Paleomagnetic and chronostratigraphic constraints on the Middle to Late Miocene evolution of the Transylvanian Basin (Romania): Implications for Central Paratethys stratigraphy and emplacement of the Tisza–Dacia plate

Arjan de Leeuw \textsuperscript{a,⁎}, Sorin Filipescu \textsuperscript{b}, Liviu Mațenco \textsuperscript{c}, Wout Krijgsman \textsuperscript{a}, Klaudia Kuiper \textsuperscript{c}, Marius Stoica \textsuperscript{d}

\textsuperscript{a} Paleomagnetic Laboratory ‘Fort Hoofddijk’, Utrecht University, Budapestlaan 4, 3584 CD, Utrecht, The Netherlands
\textsuperscript{b} Department of Geology, ‘Babeș-Bolyai’ University, Kogălniceanu str. 1, 400084 Cluj-Napoca, Romania
\textsuperscript{c} Faculty of Earth and Life Sciences, VU University Amsterdam, De Boelelaan 1085, 1081 HV Amsterdam, The Netherlands
\textsuperscript{d} Department of Geology, Faculty of Geology and Geophysics, University of Bucharest, Bălcescu Bd. 1, Bucharest, 010041, Romania

A R T I C L E   I N F O

Article history:
Received 6 May 2011
Revised 31 March 2012
Accepted 12 April 2012
Available online 15 May 2012

Keywords:
Miocene
chronostratigraphy
Paratethys
Transylvanian Basin sedimentary history \textsuperscript{40}Ar/\textsuperscript{39}Ar dating
paleomagnetism
magnetostratigraphy
tectonic rotation
environmental change

A B S T R A C T

From the Oligocene onwards, the complex tectonic evolution of the Africa–Eurasia collision zone led to paleogeographic and biogeographic differentiation of the Mediterranean and Paratethys, two almost land-locked seas, in the area formerly occupied by the western Tethys Ocean. Episodic isolation of the basins triggered strong faunal endemism leading to the introduction of regional stratigraphic stages for the Paratethys. Chronostratigraphic control on the Paratethys stages remains rudimentary compared to the cyclostratigraphically constrained Miocene stages. This lack of chronostratigraphic control restricts the insight in the timing of geodynamic, climatic, and paleobiogeographic events and thereby hinders the identification of their causes and effects. In this paper, we here derive better age constraints on the Badenian, Sarmatian and Pannonian Central Paratethys regional stages through integrated \textsuperscript{40}Ar/\textsuperscript{39}Ar, magnetostatigraphic, and biostratigraphic research in the Transylvanian Basin. The obtained results help to clarify the regions Miocene geodynamic and paleobiogeographic evolution. Six new \textsuperscript{40}Ar/\textsuperscript{39}Ar ages were determined for tuffs intercalating with the generally deep marine basin infill. Together with data from previous studies, there is now a total of 9 radio-isotopically dated horizons in the basin. These were traced along seismic lines into a synthetic seismic stratigraphic column in the basin center and served as first order tie-points to the astronomically tuned Neogene timescale (ATNTS). Paleomagnetically investigated sections were treated similarly and their polarity in general corroborates the \textsuperscript{40}Ar/\textsuperscript{39}Ar results. The integrated radio-isotopic and magnetostatigraphic results provide an improved high-resolution time-frame for the accumulation of the Dej Tuff Complex in response to a period of intensive volcanism, the onset of which is constrained between the first occurrence (FO) of Orbulina suturalis at 14.56 Ma and 14.38 ± 0.06 Ma. During the subsequent Badenian Salinity Crisis (BSC) up to 300 m of salt accumulate in the basin center. The faunal turnover that marks the Badenian–Sarmatian Boundary is dated at 12.80±0.05 Ma. A second phase of intense volcanism occurs at 12.4 Ma and leads to deposition of the middle Sarmatian tuff complex (Ghiriș, Hădăreni, Turda and Câmpia Turzii tuffs). Rates of sediment accumulation strongly diminish in the basin center at the onset of the Pannonian stage coincident with an approximately 20° CW tectonic rotation of the Tisza–Dacia plate. Concurrent enhanced uplift in the Eastern a’nd Southern Carpathians leads to the isolation of the Central Paratethys and triggers the transition from marine to freshwater conditions. An additional Pannonian to post-Pannonian 6° of CW rotation is related to the creation of antiform geometries in the Eastern Carpathians which are notably larger in the north than in the south. An 8.4 Ma age is determined for the uppermost Pannonian sediments preserved in the central part of the Transylvanian Basin. Two sections belonging to middle Pannonian Zone D, and the lower part of Zone E (Subzone E1) are found to cover the 10.6–9.9 Ma time-interval.

