaya, Goncharova, Ilyina, Paramonova, & Khondkarian, 2002; 8.85 Ma—Andrescu, 2009; 8.6 Ma—Vasiliev et al., 2011; 7.6 Ma—Trubikhin, 1989; Radionova et al., 2012) and the underlying tectonic or climatic triggering mechanisms are therefore still unexplained.



**FIGURE 1** Late Miocene stratigraphy of the Mediterranean and Paratethys (Hilgen et al., 2012, Popov et al., 2010, Steininger (1999); Palcu et al., 2015, 2017)

In this paper, we present a revised chronological framework for the upper Khersonian and Maeotian stages of Eastern Paratethys by applying integrated magneto-biostratigraphic dating to sections from the Dacian Basin of Romania. The magnetostratigraphic patterns from other independently dated sections in the Dacian and Black Sea basins are tentatively recalibrated to the new time frame. The resulting new age model allows us to date the main events in the late Miocene palaeogeographic evolution of Eastern Paratethys, which are discussed from a eustatic and basin connectivity evolution point of view.

## 2 | GEOLOGICAL BACKGROUND

At the beginning of the Oligocene, the northern branch of the Tethys Ocean lost its full connection with the global ocean and specific endemic fauna as well as water stratification and anoxia developed (e.g., van der Boon et al., 2018; Schultz, Bechtel, & Sachsenhofer, 2005). Laskarev (1924) introduced the name Paratethys for the Neogene records of this realm, which was later extended in time to cover the Oligocene as well (Báldi, 1980; Rusu, 1988). Paratethys was a heterogeneous system of epicontinental

seas and lake-seas that at its peak development, during the early Oligocene, stretched from the region north of the Alps, over Central Europe to Central Asia. Progressively, these basins became fragmented, isolated, and filled with sediments as a result of the continental collision between Africa-Arabia and Eurasia (Rögl, 1998). Periods of expansion to a large brackish sea (e.g., Popov et al., 2006; Rögl, 1998) and periods of fragmentation into smaller sub-basins (Kojumdjieva, 1983), accompanied by significant eustatic fluctuations (up to ~300m according to Popov et al., 2010) characterized Paratethys during the late Miocene.

Endemic ecosystems developed in both Central (Pannonian-Transylvanian) and Eastern Paratethys (Black Sea-Caspian Sea domain) that had little resemblance to the global ocean records. Consequently, regional stratigraphic scales were developed for both the Central and Eastern Paratethys domains, including regional stages that are defined on the basis of characteristic endemic (mainly mollusks and ostracods) faunal assemblages (e.g., Neveeskaya et al., 2002; Steininger et al., 1996). Stratigraphic correlations to the standard Geological Time Scale (GTS) are generally problematic (Piller, Harzhauser, & Mandic, 2007). In the absence of reliable biostratigraphic correlations, independent magnetostratigraphic and radio-isotopic age

determinations became crucial for direct correlations of the Paratethyan records to the GTS (ter Borgh et al., 2013; Hohenegger et al., 2009; Krijgsman, Stoica, Vasiliev, & Popov, 2010; Palcu, Tulbure, Bartol, Kouwenhoven, & Krijgsman, 2015; Palcu et al., 2017; Paulissen, Luthi, Grunert, Ćorić, & Harzhauser, 2011; Reichenbacher et al., 2013; Sant et al., 2017; Van Baak et al., 2013; Vasiliev, Krijgsman, Stoica, & Langereis, 2005; Vasiliev et al., 2004, 2010, 2011).

## 2.1 | Paratethys stratigraphy

The Tortonian and Messinian stages of the Mediterranean correspond to the Sarmatian *s.l.*, Maeotian and Pontian stages (Figure 1) of the Eastern Paratethys. The Sarmatian (Figure 1) regional stage was introduced by Suess (1866), based on his observations of highly endemic faunas in the Vienna basin sediments and the correlations to the faunas from the area north of the Black Sea. The Sarmatian deposits of the Vienna basin are relatively short-lived as the region became isolated at the beginning of the Tortonian forming Lake Pannon (11.6 Ma; Paulissen et al., 2011; ter Borgh et al., 2013). In Eastern Paratethys, Sarmatian-type of depositional environments continued much longer in time resulting in two awkwardly different definitions for the Sarmatian: a Sarmatian *sensu-strictum* (*s.s.*) and a Sarmatian *sensu-lato* (*s.l.*). The Sarmatian *s.l.*, applicable only to the Eastern Paratethys, was again divided into lower, middle, and upper Sarmatian substages by Andrussov (1899). These substages were later renamed Volhynian, Bessarabian, and Khersonian by Simionescu (1903). Due to the inherent correlation issues, it was strongly advised to abandon the use of the term “Sarmatian” in Eastern Paratethys stratigraphy and follow the substage terminology there (Lukeneder, Zuschin, Harzhauser, & Mandic, 2011; Harzhauser & Piller, 2004a, b). In this study, we will employ the terms Bessarabian and Khersonian, the Sarmatian-Maeotian boundary being thus referred to as the Khersonian/Maeotian boundary.

The Bessarabian (referred to as Basarabian by Simionescu, 1903) comprises the deposits between the Volhynian (characterized by *Ervilia*, *Plicatiforma plicata*) and Khersonian (characterized by *Maetra caspia* and *M. bulgarica*) and represents the peak development of the endemic Sarmatian mollusk fauna. These mollusks are characteristic for low salinity (Ionesi, Ionesi, Lungu, Roșca, & Ionesi, 2005), fitting in the context of a significant expansion of Paratethys (Popov et al., 2004).

The Khersonian substage (referred to as Hersonian by Simionescu, 1903; and also referred to as Kersonian or Chersonian by other authors) is defined by the presence of low-brackish mollusk associations with *Maetra caspia*, to which *Maetra bulgarica* was later added. The upper part of the Khersonian, characterized by freshwater mollusk taxa, generally covers the predominantly regressive phase

marked by the expansion of continental deposits and a fragmentation of the Paratethys basin (Kojumdgieva, 1983). Impoverished faunal assemblages (Ionesi et al., 2005) and diverse, fresh water to hypersaline environments characterize this substage. The end of the Khersonian is marked by a major lake regression: the Eastern Paratethys water level dropped 200–300 m (Jones & Simmons, 1997; Popov et al., 2006; Tugolesov, Gorshkov, Meisner, Solov'ev, & Khakhalev, 1985) exposing large swathes of the shelf that consequently became eroded. The northern Black Sea shelf became exposed and incised by rivers in large systems of canyons, such as the > 150 km long Burukshun canyon in southern Russia (Popov et al., 2010).

The Maeotian stage, also referred to as Meotian, was created by Andrussov (1890) to describe the transitional layers between the Sarmatian *s.l.* and the Pontian, which contain both brackish and marine faunal elements. It replaced the “Ante-Pontian” or the “upper Kerch limestone,” previously described by Andrusov (1885) and subdivided it in three parts: 1. brackish levels, with *Cerithium*, *Tapes*, *Maetra*, *Ervilia*, *Cardium*; 2. new fauna: *Dossinina*, *Scorbicularia*, *Hydrobia*; 3. Freshwater fauna with *Helix*, *Pupa*, *Planorbis*. The Maeotian was later subdivided into the lower (Bagerovian) and upper (Akmanaian) substages by Karlov (1937).

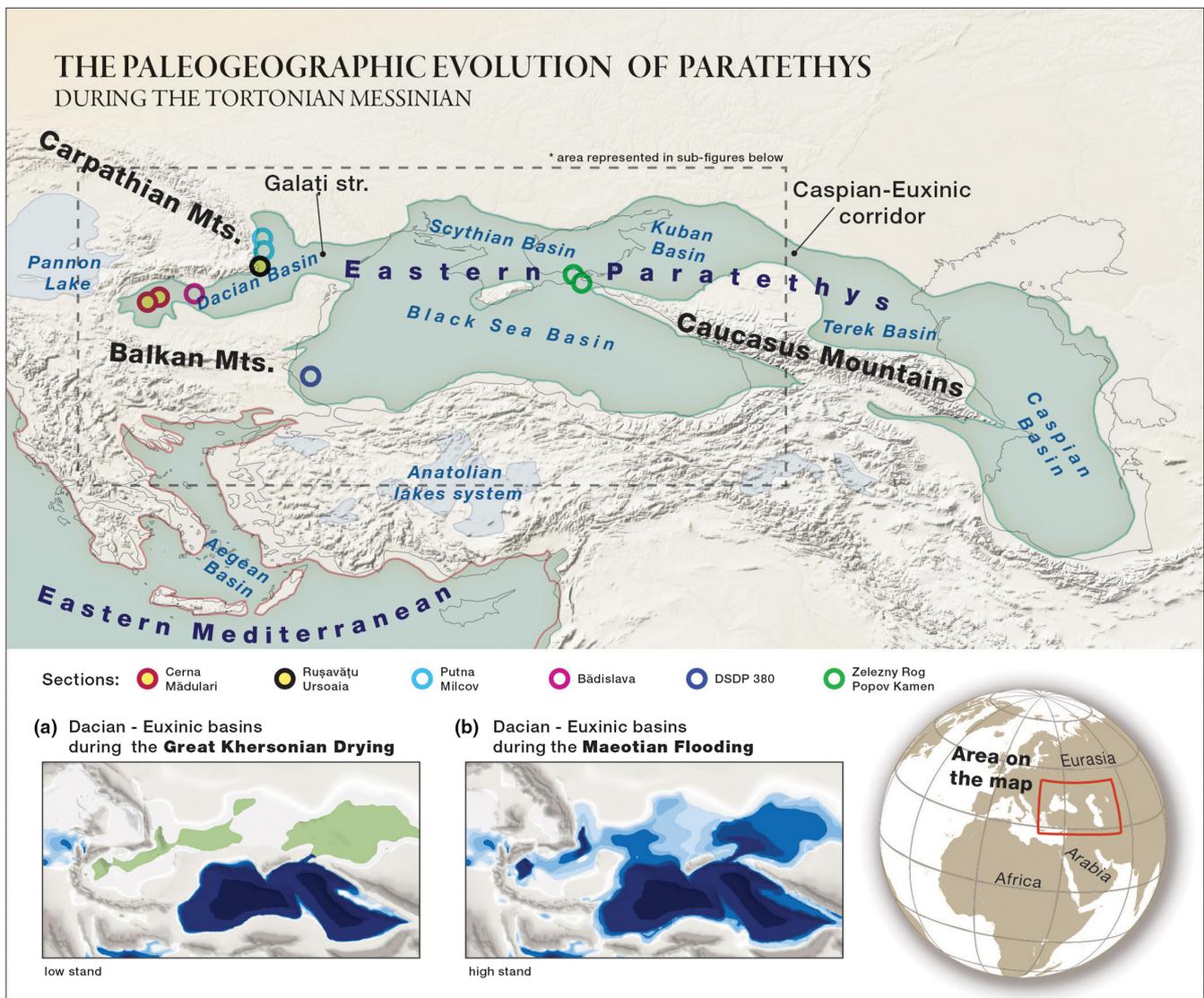
The Maeotian begins with a strong transgression, estimated at ~300 m by Popov et al. (2010), that reached its maximum flooding level in the late early Maeotian (Neveeskaya et al., 1986; Stevanovic & Ilyina, 1982). According to Neveeskaya et al. (2002), the lower Maeotian (Bagerovian) is characterized by the presence of endemic Mediterranean bivalves (*Dosinia*, *Mytilaster*, *Venerupis*, *Arba*, and *Ervilia* genera), gastropods (*Rissoa*, *Mohrensterina*, *Cerithium*, *Potamides*, and *Bittium* genera), euryhaline foraminifera (*Quinqueloculina*, *Elphidium*, *Articulina*, *Discorbis*, etc.), and ostracods (*Leptocythere*, *Xestoleberis*, and *Loxoconcha*), genera that are commonly interpreted as markers of abnormal marine environments. These Maeotian species are generally linked with the Aegean region (North Greece, Serres, Dafni Formation), where mollusk associations are similar to the lower Maeotian of Paratethys (Popov & Neveeskaya, 2000; Stevanovic & Ilyina, 1982). The upper Maeotian (Akmanaian) is characterized by brackish-water mollusks associated with scarce marine taxa. These assemblages hold evidence of episodic connections with the global ocean (Radionova & Golovina, 2012).

