Towards an astrochronological framework for the eastern Paratethys Mio–Pliocene sedimentary sequences of the Focşani basin (Romania)

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Abstract

A reliable geological timescale for the Neogene sedimentary deposits of the Romanian Carpathian foredeep did not exist so far due to the lack of accurate isotopic datings and straightforward magnetostratigraphic correlations. Here we present a new chronology for the Upper Miocene to Pliocene deposits of the westernmost part of the eastern Paratethys based on high-resolution magnetostratigraphic data from the Focşani basin. The thick and continuous river sections of Putna (2.35 km) and Rîmnicu Sârat (7.32 km) yield excellent magnetostratigraphic results and allow an unambiguous correlation to the polarity timescale for the interval between 7 and 2.5 Ma. The new age control permits the calculation of accumulation rates for each polarity zone, which shows a significant increase at approximately 5.8–6 My. The Upper Miocene to Pliocene sedimentary rocks of the Focşani basin display cyclic alternations of sandy units and finer intervals. The calculated values for the average duration of the observed sedimentary cycles are very close to the average duration of precession (21.7 ky), which indicates that the sedimentary cycles from the eastern Carpathian foredeep deposits are astronomically forced. We show that the usage of the unambiguous time-control offered by magnetic polarity stratigraphy would give a more realistic picture of the processes that have taken place in this tectonically active region.

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1. Introduction

During the Mesozoic and early Cenozoic, the Eurasian and African continents were separated by a large oceanic basin called the Tethys. Global plate tectonic processes caused a northward motion of the African plate with respect to Eurasia and led to continental collision and the gradual closure of the Tethys Ocean. These tectonic movements generated the elevation of the Alpine–Himalayan mountain belt, which started to act as an E–W striking barrier since the beginning of the Oligocene. Consequently, the Tethys Ocean evolved into two different domains, the Mediterranean basin to the south and the Paratethys to the north (Fig. 1). The Mediterranean remained an open marine basin because the connection to the Atlantic [1] guaranteed the exchange of water masses and organisms, permitting a direct biostratigraphic correlation to the world’s oceanic record. The Paratethys became semi-isolated with brackish to fresh water environments that led to the development of endemic faunas and different biozonations. In addition, ongoing tectonics caused a fragmentation of the Paratethys into various subbasins, which were affected by a continuous changing of the water surface extent and of the water connections between subbasins (Fig. 1) and the Mediterranean and Indian Ocean.

The Paratethys region is divided into a western and eastern Paratethys, which are separated from each other by the Carpathian mountain range [2,3]. The western Paratethys formed in the back (to the north and west) of the Alpine–Carpathian belt and comprises the Pannonian and Transylvanian basins (Austria, Hungary and NW Romania). The eastern Paratethys developed in the Alpine–Carpathian foredeep and consists of the Dacian, Euxinian and Caspian basins (SE Romania, Moldavia, Ukraine, SW Russia). The Black Sea and the Caspian Sea are the actual brackish water remains of this ancient water mass.

Chronostratigraphic control for the fresh to brackish water deposits of the Paratethys domain is generally poor and largely based on biostratigraphic data from endemic molluscs and ostracods. The complex paleogeographic configuration of the region resulted in the development of different biozonations and different chronostratigraphic stages for the various subbasins [2,4–8] (Fig. 2). Correlations of the Paratethys stages to the marine biochronologies of the Mediterranean and global oceans are generally impossible because of the semi-isolated position and the lack of water exchange. In addition, the political situation before 1990 did not facilitate communication between scientists of the various Paratethys countries.