© 2012 Elsevier B.V. All rights reserved.

1. Introduction

During the Cenozoic, collision of the northward moving African plate with Eurasia led to the Alpine Orogeny. Along the course of this process, the vast water mass of the Tethys Ocean disintegrated and gave birth to the Mediterranean and Paratethys Seas (e.g. Seneș, 1973; Rögl, 1999) (Fig. 1). The complex tectonic evolution of the Africa–Eurasia convergence...
zone, with its multitude of colliding micro-continents (Csontos et al., 1992; Stamfl and Kozur, 2006; Schmid et al., 2008; Ustaszewski et al., 2008), induced a similarly intricate regional paleogeographic history with frequently changing seaways and land-bridges (Rögl, 1999; Harzhauser et al., 2002; Popov et al., 2004, 2006; Harzhauser and Piller, 2007).

Gradual growth of the Alpine-Carpathian-Dinaridic orogenic system induced progressive restriction of the Western, Central and Eastern Paratethys (Fig. 1) Repeated isolation led to large changes in water chemistry (Matyas et al., 1996; Peryt, 2006; Harzhauser et al., 2007; Vasiliev et al., 2010b) and inflicted periods of severe endemism (e.g. Steininger et al., 1988; Magyar et al., 1999a; Harzhauser et al., 2002, 2003; Kovač et al., 2007). This geodynamically controlled paleogeographic and biogeographic differentiation caused major difficulties in stratigraphic correlation between the different parts of the Paratethys and the Mediterranean (Piller and Harzhauser, 2005) and thus led to the establishment of regional chronostratigraphic scales for the Central Paratethys summarized in the series “Chronostratigraphie und Neostratotypen” (Cicha et al., 1967; Steininger et al., 1971; Papp et al., 1973, 1974; Báldi and Seneš, 1975; Papp et al., 1978, 1985; Stevanovic et al., 1990). A high resolution chronologic framework was established over the past few decades for the Mediterranean stages, on which the global timescale currently relies (Lourens et al., 2004a). In comparison, chronologic control on most of the Paratethys regional stages remains rudimentary, although significant progress has been achieved in the last decade (e.g. Krijgsman et al., 2010 and references therein; Lirer et al., 2009; Paulissen et al., 2011). In particular, precise dating of the Middle-Late Miocene sediments of the Central Paratethys (Geary et al., 2000; Magyar et al., 2007; Vasiliev et al., 2010a), remains an outstanding problem.

In order to overcome part of this problem we conducted an integrated biostratigraphic, magnetostratigraphic and geochronologic study in the Transylvanian Basin. This basin was part of the Central Paratethys during the Miocene and it exposes a thick pile of siliciclastic deposits suitable for paleomagnetic investigations. A number of intercalating volcano-sedimentary products (mostly tuffs) provide key stratigraphic horizons potentially datable with the 40Ar/39Ar method. The continuity of sedimentation, reasonable exposures due to tectonic exhumation of the basin towards the end of the Miocene, and a significantly lower amount of tectonic disturbance and related structural complications and unconformities (Sztanó et al., 2005; Horváth et al., 2006; Krézsek et al., 2010) in comparison with other parts of the Central Paratethys, make the Transylvanian Basin a favorable study location. The limited size of the Transylvanian Basin and its mature stage of exploration (Krézsek et al., 2010) provide the opportunity to correlate outcrops over large distances based on subsurface imagery.

The Transylvanian Basin also provides an optimal location to examine the intriguing post-20 Ma rotation associated with the invasion of the Tisza-Dacia plate into the Carpathian embayment, because of its largely undeformed Middle-Late Miocene sedimentary succession. A precise quantification of the translation mechanics of Tisza-Dacia around the Moesian indenter, driven by the roll-back of a slab attached to the European continent (e.g. Balla, 1987; Royden, 1988; Ustaszewski et al., 2008), is hampered by contrasting amounts of rotation derived by the paleomagnetic, paleogeographic and structural geology studies, which vary from 10° to almost 70° clockwise rotation after the Paleogene (Csontos et al., 1992; Fügenschuh and Schmid, 2005; van Hinsbergen et al., 2008).