## 2.2 | The Dacain Basin

The Dacian Basin, located in between the Southern Carpathians, the PreBalkan area and the Dobrogea High, evolved as a palaeogeographic entity after the isolation of the Central Paratethys (Pannonian and Transylvanian basins) from the Eastern Paratethys at 11.6–11.3 Ma by tectonic uplift of the

Carpathian Mountains (ter Borgh et al., 2013; Saulea, Popescu, & Sandulescu, 1969; Vasiliev et al., 2010). The Carpathian Mountains were also the main source area of the thick middle Miocene deltaic units that onlap the northern margin of the basin (ter Borgh et al., 2014; Jipa, Stoica, Andreescu, Floroiu, & Maximov, 2011; Marinescu, 1978). During the middle Miocene considerable tectonic activity took place in the area and largely ceased in the western part of the basin (Getic Depression) by the end of the middle Miocene, when other factors that affect relative sea and lake levels, such as climate and basin connectivity, gained importance. The late Miocene time interval in the western part of the Dacian Basin is thus represented in the post-conglomeratic sequences that comprise parts of the Maeotian stage and the complete Pontian stage (Stoica et al., 2013). In the eastern part of the Dacian basin, the Focșani Depression comprises the thickest

Neogene sedimentary cover of the Dacian Basin, approaching ~13 km (Tărăpoancă, Bertotti, Mațenco, Dinu, & Cloetingh, 2003). Excellent exposures are on the western flank of the basin along the almost continuously outcropping river sections (Andreescu & Papaianopol, 1970; Andreescu, 1973, 1975; Andreescu and Ticleanu, 1976; Vasiliev et al., 2004). These sections consist of upper Bessarabian–Maeotian alternating shallow sandstones and shales (Saulea et al., 1969) and in the upper part (Pontian to Romanian) of shales, siltstones, sandstones, and coals (Pană, 1966; Grasu, Catana, & Bobos, 1999; Panaiotu, Vasiliev, Panaiotu, Krijgsman, & Langereis, 2007). Previous studies indicate that the late Miocene to Pliocene sedimentary successions are rather continuous without major hiatuses and unconformities (Tărăpoancă et al., 2003; Vasiliev et al., 2004; Van Baak, Vasiliev, Palcu, Dekkers, & Krijgsman, 2016).



**FIGURE 2** The palaeogeography of the Paratethys realm during the late Tortonian – early Messinian after Popov et al. (2006) and their palaeoenvironments. Medallions show the contracted (a) and expanded (b) Paratethys basin

### 3 | SECTIONS; LITHOLOGY AND PALAEOLOGY

Here, we focus on determining the age of the Khersonian/Maeotian and lower/upper Maeotian boundaries in two composite key sections located in the Dacian basin: the Ruşavăţu section in the Focşani Depression and the Cerna and Cernișoara sections in the Getic Depression, more to the west (Figure 2). These sections contain long and continuous sedimentary successions comprising Khersonian, Maeotian, and Pontian deposits that permit extensive correlations with other sections inside the Dacian Basin and outside, in the Black Sea and Indol-Kuban (Azov) basins. The sections were logged in detail and sampled for biostratigraphic, palaeomagnetic, and geochemical purposes. In Cerna-Cernișoara, our biostratigraphic study is strictly directed on locating the basal Maeotian levels while in Ruşavăţu the focus is on developing a detailed distribution chart of foraminifera and ostracods. Macrofossils are well preserved in the relatively shallow Ruşavăţu section from the beginning of the Maeotian upwards. These fossils are extensively described in the studies of Pană (1969) and were used as reference for logging the Ruşavăţu section. Macrofossils are almost absent in the Cerna-Cernișoara section, probably because the depositional environment is too deep and partly anoxic.

#### 3.1 | The Ruşavăţu section

The Ruşavăţu syncline sections preserve a very rare, rather complete lithological succession of middle-upper Miocene age. The key section is located on the Ruşavăţu river (45°15' 57.81"N/26°25'31.91"E), downstream from the Grădina Corbilor (*Ravens Garden*) hill. The sedimentary succession, with stratigraphic boundaries based on mollusk fauna (Pană, 1966), consists of sands and silts of Bessarabian age (with mollusk levels rich in *Maetra* sp.), continental deposits covering most of the Khersonian, lacustrine-semi-marine deposits of Maeotian age and marine-semi-marine clays of lower Pontian age.

Most of the Khersonian is represented by a succession of continental deposits dominated by palaeosols with colours that vary from green to yellow and dark reddish-brown (Figure 3c). The palaeosols are occasionally interrupted by whitish silts in the lower part, purple-green sands and microconglomerates in the middle part, and whitish-brown sands in the upper part.

The Maeotian starts with dark-grey clays that follow directly on top of the palaeosols. The base of the Maeotian is marked by the so-called “*Conger* clays” (1441–1446 m, Figure 3b), starting with a 0.2 m thick clay bed that contains only uncarenated *Conger* shells (ex. gr.

modioliforms). Upwards these are replaced by *Unio*, *Viviparus* (freshwater taxa) and then followed by *Littorina banatica* Jek., *L. politioanei*, *Pseudamnicola* sp., *Hydrobia vitrella* Ștef.. In this latter unit, the bryozoa *Membranipora* sp. and the benthic foraminifer *Ammonia beccarii* Linne., have also been described. The next package (1446–1480 m) begins with coarse sands with *Hydrobia* sp. at the base, followed by a middle unit of anoxic marls that are followed, in turn, by a package of silty marls. The following unit (1480–1510 m, Figure 3d) is made of fine, brown-grey sandy marls, sands, and sandstones. At the base (1483 m), a marker level characterized by carinated congerias (Figure 3e) stands out. This marker level can be followed throughout the region. The rest of the unit is characterized by silts, sands, and poorly cemented sandstones with *Unio* sp. and *Theodoxus* sp.. The upper part of the lower Maeotian (1510–1539 m) consists of alternations of anoxic clays, sandy marls, and sands rich in fauna. Four faunal reference levels are observed: a) *Ervilia minuta* and *Dosinia maeotica* sandstones; b) *Pirenella* sandy marls; c) *Dosinia maeotica* coquina (Figure 3f), and d) *S. tellinoides* in coarse dark sands rich in *Hydrobia* (Pană, 1969).

This fauna trend indicates that salinity values, although fluctuating, increase (to >15‰) towards the end of the lower Maeotian. The top of the lower Maeotian (1539 m) shows an abrupt ending of the higher salinity fauna.

The sediments of the upper Maeotian are characterized by low salinity faunistic elements that begin at 1539 m. Coarse-grained sediments characterize the first 20 m of the upper Maeotian succession (1539–1559 m), often with soft deformation patterns in the finer levels (Figure 3g). These levels consist of thick sand packages, occurring in alternations with poorly cemented sandy levels, sandstones, dark coarse silts, or loose sands with well-rounded concretions. Faunistically, this interval is characterized by the presence of intact freshwater mollusk shells (e.g., *Unio* sp., Figure 3i) and shell fragments from brackish taxa (e.g., *Dosinia maeotica*, *Paphia* sp.)

The coarse levels at the base of the upper Maeotian correspond to an important palaeoecological change in the basin, defined as the Intra-Maeotian Event (IME), placed at the sharp transition from brackish to low salinity-freshwater environments (Pană, 1969).

The upper Maeotian (1559–1668m) succession consists of a monotonous package of marls and sands, occasionally interrupted by subordinate oolitic sands and a red-brownish microconglomerate rich in andesitic clasts (Figure 3j). The top of the Maeotian succession is located below two coquina levels with fragments and shells of *Conger* *novorossica* (1668 m). The conformably overlying Pontian sediments show a shift towards finer-grain size, consisting in silty clays with abundant *C. novorossica* shells.