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**Fig. 1.** Schematic paleogeographic map of the late Miocene, showing the Paratethys area (darker gray), the Mediterranean (lighter gray) and the present-day land contribution. The gateways (dashed) are drawn according to Krijgsman [1].
Radiometric datings of the eastern Paratethys sequences are very rare [9] and magnetostratigraphic studies are mainly applied to small, scattered outcrops, generally focusing on the youngest Plio–Pleistocene deposits [4,5,10–12]. Consequently, the Neogene chronology of the eastern Paratethys is poorly defined. Ages of stages can easily vary by more than 1 or 2 million years between different basins and different studies (Fig. 2). A good example is the Pontian Stage for which estimates of the duration vary from 3.0 My [5] to 0.3 My [7]. Evidently, such uncertainties and differences in age dating will have serious consequences for studies that use these ages to establish the timing of different tectonic and climatic events or to calculate accumulation and subsidence rates (e.g. [13–15]).

The objective of this work is to establish a detailed geochronological framework for the Upper Miocene–Pliocene sedimentary deposits of the Romanian segment of the Carpathian foredeep (Fig. 3). From a paleogeographical viewpoint, the study area is known as the Dacian basin (Fig. 1) and represents the westernmost part of the eastern Paratethys (Fig. 3a).

We have selected the sedimentary sequences of the Focșani basin for a magnetostratigraphic study, because the Mio–Pliocene units of this basin are very thick (>7 km) and continuously exposed along several river valleys. In addition, the Upper Neogene deposits of the east Carpathian foredeep express a distinct sedimentary cyclicity of alternating siltstones/clays and sands. The regularity of the sandy intercalations strongly suggests a relation with regional climate changes induced by the orbital variations of the Earth. Consequently, these deposits could be very suitable to establish an astronomical polarity timescale (APTS) for the sedimentary sequence of the eastern Paratethys, allowing detailed correlations to the Mediterranean marine and continental record, for which such a timescale has already been developed [16].

2. Geological setting and sections

The Carpathians consist of two radially directed fold and thrust belts: (1) an inner belt formed by Cretaceous to early Paleogene thrusting, and (2) an
outer belt that is the result of Neogene thrusting [17].

The age of the last nappe emplacement in the Carpathians is considered to be intra-Sarmatian [18,19]. The Focșani area, located in front of the bended part of the southeast Carpathians (Fig. 3a), represents the main sedimentary basin of the Carpathian foredeep and contains more than 10 km [15] of Upper Miocene and Quaternary deposits. It is superimposed on the most rapidly subsiding and seismically active area in Romania, which shows earthquakes up to 7.2 in magnitude [17,20]. Late Miocene to Quaternary basin evolution studies demonstrate that large-scale subsidence occurred in front of the SE Carpathians, culminating in significant differential vertical motions along and across the arc during the later stages [21]. Active subduction in the Carpathian region ceased approximately 9 My ago [22], but the final phase of subduction and detachment of oceanic lithosphere still produces seismicity and subsidence.

The Miocene to Pliocene sedimentary sequences of the Focșani basin consist of cyclic alternations of
well-cemented coarse-grained rocks (sandstones and microconglomerates) and fine-grained rocks (siltstones and shales), which have been deposited in a lacustrine to deltaic environment. The cyclic pattern is constant for several kilometers, as can be observed in the various river incisions that cut through the tilted strata. A coarsening of the deposits and a decrease in faunal content take place from the south towards the northern part of the Focşani basin [23]. The sandy units are generally regularly spaced, mostly well cemented, and show sometimes small erosional surfaces at the base of channel and bar structures. Clays, marls or silts fill the space between these sandstones. For our paleomagnetic study, we selected the sections along the Putna and Rimnicu Sărat rivers that are located on the western flank of the Focşani basin, where the strata are tilted up to the vertical (Fig. 3c). The composite of these sections covers the time span from late Sarmatian to late Romanian (Upper Miocene to Pliocene).

A total number of 326 levels were sampled from the 2.35 km thick Putna section. The section starts stratigraphically above the main Sarmatian thrust contact in the first Upper Sarmatian deposits and it ends at the top of the Pontian (or base of Dacian) deposits. Throughout the entire Putna section, the layers have a subvertical bedding dip varying between 95° and 85°. The Upper Sarmatian part contains relatively coarse material, with strongly cemented sandstones, sometimes with a fine conglomeratic base. Silty units separate the sandstones and are usually laminated and friable. A reddish interval with paleosols (at approximately 175 m) indicates occasional subaerial exposure of the deposits. At approximately 480 m, there is a non-exposed gap in the section, which coincides with a change in strike of approximately 12°. There is not any known or visible fault in this non-exposed part of the section. The Meotian and Pontian part is less coarse and the sandy units are often less cemented.