We aim to derive better age constraints on the Middle to Late Miocene regional stages (Badenian, Sarmatian, and Pannonian) of the Central Paratethys through integrated magnetostratigraphic, radio-isotopic and biostratigraphic research in Transylvania. Taking advantage of the resulting chronostratigraphic and paleomagnetic data, the amount as well as the partitioning in time of the Miocene rotation of Tisza-Dacia can be unravelled. These data provide a more detailed insight in the mechanism of emplacement of the Tisza–Dacia block, supported by recent advances in the understanding of the exhumation and kinematics of the Carpathians.

2. Geological setting

The Transylvanian Basin is a 200 km long and 250 km wide semi-isolated back-arc basin (Krézsek et al., 2010) situated in the eastern part of the Central Paratethys and bounded by the Eastern Carpathians, Southern Carpathians and Apuseni Mountains (Fig. 2). The Middle to Late Miocene subsidence and subsequent exhumation of the Transylvanian Basin are both closely related to, and concurrent with, major episodes of deformation in the Carpathians in conjunction with the subduction of the Eastern European underneath the Tisza–Dacia plate and the collision that followed (Matenco et al., 2010).

During the Early Miocene deep marine conditions were restricted to the northern part of the basin while continental settings prevailed elsewhere. A regional transgression during the Middle Miocene established relatively deep marine settings throughout the basin with mixed carbonate-siliciclastic platforms near the margins and siliciclastic environments in deeper areas (Fig. 3, Filipescu and Gîrbaecu, 1997; Krézsek et al., 2010). At this time, the Central Paratethys was well connected with the Mediterranean (Fig. 1) and there was a great similarity in their paleontological record. The onset of calc-alkaline volcanism related to subduction processes in the exterior Paratethys induced acidic volcanism that led to widespread tuff deposition in the Central Paratethys (Pécskay et al., 1995; Seghedi et al., 2004; Pécskay et al., 2006). In the Transylvanian Basin an up to 50 m thick tuff complex accumulated (Fig. 3, Seghedi and Szákacs, 1991). This so called Dej Tuff thus constitutes a major lithostratigraphic marker within the upper 100 m thick lower Badenian sedimentary package.

Severe restriction of the Central Paratethys at 13.82 Ma (Peryt, 2006; de Leeuw et al., 2010) induced hypersaline conditions. These led to the deposition of around 300 m of salt in the deeper parts of the Transylvanian Basin and gypsum along the western margin (Krézsek and Bally, 2006).

In the late Badenian, the Central Paratethys re-connected with the Eastern Paratethys that occupied the current Black Sea and Caspian Sea regions. Hypersaline conditions disappeared even though the re-united Paratethys remained semi-isolated (Popov et al., 2004; Kovač et al., 2007). At this time, subsidence rates in the Transylvanian Basin increased tremendously and a locally over 3 km thick, upper Badenian to Pannonian siliciclastic sedimentary infill accumulated (Krézsek and Filipescu, 2005). The lowermost post-salt succession onlaps the evaporites in the northern, western and southern parts of the basin and thickens from a few 100 m in the west to more than 2000 m in the southeast (Krézsek and Bally, 2006). Depositional
environments were characterized by clay-dominated deep marine fans (Krézsek and Filipescu, 2005). During the late Badenian and Sarmatian, the basin’s depocenter was located at its current eastern margin and the corresponding part of the infill thus thins out towards the west (Krézsek et al., 2010).

The Sarmatian part of the basin infill comprises siliciclastics, such as marls and sandstones, subordinate conglomerates and evaporites. These demonstrate a higher variety of sedimentary environments. Sea-level changes in combination with incipient tectonics in the adjacent mountain chains triggered alternation between normal marine, brackish, and lacustrine conditions (Krézsek et al., 2010).

At the advent of the Pannonian, the Central Paratethys became fully isolated from the marine realm, and turned into a lake (Magyar et al., 1999a). The Pannonian sediments of the Transylvanian Basin comprise lacustrine fans and lowstand deltas with sand dominated facies in the more proximal-, and marl dominated facies in the central, more distal part of the basin (Krézsek et al., 2010). The regressive trend was interrupted by a phase of regional uplift towards the end of the Pannonian
that exhumed the basin to its present-day altitude of between 300 and 600 m (Matenco et al., 2010).