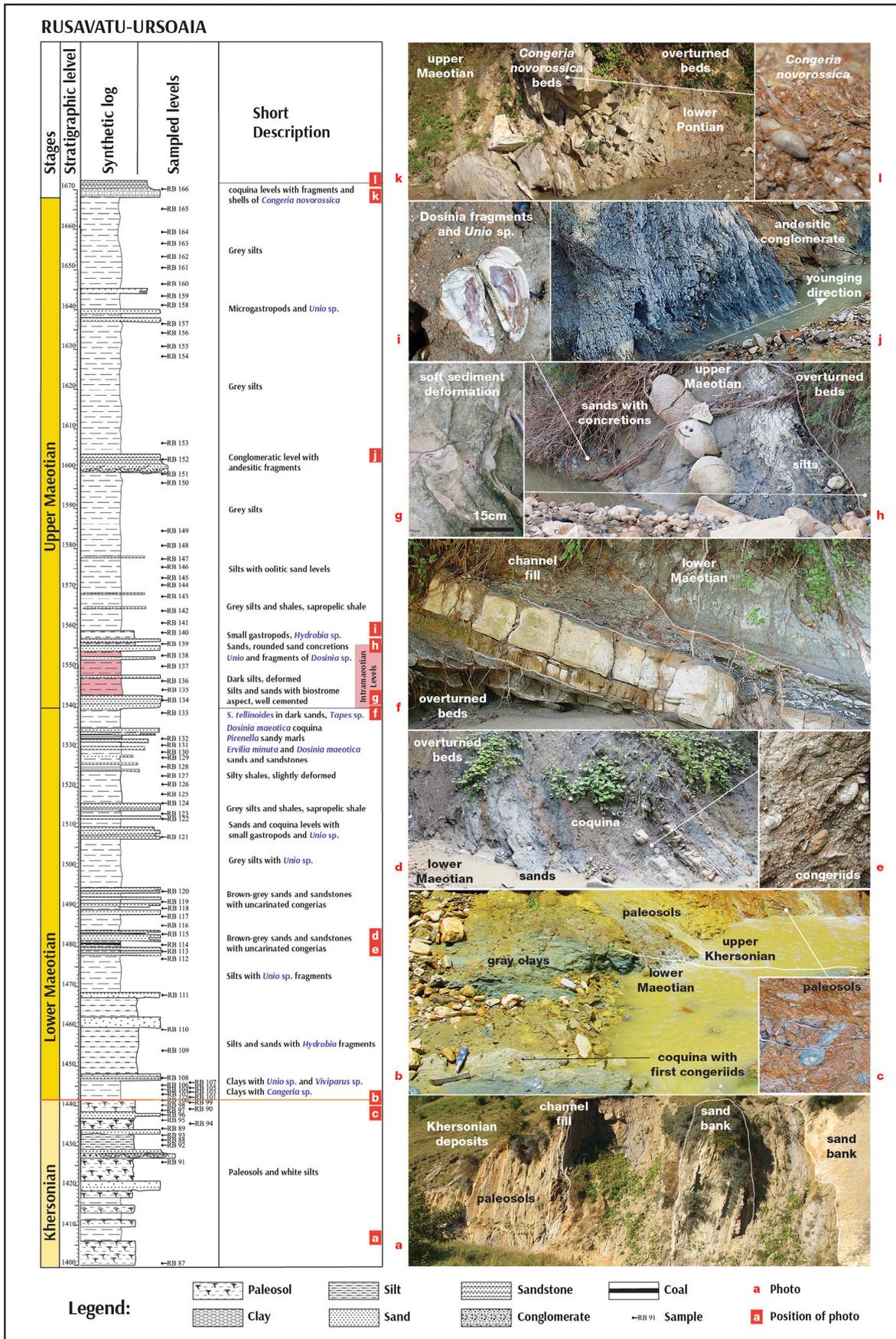
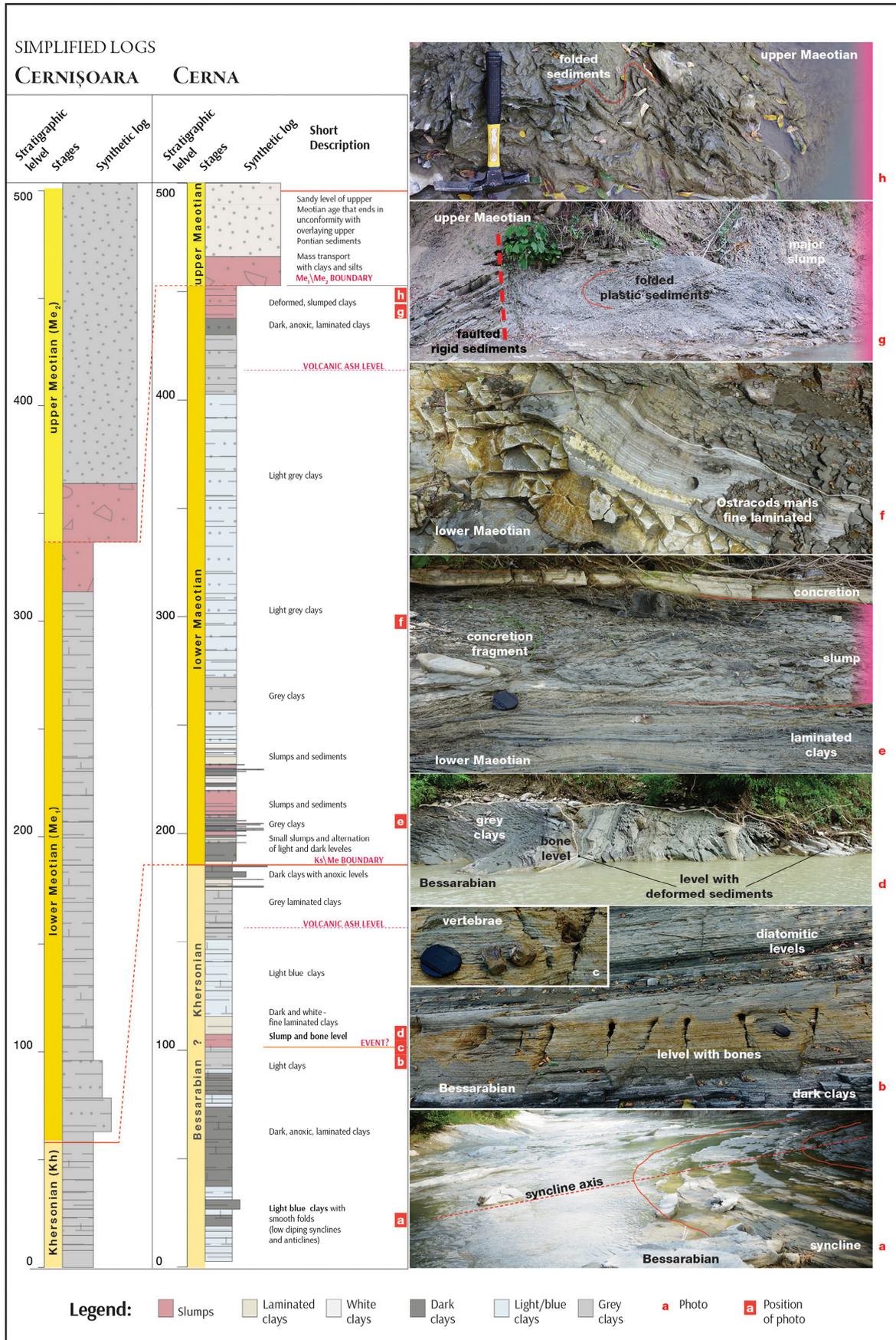


FIGURE 3 Bio-litho-stratigraphic overview of the Ruşavăţu river section



**FIGURE 4** Bio-litho-stratigraphic overview of the Cerna and Cernișoara river sections

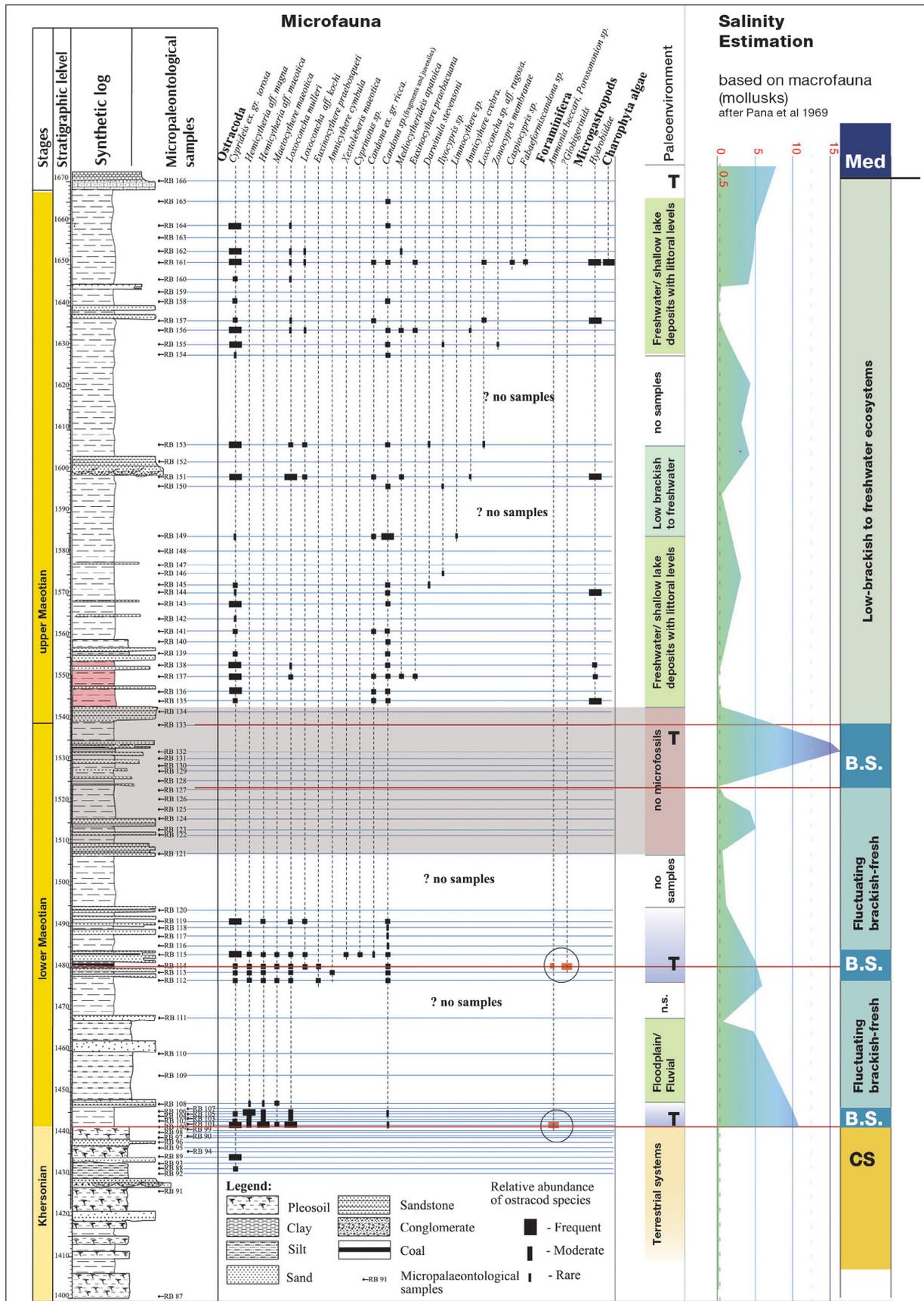


FIGURE 5 The micropalaeontological fauna from Rușavățu, correlations with the mollusk records (Pană, 1969). CS-continental, BS-Black Sea basin influence, MED-Mediterranean influence, T-transgression

### 3.2 | The Cerna-Cerņișoara section

The Cerna and Cerņișoara sections (Figure 4) preserve relatively deep-facies deposits of middle-late Miocene age, cropping out along the valleys of the Cerna and Cerņișoara (also referred to as Mădulari by Vasiliev et al., 2005) rivers. The sediments consist of massive and laminated clays of Khersonian and early Maeotian age and sands of late Maeotian age, with the stratigraphic boundaries based on microfossil fauna (Fongngern et al., 2017). The Cerna section has been sampled downstream from the villages of Stroești (45°5'32.23"N/23°53'52.48"E) and while the parallel Cerņișoara section has been sampled downstream from the village of Obârșia (45° 4'21.47"N/23°58'47.29"E). Both sections display similar lithology and can be correlated based on lithological characteristics.

The Bessarabian–Khersonian deposits are hard to differentiate, as macrofossil content is scarce in both sections. Lithological changes and sedimentary events (slumps) deliver a basic reference frame supported by microfossil assemblages. We follow here the lithological description of Fongngern et al. (2017) and aim to improve it with additional micropalaeontological information. The undifferentiated upper Bessarabian and Khersonian deposits consist of massive dark clays and laminated, diatom rich, clays both occasionally interrupted by centimetre thick lacustrine carbonate concretions. The succession is quite monotonous except for an instability level (5 m thick landslide of slumped sediments) and a 20–30 cm thick ash layer that occurs in the lower (Khersonian) part of the sampled interval of the Cerna section (~160 m). The ash layer was recently dated by the U-Pb technique at ~8 Ma (Fongngern et al., 2017). Sulphide and jarosite coatings in the dark clays of the uppermost Khersonian suggest widespread anoxia at the end of this substage.

A significant lithological shift, from dark anoxic clays to light grey clays with levels rich in microclasts (190 m), is correlated with the base of the Maeotian (180–190 m). Upwards in the successions, the microclasts levels merge into small mass transport packages that alternate with in situ deposition. Changes in lithology, from dark to light-fine laminated clays continue to occur throughout the succession, and bedding orientation remains constant until the

marked mass transport level (~440 m) that marks the Intra-Maeotian Event (IME) at the end of the lower Maeotian. The upper part of the sections contains sediments of upper Maeotian age, mostly massive sands that overlie the slumps. A sedimentation gap separates the upper Maeotian from the younger Pontian deposits.