From the 7.32 km thick Rimnicu Sărat section, 396 levels were sampled. The section starts in the Upper Sarmatian and ends in the Upper Romanian. The Sarmatian part is not very well exposed but the Meotian, Pontian, Dacian and Romanian have very good exposure. The sedimentary bedding shows a dip of approximately 90° in the lower part of the section, but goes progressively to a minimum of 20° in the upper part. The sedimentary deposits from Rimnicu Sărat are less coarse than in the Putna section, suggesting that they were deposited in a more distal part of the deltaic system, where the basin energy was lower.

Standard paleomagnetic cores were taken with an electrical drill and a generator as power supply. At each level, we took at least two standard-oriented cores. Additional samples have been collected for future biostratigraphic, sedimentological and geochemical purposes.

3. Paleomagnetic results

3.1. Methods

To establish a magnetostratigraphy for the Putna and Rimnicu Sărat sections, at least one specimen per sample layer was stepwise thermally demagnetized. The demagnetisation was performed with small temperature increments of 20–30 °C up to a maximum temperature of 620 °C, in a magnetically shielded, laboratory-built, furnace. The natural remanent magnetisation (NRM) was measured on a horizontal 2G Enterprises DC SQUID cryogenic magnetometer (noise level 3×10⁻¹² Am²). The directions of the NRM components were calculated by principal component analysis [24]. The large dip (mostly between 65° and 90°) of the layers was helpful to distinguish between primary and secondary components and to recognize a present-day overprint.

Furthermore, several rock magnetic experiments were performed to identify the carriers of the magnetisation. Thermomagnetic measurements were performed in air up to 700 °C for 36 powdered samples from diverse lithologies on a modified horizontal translation type Curie balance (noise level 5×10⁻⁹ A m²). The initial susceptibility and the anisotropy were measured on a Kappabridge KLY-2 or KLY-3. Hysteresis loops were measured for 35 samples of selected lithologies on an alternating gradient magnetometer (MicroMag Model, Princeton, noise level 2×10⁻⁸ Am²) to determine the saturation magnetisation (Ms), remanent saturation (Msr), coercive force (Bc) and remanent coercivity (Bcr). First-order reversal curves (FORC) [25,26] were measured for 33
samples to evaluate the magnetic domain situation, the
presence of magnetic interactions and the magnetic
mineralogy. For each FORC diagram, 150 curves
were measured with an averaging time of 0.2 s per
data point.

3.2. Thermal demagnetisation and rock magnetism

Thermal demagnetisation diagrams show different
types of magnetic behaviour that have a clear
relationship with lithology. Clays and silts generally
provide good quality diagrams. Sandstones commonly
provide poor quality demagnetisation diagrams (even
though magnetic intensity is generally high). Based on
rock magnetic results, we divided the samples into
two major groups.

Iron oxide group (1) type of magnetic behavior is
typical for the clays and most of the silts (Fig.
4a,c,e,f), but it may also occur in the sandstones. The
iron oxide group of samples characterizes the entire
Putna section. Only a part of the Rîmnicu Sărat Valley
samples record this behaviour and are typically found
in the oldest and coarser (Sarmatian) part of the
section.