Volcanic activity was common in the neighbouring Eastern Carpathians and Apuseni Mountains during the Badenian, Sarmatian and Pannonian (Szakács and Seghedi, 1995; Seghedi et al., 2004). It generated a large number of volcaniclastics in the sedimentary sequence of the Transylvanian Basin. These among others include the Apahida Tuff, Hâdârenei Tuff, Ghiriş Tuff, and Oarba Tuff.

3. Stratigraphic framework and sampling approach

Our research relies on the stratigraphic framework of Krézsek and Filipescu (2005), and Krézsek and Bally (2006). Sections were investigated using an integrated stratigraphic approach taking multiple samples for both biostratigraphic and paleomagnetic analysis at the same stratigraphic level. We focused on sections that include tuffs and sandstones and sampled them for $^{40}\text{Ar}/^{39}\text{Ar}$ dating, and biostratigraphy. We use the Astronomically Tuned Neogene Timescale (Lourens et al., 2004a) with adaptations by Hüsing et al. (2010) and supplemented with short term fluctuations in the geomagnetic field identified by Krijgsman and Kent (2004) as a global chronostratigraphic reference framework.

In order to retrieve the relative stratigraphic relation, all sections (Fig. 2) were traced along seismic reflectors to a composite stratigraphic column (Fig. 4) near the basin’s depocenter where the Middle to Upper Miocene succession is as complete as possible (Fig. 2). For each of the sections, an uncertainty in stratigraphic position was estimated based on the proximity to available seismic lines, clarity of line-to-line reflector correlation and the presence of potential sedimentary architectures affecting precision, such as clinoforms with lower reflectivity. The synthetic seismic section in Fig. 4 is vertically scaled in seconds two-way travel time. A depth conversion of all elements projected into this line was performed using average interval velocities derived from neighboring well logs (sonic or density logs, see De Broucker et al., 1998; Krézsek et al., 2010).

4. $^{40}\text{Ar}/^{39}\text{Ar}$ dating of key stratigraphic horizons

There are several tuffs in the Middle to Late Miocene sedimentary infill of the Transylvanian Basin, discussed in detail by Bedelean et al. (1991). Some of these, e.g. the Bor a-Apahida, Hâdârenei, Ghiriş, Sârmă el and Bazna tuffs, extend over a wide area. Their strong acoustic impedance contrast facilitates correlation across seismic lines (Ionescu, 1994) and they represent excellent regional stratigraphic markers. Several of these tuff levels were sampled for $^{40}\text{Ar}/^{39}\text{Ar}$ dating in order to acquire absolute ages for the corresponding stratigraphic horizons.

4.1. Methods

The volcanic ashes were processed at the Department of Isotope Geochemistry (VU University Amsterdam). Bulk samples were crushed, disintegrated in a calgon solution, washed and sieved over a set of sieves between 63 and 250 μm. The residue was subjected to standard heavy liquid as well as magnetic mineral separation techniques for k-feldspar. Except for the sample from Apahida, all samples contained ample k-feldspar for $^{40}\text{Ar}/^{39}\text{Ar}$ dating. The k-feldspar separates were leached with a 1:5 HF solution in an ultrasonic bath for 5 min. After careful handpick, the samples were loaded in a 10 mm ID quartz vial together with Fish Canyon Tuff (FC-2) and Drachenfels (Dra-1, IZ500-500 and Dra-2, f>500) sanidine that served as an age-monitor with Fish Canyon Tuff (FC-2) and Drachenfels. The vial was irradiated in the Oregon State University TRIGA reactor in the cadmium shielded CLICIT facility for 10 h.

Upon return in the laboratory, mineral separates were split into at least 9 duplicate fractions and loaded in a Cu-tray. The loaded Cu-tray was pre-heated to –200 °C under vacuum using a heating stage and a heat lamp to remove undesirable atmospheric argon. The tray was then placed in the sample house and the system (extraction line + sample house) was degassed overnight at ~150 °C. Incremental heating was performed with a Synrad CO₂ laser in combination with a Raylase scanhead as a beam delivery and beam diffuser system. After purification the resulting gas was analyzed with a Mass Analyzer.
Products LTD 215-50 noble gas mass spectrometer. Beam intensities were measured in a peak-jumping mode in 0.5 mass intervals over the mass range 40–35.5 on a Balzers 217 secondary electron multiplier. System blanks were measured every three to four steps. Mass discrimination was monitored by frequent analysis (~every 10 h) of aliquots of air. The irradiation parameter J for each unknown was determined by interpolation using a second-order polynomial fitting between the individually measured standards.