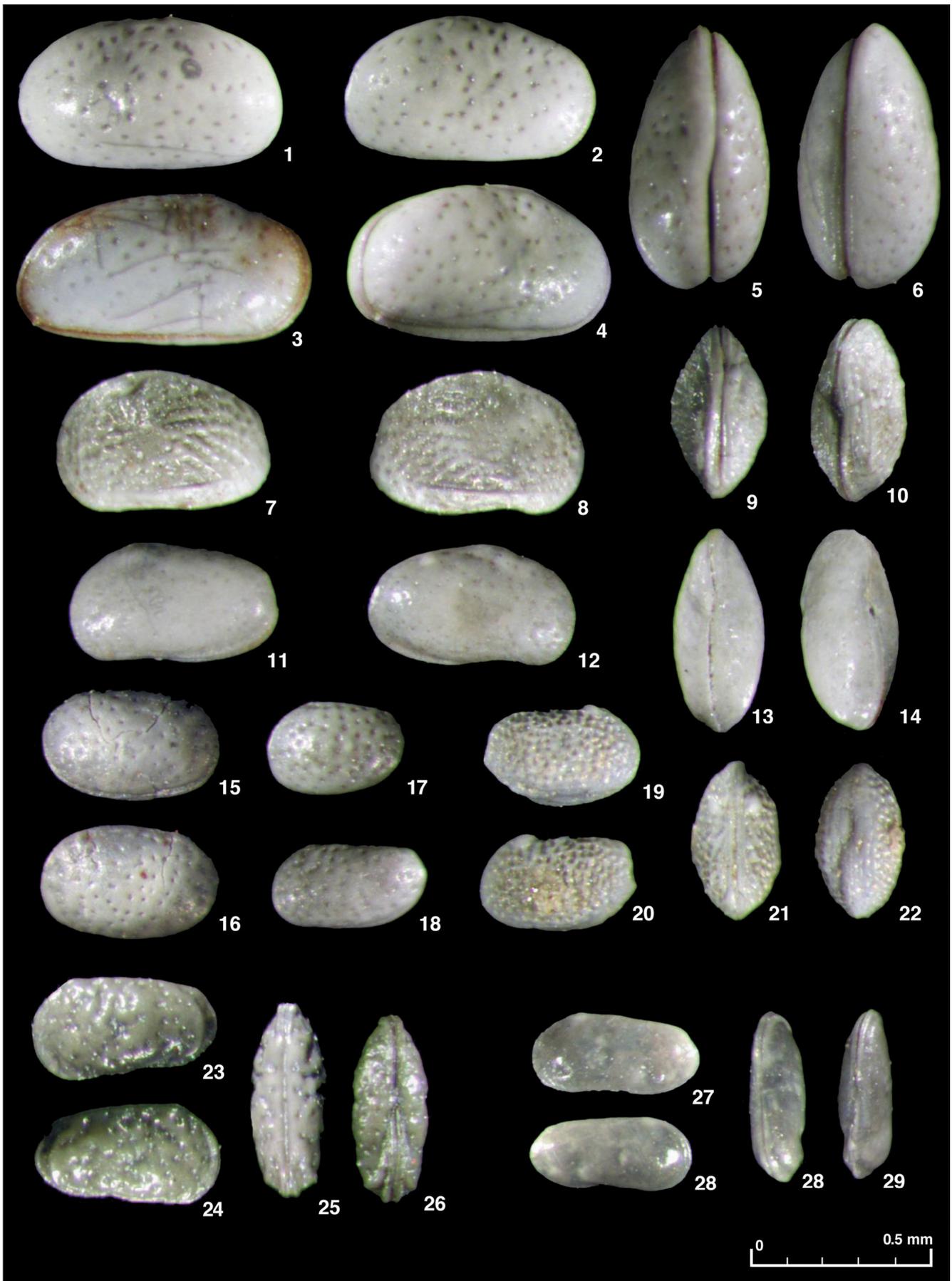
### 3.3 | Ostracod biostratigraphy

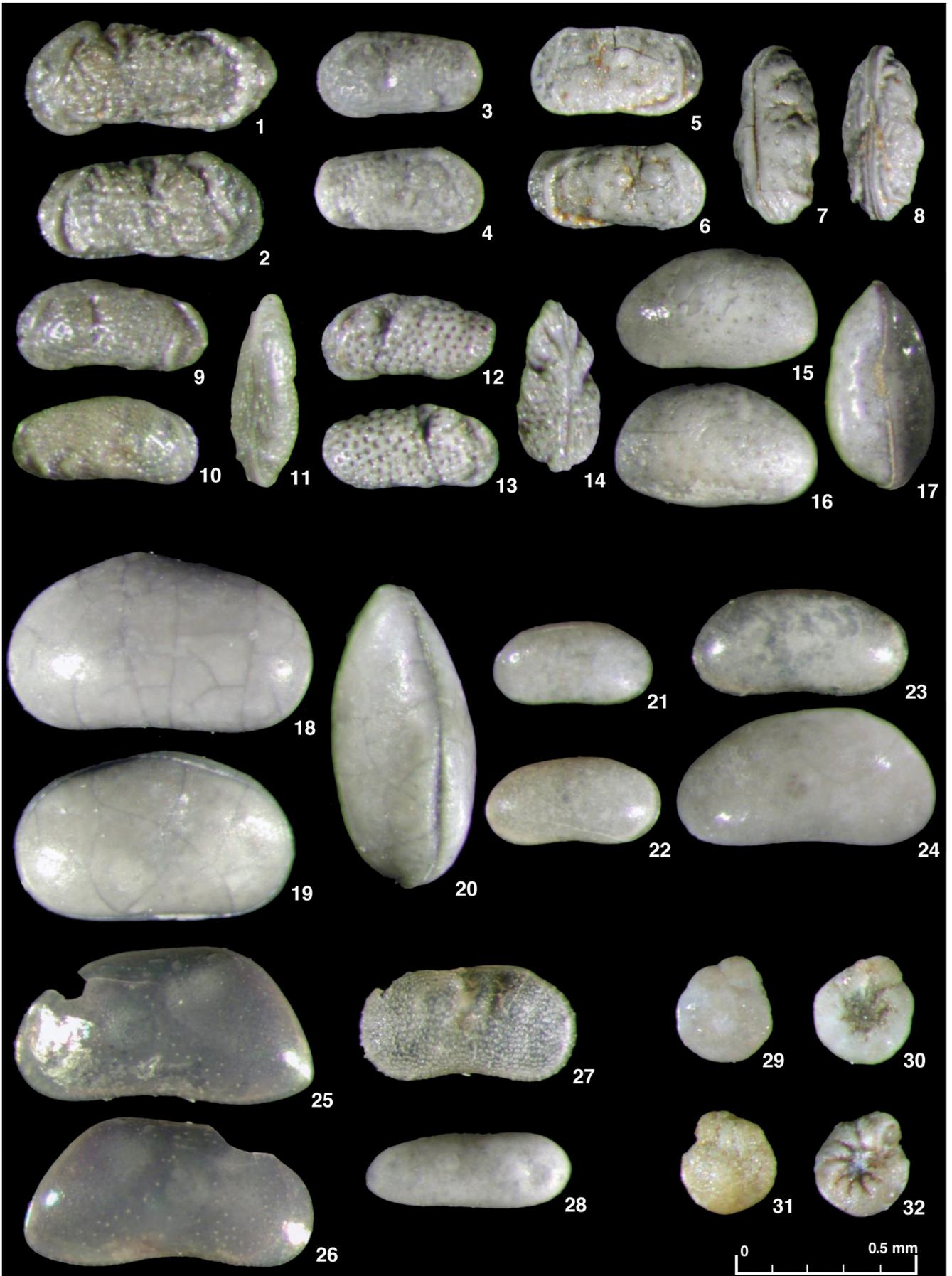
The ostracod fauna from the Khersonian–Maeotian deposits of the Rușavățu section display a temporal distribution influenced by frequent environmental changes, mainly in salinity and water depth. (Figure 5). The Khersonian part of the section shows a very poor ostracod assemblage indicative of a predominantly terrestrial environment. Some lacustrine–brackish water intercalations in the uppermost Khersonian provided an ostracod association dominated by the euryhaline species *Cyprideis torosa*.

An important event took place at the Khersonian/Maeotian transition (1440–1450 m), revealing an increase in brackish-marine ostracod species such as *Loxoxoncha mulleri*, *L. aff. mulleri*, *Maeotocythere aff. maeotica*, *Hemicytheria aff. magna*, *Hemicytheria maeotica* together with *Cyprideis torosa* (Figure 5,6,7). Benthic calcareous foraminifera like *Ammonia beccarii*, *Porosonion* sp. are also frequent at the same stratigraphic level. These ostracods and foraminifera suggest an influx of saline waters that took place at the base of the Maeotian, changing the depositional conditions to shallow brackish water with salinities of approximately 10–12 g/l. The hemicytherids, as well as some loxoxonchid ostracods identified at Rușavățu are also known to have existed in large numbers in the endemic Pannon Lake during the Pannonian regional stage (<11.6 Ma) (REF). The presence of these species at this particular interval confirm that a connection between the Pannon Basin and the Dacian Basin was established during the lower Maeotian transgression event (e.g., ter Borgh, Stoica, Donselaar, Matenco, & Krijgsman, 2014).

The Maeotian flooding event is followed by a more restrictive period when brackish environments are replaced by a more proximal or terrestrial setting where ostracods are absent (1450–1475). A new rejuvenation of ostracod fauna took place in the middle part of the lower Maeotian

**FIGURE 6** Representative ostracods from Maeotian – Rușavățu Section. ((LV-left valve, RV-right valve, C – carapace): 1–6. *Cyprideis torosa* (Jones), smooth specimens; 1. LV, external view; 2. RV, external view; 3. C, view from RV, male; 4. RV, view from RV, female; 5. C, dorsal view; 6. C, ventral view. 7–10. *Hemicytheria aff. magna* Olteanu; 7. LV, external view; 8. RV, external view; 9. C, dorsal view; 10. C, ventral view. 11–14. *Hemicytheria aff. maeotica* Olteanu; 11. RV, external view; 12. RV, external view; 13. C, dorsal view; 14. C, ventral view. 15, 16. *Loxoxoncha mulleri* (Méhes); 15. C, view from RV; 16. C, view from LV. 17, 18. *Loxoxoncha aff. mulleri* (Méhes); 17. LV, external view, female; 18. LV, external view, male. 19–22. *Loxoxoncha aff. kochi* Méhes; 19. C, view from RV; 20. C, view from LV; 21. C, dorsal view; 22. C, ventral view. 23–26. *Maeotocythere aff. maeotica* Livental. 23. LV, external view; 24. RV, external view; 25. C, dorsal view; 26. C, ventral view. 27–29. *Ammicythere cymbula* (Livental); 27. LV, external view; 28. RV, external view; 28. C, dorsal view; 29. C, ventral view





**FIGURE 7** Representative ostracods from Maeotian –Rușavățu Section. ((LV-left valve, RV-right valve, C – carapace): 1, 2. *Euxinocythere* aff. *praebosqueti* Suzin. LV, external view, male.; 2. RV, external view. 3, 4. *Amnicythere* sp.; LV, external view; 4. RV, external view; 5–8. *Euxinocythere* aff. *praebacuana* (Liventali); 5. LV, external view; 6. RV, external view; 7. C, dorsal view; 8. C, ventral view. 9–11. *Mediocytherideis apatoica* (Schweyer); 9. LV, external view; 10. RV, external view. 11. C, ventral view. 12–14. *Mediocytherideis* sp.; 12. LV, external view; 13. RV, external view; 14. C, dorsal view. 15–17. *Xestoleberis meotica* Suzin; 15. C, view from LV; 16. C, view from RV; 17. C, ventral view. 18–20. *Cyprinotus* sp.. 18. C, view from LV; 19. C, view from RV; 20. C, dorsal view. 21–24. *Candona* sp., different juvenile stages. 25, 26. *Candona* aff. *ricca* Stancheva; 25. LV, external view; 26. RV, external view. 25. *Ilyocypris* sp., RV, external view. 26. *Darwinula stevensoni* (Brady & Robertson), LV, external view. 27–30. *Ammonia beccarii*

(1475–1500), possibly related to a second salinity increase. Besides the species mentioned above, other brackish water ostracods occurred for the first time in the section: *Loxiconcha* aff. *kochi*, *Euxinocythere praebosqueti*, *Amnicythere cymbula*, *Xestoleberis meotica*, *Cyprinotus* sp. This second level rich in brackish ostracods contains also calcareous benthic foraminifera, *Ammonia* and *Porosonion* species, as well some planktonic globigeriniid species.