Fig. 4. Thermal demagnetisation diagrams. Closed (open) symbols represent the projection of the vector end-points on the horizontal (vertical)
plane; values represent temperature in °C; stratigraphic levels are in the lower left-hand corner; lithologies are in the right-hand corner. The left-
hand side diagrams of figures a, b, c and d are represented without tectonic correction (th/no tc). The right-hand side diagrams of figures a, b, c
and d and figures e, f, g and g are represented with tectonic correction (th/tc). The direction of the present-day overprint (pdf) represented with
arrows in figure a.
The initial intensities of the NRM are in the range 10–300 mA/m and susceptibilities 0.14–20 × 10⁻³ SI. Most of the samples show no significant change in susceptibility during the heating treatment. Thermal demagnetisation diagrams illustrate that, in most cases, a large secondary present-day field component is removed at temperatures varying between 240 and 380 °C (Fig. 4a). Another proportion of the samples shows the removal of the present-day overprint after heating at only 100 °C. Taking into consideration the maximum temperature up to which the NRM was preserved, this group can be subdivided in two subgroups:

**Subgroup 1A (magnetite)** is characterized by samples of which susceptibility decreases continuously, without any visible increase upon heating to the highest temperatures (585 °C). The thermal demagnetisation diagrams of this subgroup are characterized by linear decay of the NRM to temperatures of 585 °C. The Curie balance measurements reveal that the dominant magnetic carrier for this group of samples is a mineral with a maximum unblocking temperature (Tb) in the range 540–580 °C, close to the Curie temperature (Tc) of magnetite (Fig. 5a). The lowering of the Tc can be induced by enrichment in titanium or some other type of impurities. The hysteresis curves, up to fields of 1.6 T, are closed well below 150 mT (Fig. 6a), and confirm the presence of a low coercivity mineral like magnetite. The FORC diagrams (Fig. 7a) furthermore infer an interacting multi-domain (MD) state of the magnetic minerals [25,26].

**Subgroup 1B (magnetite–hematite).** The hysteresis loops acquired for this subgroup are not closed in fields of 300 mT (they are saturated at approximately 800 mT) and indicate the presence of a high-coercivity mineral (Fig. 6b). The unblocking temperature (Tb) observed in thermomagnetic runs on the Curie balance is approximately 650 °C. The demagnetisation diagrams combined with the hysteresis loop interpretations (Fig. 6b) indicate that the high coercivity mineral with Tb around 650 °C is hematite. The magnetite–hematite type magnetisation is limited to the 28 m containing paleosol levels from the Sarmatian/Meotian part of the Putna section, where the red-orange color of the samples indeed indicates the existence of hematite in the sediment.

**Iron sulphide group (2)** characterizes the largest (and younger) part of the Rîmnici Sârat samples.
removed at temperatures of 100–180 °C (Fig. 4b,d,g,h). In about 20% of the samples, there is (virtually) no low temperature overprint component (Fig. 4h). Some of the samples record a low temperature secondary component of reversed polarity (Fig. 4b). Curie balance measurements of samples from group 2 show invariably an increase of total magnetisation after heating to 420 °C (or even after 290 °C) (Fig. 5b). This behavior may indicate the oxidation of an iron sulphide. The FORC diagrams show in the majority of these cases the strongly interacting single domain (SD) state with a $B_C$ varying between 45 and 90 mT (Fig. 7b). The hysteresis loops recorded for this group have a rectangular shape (Fig. 6c), which is characteristic for pyrrhotite [27]. Using the biplot of SIRM/$\chi$ versus remanent acquisition coercivity for magnetic minerals, we observed that all the analyzed samples from this group fit perfectly in the pyrrhotite area [28]. An identical conclusion was reached by plotting the results of the same samples on the biplot of $M_{RS}/M_S$ versus $(Bo)_C/(Bo)_{CR}$ [28].

3.3. Magnetostratigraphy

In the Sarmatian and the lower Meotian segment of the Putna section, the occurrence of overprinted samples is higher than in the rest of the section. This is probably caused by the dominance of coarser sandy–silty sequences in this interval and the relative lack of finer sediments capable of retaining a better magnetic signal. Nevertheless, sufficient samples with a good signal allow a convincing interpretation of the overall polarity in this part of the section. Typically, the occurrence of a present-day overprint is rare in the Rimnicu Sărat section and is only associated with yellowish, coarser material. In both sections, some coarser-grained rocks show viscous scattering of the NRM component. We include the present-day overprint and viscous NRM types in the category of unreliable directions. In some levels, in the vicinity of a polarity reversal, the antipodal magnetisation of two different components could be observed (Fig. 4b). We believe that in these cases early diagenetic processes cause a delayed acquisition and a (partial) overprint of the original (earlier acquired) component [29,30]. Hence, the direction of the following (younger) polarity interval will overprint the original direction of the studied level. For this reason, it was very important to separate the low and high temperature components.