All age calculations use the decay constants of Steiger and Jäger (1977). Steps with less than 1% radiogenic argon were immediately discarded. The age for the Fish Canyon Tuff sanidine flux monitor used in age calculations is 28.201±0.03 Ma (Kuiper et al., 2008). The age for the Drachenfels sanidine flux monitor is 25.42±0.03 Ma (Kuiper et al. in preparation). Correction factors for neutron interference reactions are 2.64±0.017×10^{-4} for (^{36}Ar/^{37}Ar)_{Ca}, 6.73±0.037×10^{-4} for (^{39}Ar/^{37}Ar)_{Ca}, 1.211±0.003×10^{-2} for (^{38}Ar/^{39}Ar)_{K} and 8.6±0.7×10^{-4} for (^{40}Ar/^{39}Ar)_{K}. Errors are quoted at the 1σ level and include the analytical error and the error in J. All relevant analytical data as well as error determination can be found in the online supplementary material.

4.2. Results

The calculated weighted mean ages, peaks in probability density distribution, isochron ages and several other parameters for each of the investigated ashes are listed in (Table 1). The full analytical data are available in the supplementary material. The MSWD was smaller than the statistical T ratio at the 2 sigma level for all reported weighted mean ages. Isochron ages are concordant with the weighted mean ages and the trapped $^{40}Ar/^{36}Ar$ component is in all cases atmospheric. The calculated weighted mean ages are also concordant with the peaks in probability density distribution and we interpret them to reflect the respective tuff’s crystallization age.

The lower Badenian part of the basin infill is dominated by volcaniclastic deposits of the Dej Tuff complex (Fig. 2). We dated two samples of the Dej Tuff taken from outcrops near Ciceu Giurgești and Dej. For these samples a larger number of repetitive experiments had to be performed, since both contained a large reworked component. This resulted in abnormally high ages for part of the experiments that therefore had to be rejected. For each of the samples at least 7 experiments provided consistent, homogenous and stratigraphically realistic ages (Fig. 5). These were selected to calculate their weighted mean ages. The resulting statistically equivalent weighted mean ages of the samples of the Dej Tuff taken at Ciceu Giurgești and Dej are 14.38±0.06 Ma and 14.37±0.06 Ma (Table 1).

Several tuff levels occur within a small stratigraphic distance in the middle Sarmatian part of the basin infill. The lowermost of these is the Hădăreni Tuff, which is recognized as a seismic reflector throughout the larger part of the basin (Fig. 4). It crops out in a classic locality at the base of the Brâul Alb (‘White Belt’) hill near Hădăreni where it was sampled. At the top of the same hill, stratigraphically about 100 m higher, there is another tuff level. This level, known as the andesitic Chiriș Tuff (Márza and Mészáros, 1991), also crops out at a site just west of Chiriș Român where it was sampled. Between the Apuseni Mountains and Hădăreni, where the E-W flowing Arieș...
River has cut outcrops in the south-directed hillsides, the diapirs of the western salt diapir alignment (Krózek and Bally, 2006) cause tilting and multiple repetition of upper Badenian to middle Sarmatian strata. Two additional middle Sarmatian tuff levels were sampled in outcrops near Turda and Câmpia Turzii respectively (Fig. 2). The $^{40}\text{Ar}/^{39}\text{Ar}$ experiments for the Turda, Hâdârâni, Câmpia Turzii and Ghiriş tuffs provided homogenous age populations (Fig. 5). The peaks in the probability density distributions are concordant with the calculated weighted mean ages and we interpret them to reflect the respective tuff’s crystallization age. The Turda, Hâdârâni, Câmpia Turzii and Ghiriş tuffs are thus respectively 12.37 ± 0.04 Ma, 12.35 ± 0.04 Ma, 12.38 ± 0.05 Ma and 11.62 ± 0.04 Ma old (Table 1).

### 5. Resulting time frame and sedimentation rates

The performed $^{40}\text{Ar}/^{39}\text{Ar}$ measurements thus provided ages for six stratigraphic horizons. The number of horizons with a radio-isotopically determined age can be augmented with the base and top of the evaporite sequence for which a 13.8 Ma and 13.36 Ma $^{40}\text{Ar}/^{39}\text{Ar}$ age was recently established (de Leeuw et al., 2010; de Leeuw et al., unpublished results; de Leeuw, 2011). Vasiliev et al. (2010a) furthermore determined an 11.62 ± 0.04 Ma age for Oarba Tuff that is located in the uppermost Sarmatian part of the basin infill (Fig. 4). These radio-isotopic ages provide a first order timeframe for the Middle to Late Miocene infill of the Transylvanian Basin that enables the calculation of average sedimentation rates.