The upper Maeotian (1540–1670 m) is dominated by fresh water environments with some brackish intercalations. The fresh water intervals contain ostracod associations dominated by candonids, juveniles, or fragments of *Candona* sp., *Pseudocandona*, *Candona* aff. *ricca*, *Fabaeformiscandona* sp., together with rare *Ilyocypris* sp., *Darwinula stevensoni*, *Zonocypris membranae*, *Limnocythere* sp. Brackish intercalations contain species of *Loxiconcha*, mainly *L.* aff. *mulleri*, *L.* aff. *kochi*, *L.* aff. *rugosa*, *Euxinocythere praebacuana*, *Amnicythere* sp., *A. crebra* as well as *Mediocytherideis apatoica* and *M.* sp. This typical upper Maeotian ostracod assemblage is replaced by Pontian ostracod fauna, mainly of Pannonian type, after the Pontian flooding event of the Eastern Paratethys at 6.1 Ma (e.g., Floroiu, Stoica, Vasiliev, & Krijgsman, 2011; Krijgsman et al., 2010; Stoica et al., 2013; Van Baak et al., 2016, 2017).

#### 4 | PALAEOMAGNETIC PROPERTIES

The palaeomagnetic properties of the sediments from Cerna, Cernișoara, and Rușavățu have been investigated following the standard methodology, described in detail in the supplementary material attached to this article and Fig. S1.

The NRM intensity ranges  $1.5 \cdot 10^{-5}$  A/m to  $4.6 \cdot 10^{-2}$  A/m for the Rușavățu samples,  $4.9 \cdot 10^{-5}$  A/m to  $5.6 \cdot 10^{-2}$  A/m for the Cerna and Cernișoara samples. Several characteristic thermal demagnetization diagrams of mostly marls and clays are depicted in Figures 8, 9 and 10. We identified the ChRM by analyzing the decay-curves and vector end-point diagrams (Zijderveld, 1967). In both the thermal demagnetization and the alternating field demagnetization diagrams, two magnetic components can be recognized. A very weak, low-temperature, viscous overprint is generally removed at

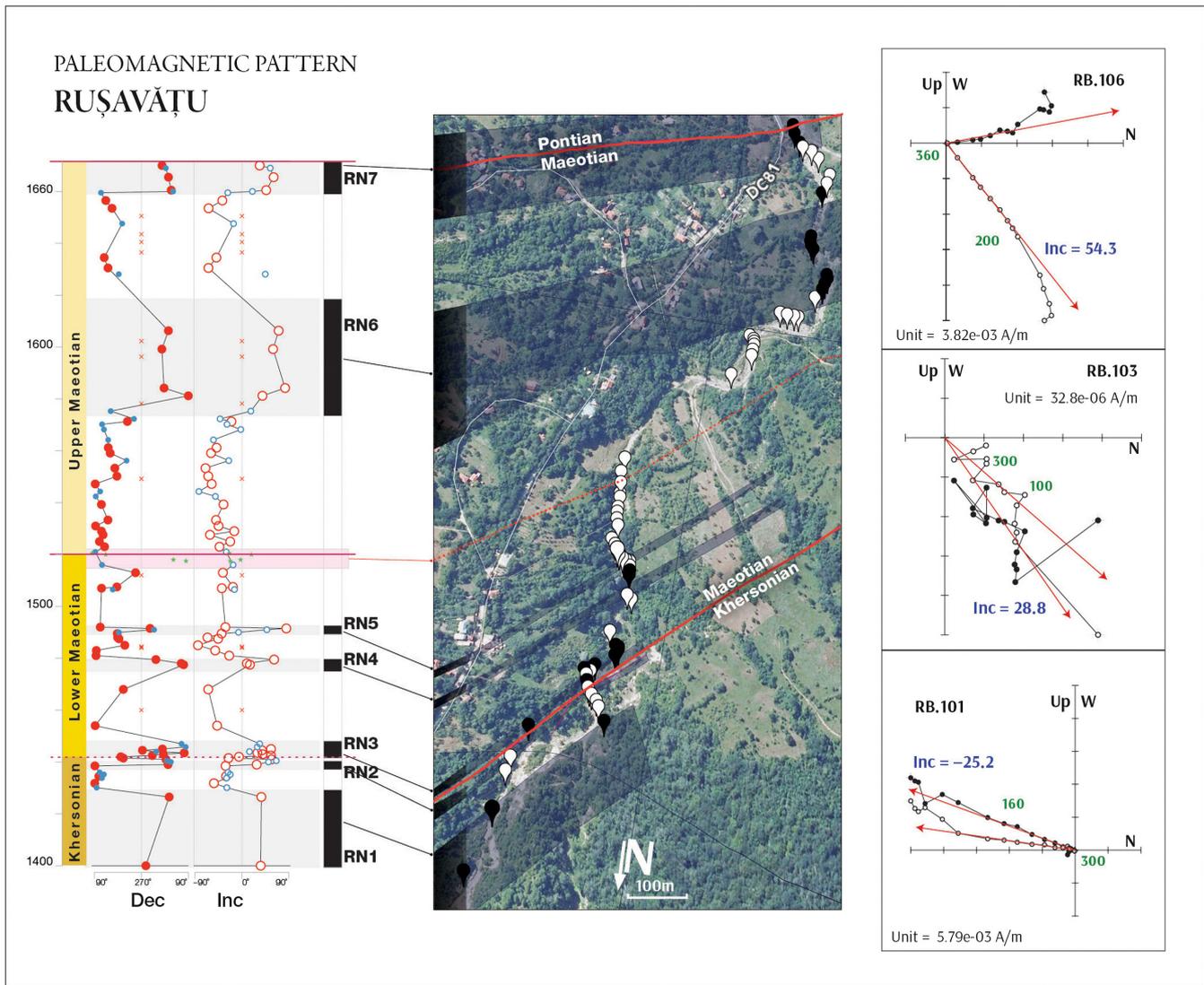
150°C and 15mT (Figures 8c,d,e; 9c,d,e and 10c,d,e). The second, higher temperature, component is demagnetized at temperatures between 120°C and 400°C. This component is of dual polarity and is interpreted as the ChRM. The ChRM directions are all determined by four or more consecutive temperature steps and calculated with the use of principal component analysis (Kirschvink, 1980).

We use the maximum angular deviation (MAD) of the calculated directions to separate the results into three qualitative groups. The 1st quality (MAD = 0–5) and 2nd quality groups (MAD = 5–10) have all been plotted against stratigraphic levels to determine the polarity pattern of the sections (Figures 8,9,10). The 3rd quality results (MAD > 10) have not been taken into account when interpreting the magnetostratigraphic pattern.

The polarity pattern of the Rușavățu section comprises 13 polarity intervals, 6 of reversed (RR1-6), and 7 of normal (RN1-7) polarity. The polarity pattern of the Cerna section comprises 17 polarity intervals, 8 of reversed (CR1-8), and 9 of normal (CN1-9) polarity. The polarity pattern of the Cernișoara section shows 13 different polarity intervals, 6 of reverse (XR1-6), and 7 of normal (XN1-7) polarity.

#### 5 | A MAGNETOSTRATIGRAPHIC AGE MODEL FOR THE MAEOTIAN STAGE

We aim to obtain a robust time frame for the Khersonian–Maeotian deposits of the Dacian Basin relying on the palaeomagnetic polarity patterns of the Rușavățu section (Figure 11d), located in a marginal piggyback basin that is characterized by relative quiet sedimentation settings (Pană, 1969), and the polarity pattern of the Cerna and Cernișoara sections (Figure 11c), indicative for the relatively deep parts of the basin (Fongngern, Olariu, Steel, & Krézsek, 2016; Fongngern et al., 2017). Two volcanic ash layers from Cerna and Cernișoara, from levels situated above and below the Khersonian/Maeotian boundary have previously been dated with the U-Pb method and constrain the Khersonian/Maeotian boundary to the 7.5–8 Ma age interval (Fongngern et al., 2017). An alternative age model, however, was proposed by Andreescu (2009) based on palaeomagnetic results by Snel, Mărunțeanu, Macaleț, Meulenkamp, and Van Vugt (2006),

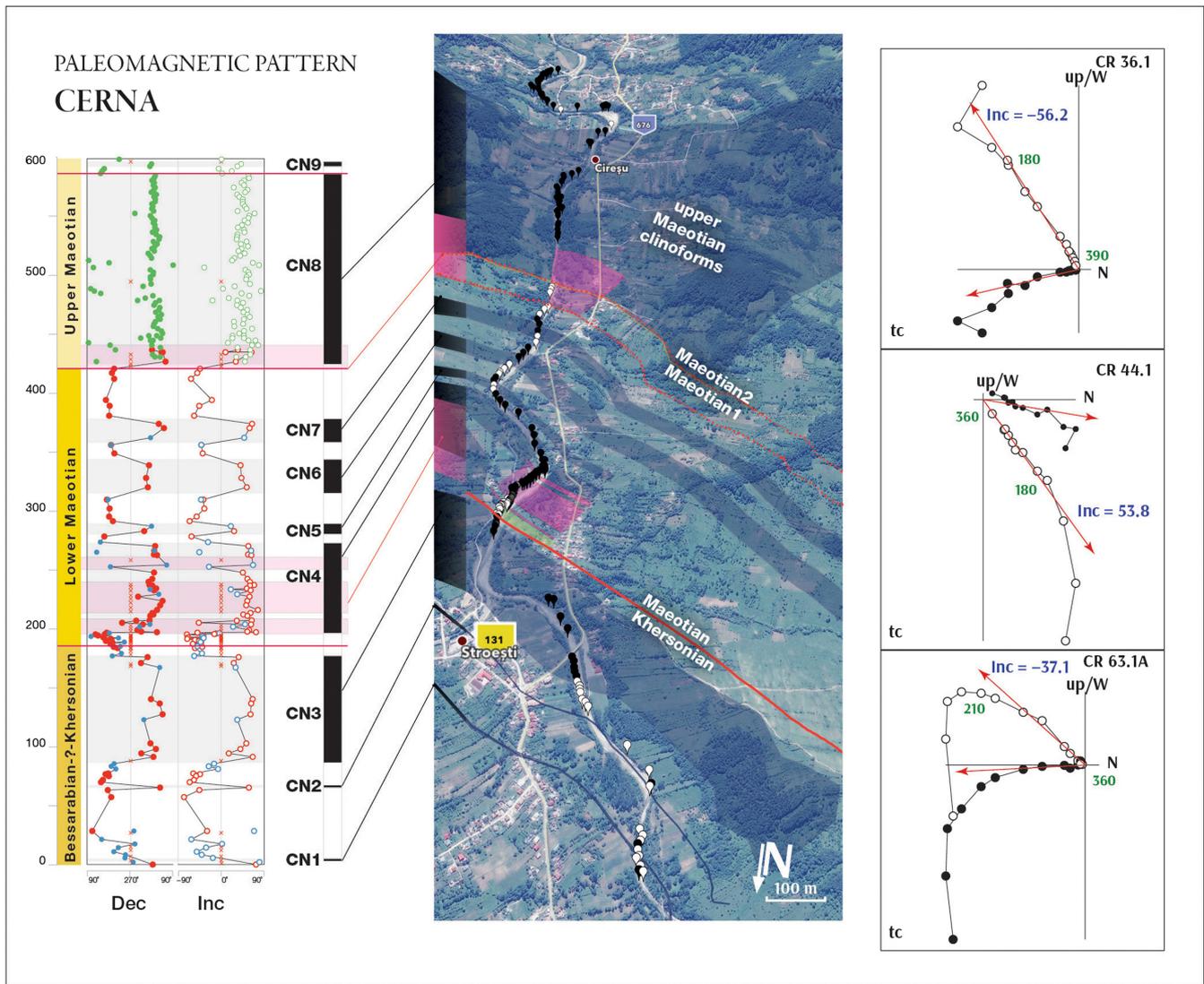


**FIGURE 8** Demagnetization results for the Rușavățu section. The magnetic declination (dec), inclination (inc), and the polarity pattern are plotted in stratigraphic view and map view; the red interval corresponds to the Intra-Maeotian Event when disturbances in the sedimentation affected the declination and inclinations (left and centre). Representative thermal demagnetization diagrams (after tilt correction), solid (open) circles denote projection on the horizontal (vertical) plane and the attached numbers in green indicate temperatures in °C. The samples code in indicated black, bold capital letters and the inclination angle is indicated in blue letters

with ages of 8.85 Ma for the base Maeotian (top of chron C4An) and 8.1 Ma for the IME (base of C4n.2n).