The magnetisation for the iron sulphide group, most probably, was acquired during early diagenesis of the sediments [31]. The acquisition of the magnetisation for hematite is uncertain [29]. The contribution of the magnetite–hematite type of magnetisation to the total amount of samples is less than 2%. The hematite contribution to the total magnetisation is very small and the directions are the same as of the magnetite (the main magnetic mineral in that samples). Most likely, the occurrence of hematite in our sections...
suggests a magnetisation acquired during or soon after deposition.

Based on the results of rock magnetic analyses, the ChRM component can be reliably determined in the demagnetisation diagrams of both groups 1 and 2. Both normal and reversed components are revealed in both sections, which suggest a primary origin of the magnetic components. Due to the large dip of the strata, the viscous secondary magnetisations (Fig. 4a,b) can easily be distinguished from what we interpret as the characteristic remanent magnetisation (ChRM), which is here the component acquired before tilting [32]. The consistent directions after (variable) tilt corrections and the results of the rock magnetic analyses warrant the conclusion that the ChRM is the original NRM acquired during deposition.

The ChRM directions and polarity zones of the Putna section (Fig. 8) reveal 16 polarity reversals: 8 normal and 8 reversed intervals. The Râmnicu Sărat section (Fig. 9) recorded at least 17 reversals with 8 normal and 9 reversed polarity intervals. Some reversal boundaries are not exactly determined because of non-exposed intervals, but the long and clear polarity patterns allow an excellent correlation to the APTS.

4. Correlation to the APTS

We used the most recent APTS [16] to establish the magnetostratigraphic correlation of our sections. The most striking polarity pattern in the overlap interval between the Putna (Fig. 8) and Râmnicu Sărat (Fig. 9) sections concerns the two relatively short normal polarity zones in the Meotian, which are followed by a very long reversed zone in the Pontian (Fig. 10). The lengths of these polarity zones are in excellent agreement between the two sections. This characteristic pattern suggests a correlation to the interval in the timescale comprising subchrons C3An.2n, C3An.1r, C3An.1n and C3r, covering the time span between 6.7 and 5.2 Ma (Fig. 10). Upwards, this sequence is followed in Râmnicu Sărat by an interval of four short normal zones (and three short reversed zones) until another very long reversed zone is found. The four normal intervals correlate very well with the four C3n subchrons (Thvera, Sidufjall, Nunivak and Cochiti) of the reversed Gilbert Chron. The uppermost part of the Râmnicu Sărat section then probably comprises the Gauss Chron with normal subchrons C2An.3n and C2An.1n. The absence of the subchron C2An.2n in the Râmnicu Sărat magnetostratigraphic record can be explained by the lack of exposure in the interval.
between 5500 and 6100 m. The correlation of the polarity record of the lowermost part of the Putna section, below the gap coinciding with the 12° change in strike, is far from straightforward. The two most likely correlations of the relatively long normal interval are to C4n.2n or C4An (Fig. 10). Apparently, the change in strike is related to some serious deformation and/or possibly to a significant hiatus in the sequence.

By plotting the stratigraphic thickness versus the calibrated APTS ages (inset in Fig. 10), we observe that the accumulation rate records two clearly distinct trends, each having a nearly constant slope. The mean accumulation rate is 60 cm/ky for most of the Meotian. The accumulation rate increases more than two times (150 cm/ky) for the Pontian–Romanian interval.

5. Discussion

5.1. Astronomical forcing of sedimentary cyclicity

Astronomical forcing of the sedimentary cycles can be demonstrated by analyzing the average duration of a cycle. When the average period of the cycles in the time domain is constant and identical to one of the known Milankovitch periods, astronomical forcing is the most probable cause. Knowing the average sedimentation rate (per polarity zone) and the mean cycle thickness for specific intervals the average periodicity of the sedimentary cyclicity can be estimated. For such a demonstration, we have selected those intervals of the Putna and Rîmnicu Sârat
Sedimentary cyclicity is very distinct in the Pontian of the Putna section (Fig. 11a). The average sediment accumulation rate during the correlative reversed polarity interval of C3r is 137 cm/ky. The average thickness of the sedimentary cycles is 31.1 m, which results in a periodicity of 22.7 ky. The Meotian part of the Putna section also displays a clear sedimentary cyclicity, especially in the interval corresponding to C3An.1r (Fig. 11b). Sediment accumulation rate during this subchron is 80 cm/ky and the average thickness of the cycles is 18.5 m. The average periodicity for the sedimentary cyclicity in the Meotian is thus 23.5 ky. These results from the Putna section are in excellent agreement with the periodicity for precessional cycles of the Earth’s axis, and suggest that the sedimentary cyclicity in the Pontian and Meotian is indeed astronomically forced.

The Pontian of the Rımnicu Sărat section shows a similar type of cyclicity as in Putna although the sandy units are less distinct (Fig. 12a). Here the average sedimentation rate during C3r is 133 cm/ky and the average cycle thickness is 30.6 m. This results in a periodicity of 23 ky for the sedimentary cyclicity. The Dacian deposits of the Rımnicu Sărat also show a cyclic alternation of thick sandy units and more softer, clay-rich, intervals (Fig. 12b). The best-exposed interval correlates to subchron C3n.3n (Sidufjall) in which the accumulation rate is 91 cm/ky. The average cycle thickness is 21.75 m, which gives a mean periodicity of 23.9 ky. This is again in excellent agreement with a precessional forcing for the sedimentary cycles.

In all four cases, the values obtained for the duration of a cycle are very close to the average
periodicity of precession. If we assume that the very similar type of sedimentary cycles in the lower part of the Putna section are also forced by precession, we can use this relation to estimate the durations of the polarity zones in the lower part of the Putna section. There, the Sarmatian–Meotian interval also shows an interval with a similar regular alternation of sands and silts, correlative to the longest normal polarity zone of that part of the section (Fig. 13). Cycle thickness in this normal interval is 15 m, which suggests an accumulation rate of 0.69 m/ky and duration of 209 ky. This accumulation rate agrees very well with the uppermost Meotian values (Fig. 10), but the calculated duration does not allow a positive discrimination between the two most likely correlations (Fig. 10).

In all parts of the sections, the sedimentary cyclicity appears to be induced by precession/insolation processes. Clearly, a more rigorous analysis of quantitative records combined with biostratigraphic-environmental confirmation is needed to provide definitive proof for the astronomical origin of the cyclicity and to understand the phase relation in the succession. For example, it is still uncertain if the sandy units from Putna and Rımnicu Sărăț are the expression of wet or dry periods. There are two possible scenarios. The first suggests that sandy units were deposited during wetter periods, when the discharge from the rivers was bigger; the softer parts imply a low energy environment coincident with a decline of deposition. The second scenario presumes that the sand originated from the dry period, with poor coverage of vegetation, causing a rapid erosion and

Fig. 10. Correlation of the polarity sequences of the 2.3 km of Putna and of 7.3 km of Rımnicu Sărăț sections to APTS 2004. Solid lines between the section records and APTS 2004 connect (interpretative) simultaneous polarity boundaries. The dotted lines connect the most possible position of the C2An.2n into the Rımnicu Sărăț record. The names of the subchrons are in the column attached to the APTS. The C (Cochiti), N (Nunivak), S (Sidufjall) and T (Thvera) are the popular names for the names of the subchrons from the Gilbert Chron. The ages extracted from the map are on the right-hand side for Putna (internal column) and Rımnicu Sărăț (external column). The lighter gray area suggests the first possible correlation of the older (Sarmatian–Lower Meotian) Putna Valley, while the darker area is the second most possible correlation. The question marks indicate the position of the Sarmatian/Meotian boundary in the cases of the two possible correlations. The arrow indicates the place of the 12° changing in strike and the beginning of the certain magnetostratigraphic record. Inserted is the diagram illustrating the changes in the accumulation rates through time (according to APTS calibration points [2]). Note significant change of slope between the two intervals, at ca. 6 Ma. The calculated values are for the already compacted deposits.
increased sediment transport. The softer parts are related, in this case, to more vegetation and more protection against erosion and hence less sediment removal and transport.