The polarity pattern of the Rușavățu section can be correlated straightforwardly to the GPTS. Starting from the Maeotian-Pontian boundary in RN7, that is basin-wide located in chron C3An.1n (Chang et al., 2014; Krijgsman et al., 2010; Stoica et al., 2013; Vasiliev et al., 2004, 2011), all polarity intervals can unambiguously be correlated downward in the GPTS, with the lowermost normal polarity interval RN1 to C4n.2n (Figure 11). This correlation indicates an age of 7.5 Ma for the Khersonian/Maeotian boundary (1<sup>st</sup> Maeotian transgression), 7.25 Ma for the 2<sup>nd</sup> Maeotian transgression, and 6.9 for the lower/upper Maeotian transition and the corresponding IME (Figure 11).

The polarity pattern of the Khersonian and lower Maeotian in the Cernișoara section can also be unambiguously correlated with the GPTS with XN1 to C4n.2n and XN 5 to C3Bn (Figure 11). The upper Maeotian only comprises normal polarities, which probably correlate to C3An, but the presence of slumped intervals and sandy lithologies suggest that no robust magnetostratigraphic correlation can be made here. The polarity pattern of the Cerna section is in good agreement with Cernișoara for the interval CN4–CN7, and only the Khersonian normal interval CN3 is significantly shorter. The correlation of the Cerna-Cernișoara composite section to the GPTS provides an age of 7.65 for the Khersonian/Maeotian transition (here marked by a change in lithology) and of 6.7 Ma for the IME



**FIGURE 9** Demagnetization results for the Cerna section. The magnetic declination (dec), inclination (inc) and the polarity pattern are plotted in stratigraphic view and map view; the red interval corresponds to the Intra-Maeotian Event when disturbances in the sedimentation affected the declination and inclinations (left and centre). Representative thermal demagnetization diagrams (after tilt correction), solid (open) circles denote projection on the horizontal (vertical) plane and the attached numbers in green indicate temperatures in °C. The samples code in indicated black, bold capital letters and the inclination angle is indicated in blue letters

(Figure 11). The lowermost part of the upper Maeotian, however, contains numerous examples of slumping and mass-transport, and it is possible that some time is missing in both sections at the IME interval.

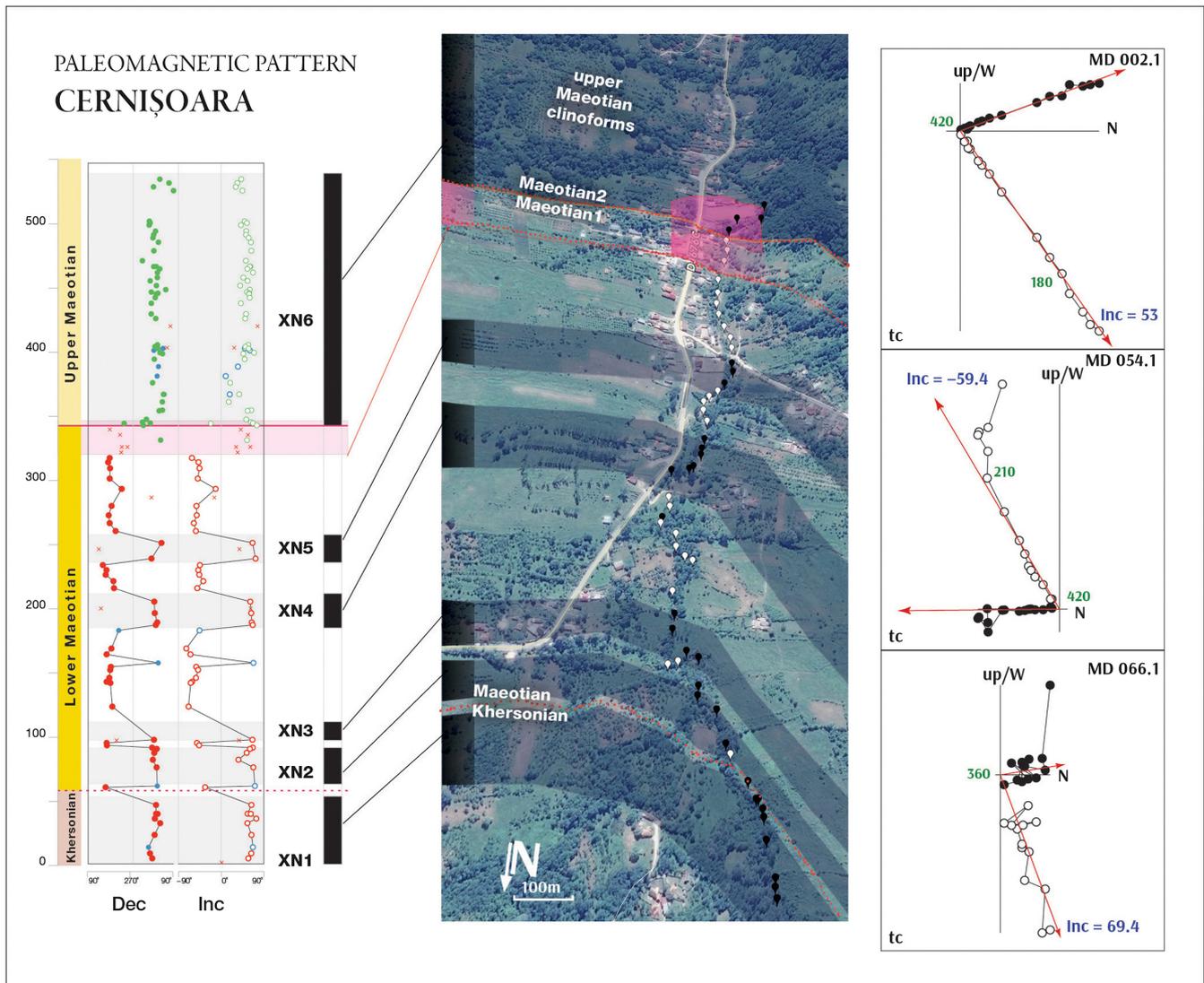
## 6 | TOWARDS A REVISED LATE MIOCENE AGE MODEL FOR THE EASTERN PARATETHYS

A significant effort has previously been put in obtaining palaeomagnetic data from late Miocene sedimentary successions from the Eastern Paratethys (Figure 11f,g) (Filipova & Trubikhin, 2009; Pevzner & Chikovani, 1978;

Pilipenko & Trubikhin, 2014; Trubikhin & Pilipenko, 2011; Vasiliev et al., 2004, 2005, 2011). The lithological successions of these sections, however, are in many cases incomplete, fragmented by several unconformities, which makes the correlations with the GPTS less straightforward (Figure 11).

### 6.1 | Dacian Basin

The age of 7.65–7.5 Ma obtained for the Khersonian/Maeotian transition in the Rușavățu section can be used to clarify the magnetostratigraphic correlation problems of the previously studied Putna and Râmnicu Sărat sections in the Focsani Depression of the East Carpathians foredeep



**FIGURE 10** Demagnetization results for the Cernișoara section. The magnetic declination (dec), inclination (inc) and the polarity pattern are plotted in stratigraphic view and map view; the red interval corresponds to the Intra-Maeotian Event when disturbances in the sedimentation affected the declination and inclinations (left and centre). Representative thermal demagnetization diagrams (after tilt correction), solid (open) circles denote projection on the horizontal (vertical) plane and the attached numbers in green indicate temperatures in °C. The samples code in indicated black, bold capital letters and the inclination angle is indicated in blue letters

(Vasiliev et al., 2004). Because of uncertainties in correlation of the acquired magnetostratigraphic record, Vasiliev et al. (2004) did not want to appraise an age for the Khersonian/Maeotian boundary. Their correlation figure suggests three options for this boundary, which is not based on biostratigraphic constraints but directly taken from the detailed geological maps of the region. The magnetostratigraphic correlation of the polarity pattern of the Râmnicu Sărat section resulted in an age of ~7.5 Ma, although the polarity pattern of the lowermost Maeotian is not very well documented. This correlation is in very good agreement with our Rușavățu data. The magnetostratigraphic correlation of the Putna section, however, provided an old ~9 Ma option and a young ~8 Ma option, both significantly older

than the Rușavățu results. The Putna section comprises relatively proximal alluvial sediments from the Balta delta system (Jipa & Olariu, 2009; Matoshko, Matoshko, Leeuw, & Stoica, 2016), which are very poor in fossil assemblages. The 1:20000 geologic maps of the area located the Khersonian/Maeotian boundary at the last level with Khersonian mollusks (the *Maetra bulgarica* fauna), just below a thick interval of reddish to purple-green continental beds (Dumitrescu, 1952) of normal polarity completely lacking age diagnostic fossils. The first Maeotian fossils, however, appear much higher in the section, above the long normal polarity interval. If we accept this normal polarity interval to correspond to C4n.2n (option 1 of Vasiliev et al. (2004)), the position of the lowermost Maeotian fossils

would not be in disagreement with the Ruşavăţu results. The Putna section also suffers from a nonexposed interval straddling the 7.7–7.3 Ma interval, which is represented by a change in bedding orientation (Vasiliev et al., 2004). Above this stratigraphic gap, the correlation of the upper Maeotian polarity pattern to the GPTS is robust. The upper two normal chrons of the section, corresponding to the upper Maeotian, are correlated with chrons C3An.1n and C3An.2n, in excellent agreement with our Ruşavăţu correlation (Figure 11).

The Bădăslava section (Figure 11a), located in the South Carpathians region, consists of Maeotian and Pontian deposits (Floroiu et al., 2011; Stoica, Lazar, Vasiliev, & Krijgsman, 2007; Vasiliev et al., 2005, 2007). The Maeotian sediments are found transgressively overlying Burdigalian sediments and are in turn overlain by upper Pontian (Bosphorian substage) deposits, indicating that at least the lower and middle Pontian are missing (Floroiu et al., 2011). The Maeotian part of the section is difficult to interpret biostratigraphically as it contains no marker fossils