The astronomical forcing of the sedimentary succession of the westernmost part of the Dacian basin has earlier been reported by Van Vugt et al. [12], who demonstrated the expression of eccentricity in the siliciclastic Lupoaia section of the southern Carpathian foredeep. Van Vugt et al. [36] suggested that the lacustrine detrital basins are dominantly forced by eccentricity and that the lignite seams represent the eccentricity maxima while the detrital intervals denote eccentricity minima. By analyzing the pollen from the same Lupoaia section, Popescu et al. [33] arrived at the same conclusion. In our study, however, we find evidence for precession rather than eccentricity-related cyclicity affecting the deposits of the eastern part of the Dacian basin. An explanation for this difference in response to astronomical forcing could
Fig. 12. Subvertically bedded sandy intervals, alternating with finer sediments on Rimpanicu Sarat Valley expressing the Milankovitch precession cycles: (a) in Pontian and (b) in Dacian deposits. See also the caption of Fig. 11.

Fig. 13. Vertically bedded sandstones, alternating with finer sediments on Putna Valley at the transition between Sarmatian and Meotian. See also the caption of Fig. 11.
be the difference in tectonic setting and basin configuration: the eastern Carpathian foredeep includes more than 8 km of Sarmatian–Romanian deposits, whereas the southern Carpathian foredeep consists of a maximum of 1.5 km for the time-equivalent deposits [34].

5.2. Age constraints for the Paratethys stages

Many recent studies of the Focșani basin use the boundaries of the Paratethys Sarmatian, Meotian, Pontian, Dacian and Romanian stages to estimate sediment accumulation rates and tectonic subsidence rates [13–15]. The stage boundaries are mainly defined by biostratigraphic studies [23,35–37] on molluscs and are displayed on the geological maps of the region (29 Covasna, 1:200 000 scale [38] and 113b Dumitrești, 1:50 000 scale [39]). These stage boundaries commonly correlate to lithological changes and are recognized as such in the seismic reflection profiles of the Carpathian foredeep [13,15]. Subsequently, the lateral changes in sediment thickness can be quantified for each stage. For estimation of (subsidence and accumulation) rates, however, additional age constraints have to be incorporated. The most recent studies on the tectonic evolution of the Focșani area use the timescale of Andreescu et al. [5] for ages of the stage boundaries [13–15].

Although we would prefer to have a detailed biostratigraphic record directly tied to our sequences, we must limit ourselves, in this phase, to the biostratigraphic data and the stage boundaries from the published geological maps [38,39] and the seismic profiles. Using the direct correlation to the APTS and the (biostratigraphically) determined stages on the maps and profiles, we can now numerically date the various stage boundaries of the sedimentary sequences of the Romanian segment of the eastern Carpathian foredeep (Fig. 10). We conclude that the Meotian/Pontian boundary is placed at the beginning of chron C3r at 5.9±0.1 Ma. The Pontian/Dacian boundary is located at the beginning of subchron C3n.3n (Sidufjall) and dated at 4.9±0.1 Ma. The Dacian/Romanian limit is situated at the base of chron C2Ar with an age of 4.07±0.5 Ma. Note that our errors relate to uncertainties in magnetostratigraphic age, but do not include uncertainties in the biostratigraphic positions of the stage boundaries. Nevertheless, our ages are in great contrast to the commonly used timescale of Andreescu et al. [5] (Fig. 2) and will thus have major consequences for tectonic studies. Correlation to the western Paratethys timescale is not useful at this stage because of the lack of detailed biostratigraphic criteria. We do not appraise the age for the Sarmatian–Meotian boundary because of the uncertainties in correlation of the acquired magnetostratigraphic record (of this older part) to the APTS.