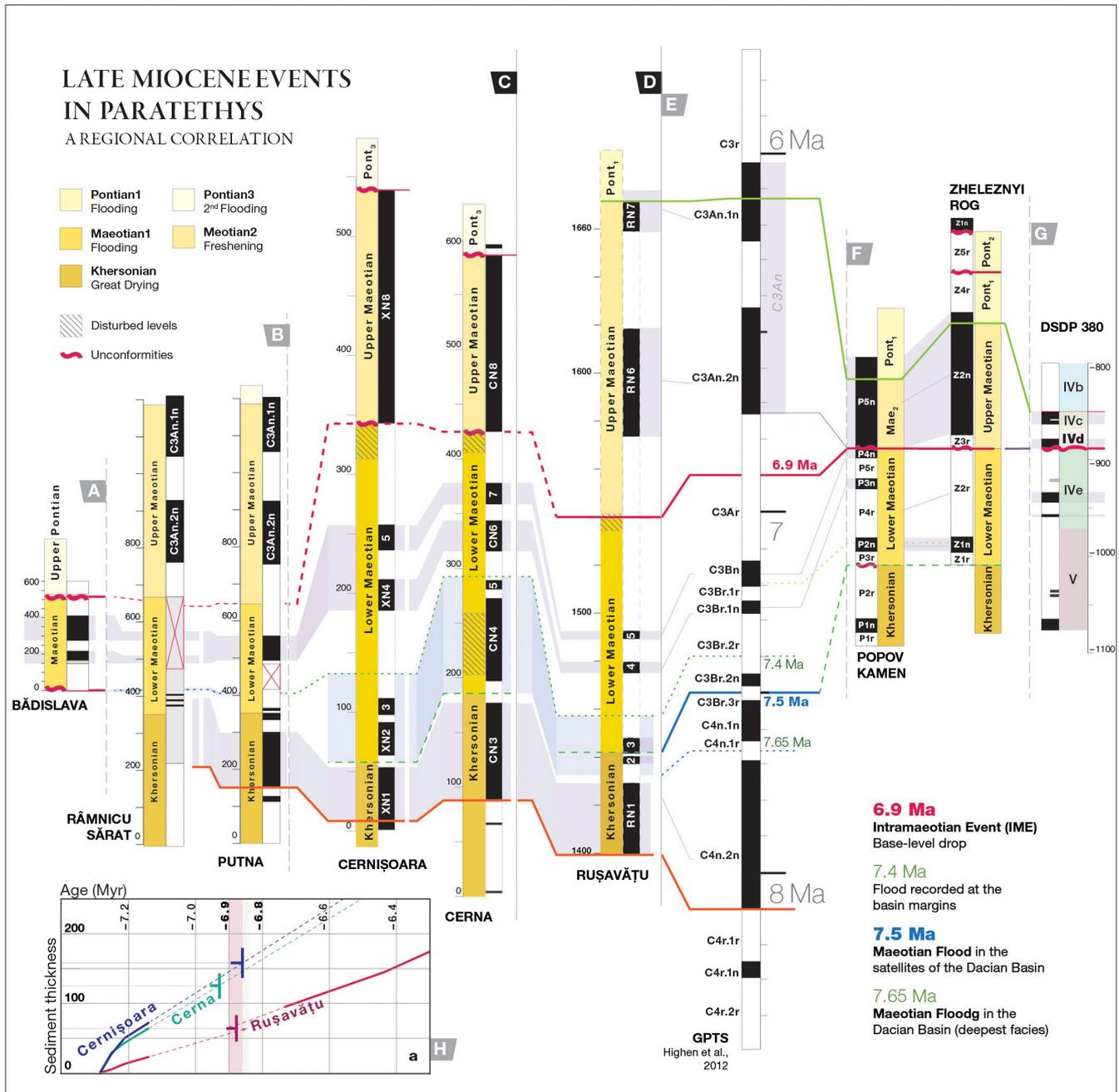


FIGURE 11 Correlations and age calculations between sections from various locations from Dacian Basin (A-Marginal basins, B-East Carpathian Foredeep, C-Getic Basin, D-piggy back basins), Kuban Basin (F), Black Sea Basin (G) and GPTS 2012 (E)

(Stoica et al., 2007). Magnetostratigraphically, the two normal polarity intervals correlate best to chrons C3Bn and C3Br.1n (Figure 11) indicating that this part of the section corresponds to the lower Maeotian (e.g., Stoica et al., 2007). This interpretation implies that the Maeotian sediments of Bădislava corresponds to the maximum Maeotian transgression between ~7.4 and 6.9 Ma.

## 6.2 | Euxinic Basin

The Popov Kamen and Zheleznyi Rog sections on the Taman Peninsula of the northern Black Sea Basin both comprise Khersonian, Maeotian, and Pontian sediments. They have been subject of multiple detailed magneto-biostratigraphic and radiometric dating studies (Krijgsman et al., 2010; Pevzner & Vangengeim, 1993; Pevzner, Lungu, Vangengeim, & Basilyan, 1987; Stoica, Krijgsman, Fortuin, & Gliozzi, 2016; Trubikhin, 1989; Trubikhin & Piliipenko, 2011; Vasiliev et al., 2011) but the age of the Khersonian/Maeotian boundary and the chronology of events during the Maeotian remain unclear. The base of the lower Maeotian is marked in both sections by a short-reversed polarity interval, followed by a short normal polarity interval (P3r/P2n and Z1r/Z1n) and a long interval of reversed polarity (P4r/Z2r). The upper reversed interval in Popov Kamen contains an additional short normal polarity interval that is apparently missing in Zheleznyi Rog (Figure 11).

A significant unconformity is observed between the lower and upper Maeotian (equivalent to the Intra-Maeotian Event), represented by clay breccia units (Popov, Golovina, Jafarzadeh, & Goncharova, 2015). Remarkably, the upper Maeotian pattern of both Popov Kamen and Zheleznyi Rog is very similar to those of Cerna and Cernișoara, being dominated almost entirely by a long normal polarity interval (P5n and Z2n in Figure 11). The lower/upper Maeotian transition closely corresponds in both sections to a reversed to normal polarity transition interpreted as the base of C3An (6.7 Ma; Trubikhin & Piliipenko, 2011; Rostovtseva and Rybkina, 2014). Correlating downwards result in two different age options for the Khersonian/Maeotian boundary: the polarity pattern of Zheleznyi Rog suggests a location in C3Br.1r at an age of 7.25 Ma; the pattern in Popov Kamen best fits to C3Br.3r at an age of 7.5 Ma (assuming that the short subchron C3Br.1n has been missed in the section; Filippova & Trubikhin, 2009). A cyclostratigraphic study, based on magnetic susceptibility patterns in the Popov Kamen section, revealed a 41,000-year obliquity cycle and suggested an age of ~7.6 Ma for the Khersonian/Maeotian boundary (Rybkina, Kern, & Rostovtseva, 2015).

In conclusion, we see no argument to place the Khersonian/Maeotian boundary below C4n.2n (>8 Ma) as proposed by Semenenko et al. (2009) and Vasiliev et al.

(2011) who favour an age of 9.8 Ma and 8.2 Ma, respectively. Vasiliev et al. (2011) based their age on radiometric dating of a volcanic ash layer in the uppermost part of the Khersonian in Zheleznyi Rog, although they admit that their palaeomagnetic pattern is difficult to correlate. According to Popov et al. (2015), however, the Khersonian/Maeotian boundary in Zheleznyi Rog represents a hiatus. We revisited the boundary interval in Zheleznyi Rog again to critically evaluate the previous interpretations and conclude that there are indeed signs of tectonic disturbance by potentially major faults in that part of the section. The radiometric age of  $8.69 \pm 0.18$  by Vasiliev et al. (2011) can thus not be interpolated to the Khersonian/Maeotian boundary, which makes the age of 8.2 Ma obsolete.

Unfortunately, not much age information is available from the late Miocene sediments of the deep Black Sea Basin. This is mostly related to limited exposures and limited access to industrial drill core data. The DSDP Site 380 is one of the most important sources of information but initially the presence of greigite ( $\text{Fe}_3\text{S}_4$ ) severely complicated establishing a robust polarity zonation (Giovanoli, 1979). Only recently it became clear that greigite could be a reliable carrier of magnetization (Vasiliev et al., 2007, 2008) and a greigite-based magnetostratigraphic time frame was established for these old DSDP cores (Van Baak et al., 2016). Among the three studied drill cores, the magnetic polarity pattern of Hole 380B covers the Khersonian-Maeotian time interval best (Figure 11f). In this core, unit IVb has been correlated with the Pontian, while units IVc and IVd, containing a normal-reversed-normal polarity triad, are correlated with chron C3A (Van Baak et al., 2016). Unit IVe, containing dominantly reversed polarities and only two short normal polarity intervals, has been correlated to the C3Ar-C3Br.2r interval while unit V is too poorly constrained for a clear correlation. This pattern fits very well with the results from the other sections and suggests that the base of unit IVd, the characteristic “pebbly breccia” unit, has an age of ~6.7 Ma. Unit IVd was first initially considered an expression of Black Sea desiccation during the Messinian Salinity crisis (Hsü & Giovanoli, 1979). Later studies, however, revealed it belonged to an older erosional event (Grothe et al., 2014) that according to our correlations closely corresponds to the IME.

Recently, Tari et al. (2015) have shown that the pebbly breccia unit is belonging to a mass transport body, linked with a phase of slope instability. In addition to this, the study of Sipahioğlu and Bati (2017) on Miocene canyons in the Black Sea show that their Mass Transport Complexes (MTC) of assumed Messinian age are filling older erosional features (e.g., the Karaburun Canyon). This suggests that the Miocene erosional features in the Black Sea belong to an older event than the MTC, which potentially can be linked with the Intra-Maeotian Event. The upper

Miocene interval in the DSDP record does not show any indications for long normal polarity intervals and can thus be interpreted as younger than 7.6 Ma (Figure 11).

### 6.3 | Continental records

Several magnetostratigraphic studies have been conducted on late Miocene mammalian localities surrounding the Paratethys realm. The ages attributed to the Khersonian/Maeotian boundary in these terrestrial records strongly depended on previous (widely varying) age estimates of the overlying Pontian deposits and consequently arrive at a broad range of ~9 Ma, Pevzner and Vangengeim (1993); 9.4 Ma, Vangengeim, Lungu, and Tesakov (2006) and a radio-isotopic age of 9.3 Ma; 9.6 Ma, Vangengeim and Tesakov (2008); 6.4 Ma, Tesakov, Titov, Syromyatnikova, Danilov, and Frolov (2013); 7.8/8.2 Ma, Tesakov et al. (2017). The magnetic polarity patterns of the continental sections do, however, contain some recurrent patterns: the Khersonian is commonly dominated by a long normal polarity interval (sections Poksheshty, El'Dari, Starya Kubanka, Yurievka, Tiaginka, Krivoj Rog, Tiaginka and Kajnary in Pevzner and Vangengeim (1993); El'Dari section, in Vangengeim et al. (2006)). This long normal polarity interval was originally correlated with C5n and an age between 11 and 10 Ma (Vangengeim et al., 2006). In light of a much younger age for the base of the Pontian (6.1 Ma instead of 7.5 Ma) we re-evaluate this normal polarity pattern of the Khersonian and correlate it with chron C4n. This would bring it in agreement with the recent correlation of the Khersonian/Maeotian boundary in the north Caucasus region (Tuapse highway bridge, Gaverdovsky and Volchaya Balka in the Blinov and Gaverdovsky formations near the city of Maikop) to the base-C3Br/top-C4n interval at an age of ~7.5 Ma (Tesakov et al., 2017).

## 7 | DISCUSSION

### 7.1 | The end of the Great Drying of Eurasia; the Maeotian transgression

The Late Miocene Tortonian stage (11.6–7.24 Ma; Hilgen et al., 2005) is characterized by significant climatic changes (Herbert et al., 2016) and a major palaeogeographic reorganization of the Mediterranean region (e.g., Jolivet, Augier, Robin, Suc, & Rouchy, 2006) including its Atlantic gateways (e.g., Flecker et al., 2015). In this period, Paratethys was a long-lived lake without any significant connection to the global ocean (Popov et al., 2010). Paratethys' water level thus contracted and expanded only in response to regional climate. Paratethys stretches especially in the latitude that straddles the present-day boundary between a negative and positive hydrological balance. Small changes

in climatic conditions can thus have major impact on regional palaeoenvironments, explaining shifts from moderately dry to moderately humid conditions (Eronen et al., 2012). In dry periods, lake-level drop leads to the reduction or loss of intra-basinal connectivity (Figure 2a) and the formation of seas with obstructive environmental conditions. In the relatively wet phases, when lake level rises enhanced communication and mixing between basins (Figure 2b), uniform environments such as Caspian-type brackish lake-seas will thrive.