In a recent study [13], based on a composite of seven outcrops from the western and central part of Dacian basin (southern Carpathian foredeep), the Meotian–Pontian boundary had been estimated at 6.15±0.11 Ma, within subchron C3An.1n, while the age of the Pontian–Dacian boundary is 5.30±0.1 Ma (Fig. 2). In the same source, the Dacian–Romanian boundary is recorded either at 4.58±0.05 Ma, in subchron C3n.2n (Nunivak), or at 4.25±0.0 Ma in subchron C3n.1n (Cochiti). These ages are slightly different from ages in our study. Nevertheless, both studies show the reduction in the duration of the Pontian to approximately 0.85 My.

5.3. Consequences for tectonic processes

Our magnetostratigraphic results infer that the accumulation rate increased considerably at the Meotian–Pontian boundary, from an average of 60 cm/ky during the Meotian to an average of 150 cm/ky during the Pontian to Romanian (Fig. 10). Until now, tectonic models suggested that the subsidence decreased during Meotian to Pontian times, but that a significant increase took place during the Dacian–Romanian time span [15]. Evidently, this difference in timing is related to the use of a different timescale [8], in which the Meotian is placed between 10 and 8 Ma, and the Pontian between 8 and 5 Ma. Our new results show that the duration of the Pontian is more than three times shorter than in most published timescales (Fig. 2). As a consequence, the calculated sedimentation and subsidence rates will also be significantly (about three times) higher. Our new age control demonstrates that the previous timescales for the eastern Paratethys are not appropriate for calculation of accumulation rates from the Carpathian foredeep. We show that the usage of unambiguous time control, offered by magnetic polarity stratigraphy, would substantially improve the interpretation of tectonic processes.
models and give a more realistic picture of the processes that have taken place in this tectonically active region.

A major change in accumulation rate in the eastern Paratethys thus appears sometime around the Meotian–Pontian transition, at approximately 5.8–6 Ma. This coincides remarkably well with the onset of the Messinian Salinity Crisis of the Mediterranean, dated at 5.96 Ma [1]. The increase in accumulation rate probably indicates a change in basin configuration, which may be related to the closure of the water connection between or within this part of the Paratethys and the Mediterranean.

6. Conclusions

A magnetostratigraphic record was made on long and continuous sections, comprising more than 7 km of sedimentary rocks, from the extremely thick Focșani sedimentary basin of the eastern Carpathian foredeep of Romania. The excellent results from the Putna and Rămnicu Sărat sections show that an almost complete chronostratigraphic framework can be derived, from approximately 7.2 to 2.5 Ma. Correlation of the magnetic polarity patterns to the APTS is straightforward (Fig. 10) except for the lowermost (Sarmatian) part of the Putna section.

The Meotian/Pontian, Pontian/Dacian and Dacian/Romanian boundaries have been dated, according to their biostratigraphic position on the geological maps. We conclude that the Meotian/Pontian boundary is placed at the beginning of chron C3r at 5.9±0.1 Ma. The Pontian/Dacian boundary is located at the beginning of chron C3n.3n (Sidufjall) and dated at 4.9±0.1 Ma. The Dacian/Romanian limit is situated at the base of chron C2Ar and its age is 4.07±0.5 Ma.

The new age control permits to calculate the accumulation rates for each polarity zone, i.e. for well-defined time intervals. Two major trends in the accumulation rate imply a significant increase (more than doubling) occurring at the transition Meotian to Pontian, during chron C3r. This change in accumulation rate appears at approximately 5.8–6 My, which coincides with the onset of the Messinian Salinity Crisis from the Mediterranean.

The calculated values for the average duration of the observed sedimentary cycles are very close to the average duration of a precession (21.7 ky) cycle, and indicate that the sedimentary cycles from the eastern Carpathian foredeep are astronomically forced.

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