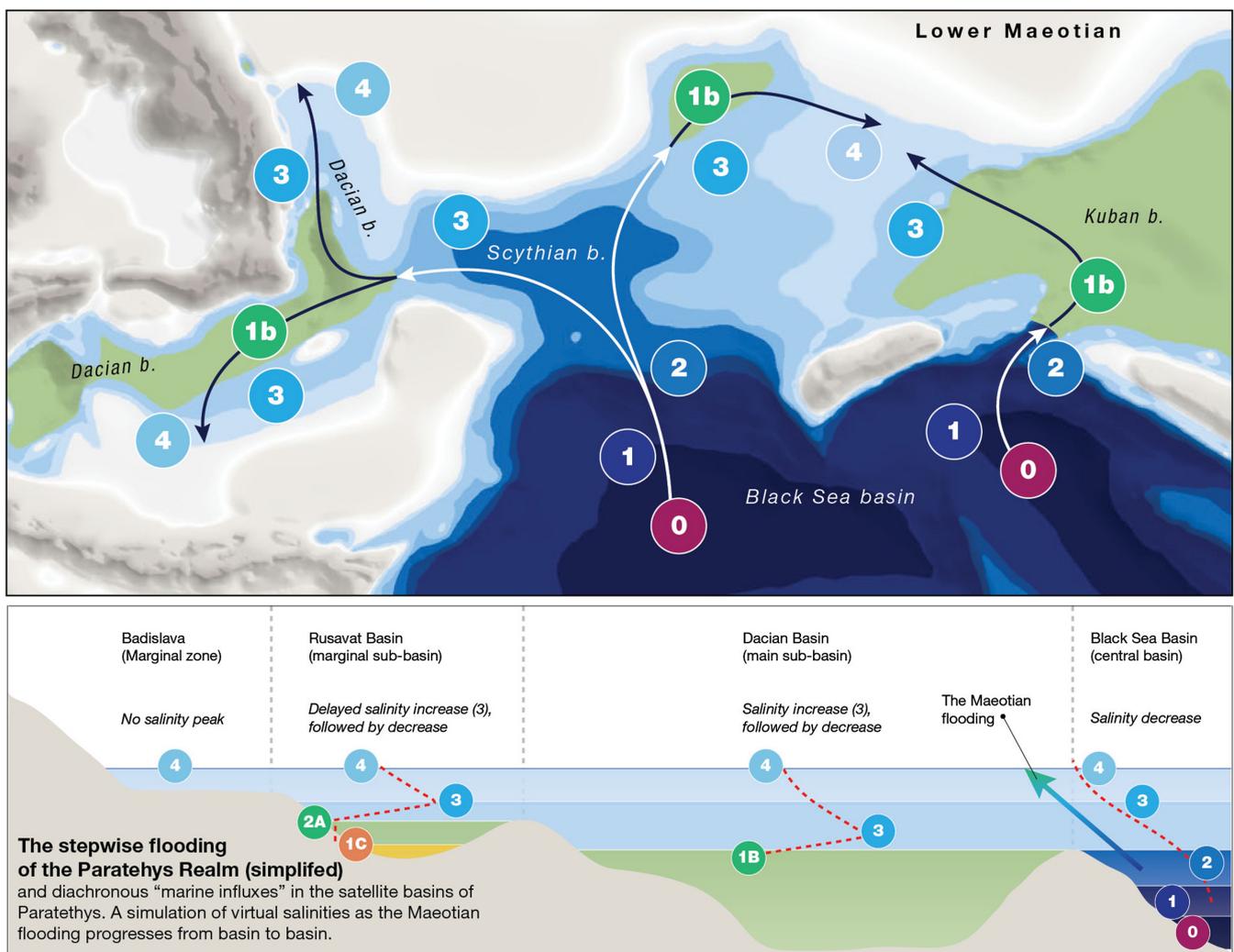
The Khersonian stage of the Eastern Paratethys is generally considered as the time period of driest conditions in Central Eurasia (Stekina, 1979; Syabryaj, Utescher, Molchanoff, & Bruch, 2007). Palaeoclimatic data indicate that the fluctuating Khersonian climate contained episodes that were relatively cool (particularly in winter) and significantly drier than the Bessarabian (Ivanov et al., 2011). These episodes led to an aridization trend, reflected in the partial replacement of forest vegetation by herbaceous communities (Ivanov et al., 2011). The Black Sea water level experienced its major lowstand of ~300 m during the Khersonian, and the marginal regions, including the Dacian basin, partially desiccated (Jipa, 2009; Jones & Simmons, 1997; Tugolesov et al., 1985). The previous saline waters of the marginal Bessarabian basins migrated towards the central Black Sea Basin, which consequently became significantly enriched in salt. Unconformities between the brackish or hypersaline Khersonian and the semi-marine lower Maeotian deposits are very common, suggesting drying up of large marginal areas (Popov et al., 2006). In the Dacian basin, the Khersonian is generally marked by continental conditions, represented by reddish soils in the Rușavățu, Putna, and Milcov sections (Ioniță, 1963; Korotkevich, 1970; Vasiliev et al., 2004) and seriously reduced sediment accumulation rates. According to the isopach map of Saulea et al. (1969), the maximum thickness of the Bessarabian and Khersonian sediments in the Dacian basin is 400 m in two key distribution areas: the Focșani Depression in the Carpathian bend zone and the Getic Depression in the west. The thickness of Khersonian sediments decreases to zero towards the southern and eastern limit of the basin (Saulea et al., 1969). Regarding the long duration (~4 Myr) for the combined Bessarabian–Khersonian interval, the maximum sedimentation rates of 10 cm/kyr are much lower than for the Maeotian (60 cm/kyr) and Pontian (155 cm/kyr) (Vasiliev et al., 2004).

Our new magneto-biostratigraphic results indicate that the dry Khersonian environments mainly correspond to the normal polarity intervals of chron C4n, and that they are thus formed in the time interval between 8.6 and 7.5 Ma (Figure 11). This allows a new and better calibration of all available palaeoenvironmental proxy data for this time period, resulting in recalibrated ages that can be up to 1–3 Ma

year younger than previously estimated, depending on which time scale was used. The end of the dry Khersonian conditions are linked to the first Maeotian transgression which took place at an age of  $\sim 7.5$  Ma, possibly slightly diachronous over the Eastern Paratethys basins. The maximum of the transgression took place in the second part of the early Maeotian (Neveskaya et al., 1986; Popov et al., 2004; Stevanovic & Ilyina, 1982).

The magnetostratigraphic records of Ruşavăţu -and Cerna-Cerņișoara indicate that the Maeotian transgression may also be slightly diachronous within the Dacian basin, as the main lithological transition in Cerna is  $\sim 200$  kyr older than the first saline ingression in Ruşavăţu. The Maeotian transgression occurs in a time of relatively low subsidence and is therefore most likely climatically induced. Our biostratigraphic records (Figure 5) show a loss in salinity as the transgression progresses, in

agreement with results from the Black Sea region (Popov et al., 2016). This is opposite to the expected effect of a marine transgression by Mediterranean waters through the establishment of a reconnection with the Mediterranean. We therefore consider it more likely that the Maeotian transgression is caused by intra-Paratethyan hydrological changes, progressively raising the level of the Black Sea. The Maeotian transgression is accompanied by a small change in salinity, testifying that the Khersonian Black Sea was filled with relatively saltier water, as previously suggested by Kojumdgieva (1983). The stepwise flooding of the Paratethys basins can potentially be traced from the deeper basins to the marginal regions according to their bathymetry and palaeogeographic position (Figure 12). The marginal lake relicts of Paratethys that was fragmented during the Khersonian then progressively all reconnected in the Maeotian highstand.



**FIGURE 12** The stepwise flooding of the Dacian Basin during the Maeotian transgression. The central basin expands (0, 1, 2) and reconnects with the previous lacustrine sub-basins (1B) and the marginal subbasins (1C) and further expands to (3, 4). As the basin continues to expand, the overall salinity (marked with dotted red lines) drops in the central basin and shows positive pulses in the sub-basins and the marginal basins

The biostratigraphic record of Ruşavăţu shows several waves of “marine influxes” that consist in levels foraminifera. These influxes are resulting from highstand episodes when water and fauna from the saltier Black Sea basin cross the sills and spill into the marginal basins. In the case of the Dacian Basin (Figure 12) the initial freshwater systems of the Khersonian lakes (1B) will be flooded by- or will mix with- saltier water from the Black Sea (1). The water mixing then decreases salinity (3) and as the next basins will be flooded salinity will continue to drop (4). In the case of the Ruşavăţu basin, that shows a transition from continental and freshwater systems (1), (2), the salinity spike will be delayed with respect to Cerna-Cernisoara (3) and will also be followed by a drop in salinity (4).

We conclude that, although climate significantly fluctuated during the Khersonian and Meotian (Böhme et al., 2017; Vasiliev et al., 2015), the end of the dry Khersonian phase in the Eastern Paratethys corresponds to a change in the hydrological balance of the region. The shift to an overall positive hydrological balance caused a progressive increase in water level and subsequently resulted in flooding of the marginal basins during the earliest Maeotian.

## 7.2 | The Intra-Maeotian Event: reconnecting the Mediterranean?

The end of the transgressive trend in the Maeotian corresponds to a phase of slope instability that frequently characterizes the lower–upper Maeotian boundary. This Intra-Maeotian Event (IME), also locally referred as the Intra-Maeotian unconformity (e.g., Dacian Basin, Munteanu, Matenco, Dinu, & Cloetingh, 2012), can be observed both in the seismics (Tărăpoancă et al., 2004) and in outcrops throughout Paratethys. The most visible expressions of the IME are slumps and mass transports near the high slope margins of the basin (e.g., Cerna and Cernișoara; DSDP) or major unconformities in marginal successions (e.g., Zheleznyi Rog). The Intra-Maeotian Event corresponds to the upper part of chron C3Ar and is dated between 6.9 and 6.7 Ma (Figure 11h). A logical explanation for the IME is a significant sea-level drop, which can explain a basin-wide destabilization of sediments in the high slope parts of the basin.

The presence of Aegean species in the upper Maeotian (Radionova & Golovina, 2012) suggests that the IME corresponds to the creation of a connection to the Mediterranean through the proto-Bosporus—Aegean. If the Paratethys water level had been higher than the global sea level, this would have caused a massive flow of waters to the Mediterranean and a sudden lowering of the Paratethys level. After the IME, the marginal Dacian Basin reveals oligo-haline to fresh water conditions, probably indicating restrictions in mixing, while the Black Sea remains brackish with some freshwater episodes. Later in the upper

Maeotian, more elements of the Mediterranean fauna appear in the Black Sea as the prelude of the Pontian flooding at 6.1 Ma (Radionova & Golovina, 2012).

## 8 | CONCLUSIONS

We provide integrated magneto-biostratigraphic results for three continuous, 250–600 m long, sedimentary successions, situated in the Dacian basin Depression in a relatively deep (Cerna and Cernișoara Rivers) and a shallower setting (Ruşavăţu River), in present day Romania. The successions comprise the uppermost Khersonian (upper Sarmatian s.l.) – Maeotian – lower Pontian interval of the Eastern Paratethys. The magnetostratigraphic record from the Cerna section consists of 17 polarity intervals, eight of reversed, and nine of normal polarity. The polarity pattern of the Cernișoara section comprises 13 different polarity intervals, 6 of reverse, and 7 of normal polarity. The polarity pattern of the Ruşavăţu section comprises 13 polarity intervals, 6 of reversed, and 7 of normal polarity. The three sections have a common pattern that can be correlated with the GPTS. Our correlation shows that the entire succession covers the time interval from ~8 to ~6 Ma. We date and describe the major palaeoenvironmental phases, which can all be related to changes in the climate and connectivity regime of Eastern Paratethys with the Mediterranean.

We show that the great drying episode of the late Khersonian ended because of a shift in the water budget in the region of Paratethys at 7.65 Ma leading to a major transgression (flooding) that went on for 250 kyrs with ups and downs, probably due to climate change. As the stepwise flood progressed, between 7.65 and 7.4 Ma, the marginal basins were flooded and increased connectivity with the rest of the lake. We date the Intra-Maeotian Event at 6.9 Ma, corresponding to sedimentary anomalies such as the mass transport units in the west Dacian Basin and the Pebbly breccia in the Black Sea due to a sudden lake level fall. This event corresponds to the boundary between lower and upper Maeotian that we suggest it marks the reconnection between Paratethys and the Mediterranean.

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