At the onset of the Badenian the Paratethys was well connected with the Mediterranean and endemism does not hamper biostatigraphic correlation to the global timescale. The base of the Badenian coincides with the Early-Middle Miocene boundary (Papp et al., 1978) which is dated at 15.97 Ma (Lourens et al., 2004a). The 13.8 Ma onset of evaporite deposition provides a minimum age for the top of the lower Badenian. Whereas the early Badenian thus lasted for over 2 Ma, not more than 100 m of sediments accumulated (Krózek and Filipescu, 2005). Rates of deposition were consequently rather low (maximum of 0.05 m/k.yr).

After the Badenian salinity crisis (BSC), rates of deposition increase tremendously. The upper Badenian to upper Sarmatian part of the basin infill is around 2.6 km thick in the basin centre (Fig. 6). The 13.36 Ma termination of the BSC provides a maximum age for its base and the 11.62 Ma old Oarba Tuff provides an approximate age for its top. Based on these two absolute ages an average sedimentation rate of 1.5 m/k.yr can be calculated.

Our synthetic seismic section (Fig. 4) demonstrates that the Hâdârâni, Ghiriş, Turda and Câmpia Turzii tuffs are all distributed in an interval with a thickness lower than 0.1 sTWT, which is around 130 m given the average interval velocity of Sarmatian sediments. We collectively call these tuffs the ‘middle Sarmatian tuff complex’ (MSTC) since they demonstrably co-occur in a relatively limited stratigraphic interval and according to our $^{40}\text{Ar}/^{39}\text{Ar}$ data follow upon each other in a short time-interval.

Based on our composite stratigraphic column for the late Badenian and Sarmatian (Fig. 6), there are about 1.44 km of sediments between the top of the salt (13.36 Ma) and the base of the MSTC (12.36 Ma). The respective upper Badenian and lower Sarmatian sediments were thus deposited at a rate of approximately 1.44 m/k.yr. The upper Sarmatian sediments overlying the MSTC have a stratigraphic thickness near 1.16 km. Using the 11.62 Ma age for the Oarba Tuff as tie-point this package was deposited at a comparable rate of

### Table 1

$^{40}\text{Ar}/^{39}\text{Ar}$ ages for several tuffs of the Transylvanian Basin. MSWD is Mean Square Weighted Deviates, N is the total number of repetitions in the single fusion experiments and n is the amount included when calculating the weighted mean and inverse isochron age. %$^{40}\text{Ar}(r)$ is the radiogenic amount of $^{40}\text{Ar}$. Errors are given at 95% confidence level. MSWD, $^{40}\text{Ar}/^{39}\text{Ar}$, and Inverse isochron intercept were determined based on the experiments selected for calculation of the weighted mean or isochron age. Ages in bold are considered to most accurately represent the respective tuffs crystallization age.

<table>
<thead>
<tr>
<th>Sample Laboratory ID</th>
<th>Location</th>
<th>Method</th>
<th>Weighted mean age (Ma)</th>
<th>n/N</th>
<th>MSWD</th>
<th>%$^{40}\text{Ar}(r)$</th>
<th>Mineral</th>
<th>Cystal size (μm)</th>
<th>Inverse isochron age</th>
<th>Inverse isochron Intercept</th>
<th>Isochron Reference</th>
</tr>
</thead>
<tbody>
<tr>
<td>Dej AL62, VU78-25</td>
<td></td>
<td></td>
<td><strong>14.37 ± 0.06</strong></td>
<td></td>
<td></td>
<td>7/13</td>
<td>14.34</td>
<td>95 k-feldspar</td>
<td>200–250</td>
<td>14.29 ± 0.13</td>
<td>328 ± 48 1.46</td>
</tr>
<tr>
<td>Ciceu</td>
<td></td>
<td></td>
<td><strong>14.38 ± 0.06</strong></td>
<td></td>
<td></td>
<td>8/18</td>
<td>14.41</td>
<td>90 k-feldspar</td>
<td>180–200</td>
<td>14.31 ± 0.10</td>
<td>309 ± 17 0.94</td>
</tr>
<tr>
<td>Ghirișu AL49, Câmpia Turzii VU79-19</td>
<td>Multi total fusion AL58, VU78-24</td>
<td>Multiple total fusion</td>
<td><strong>12.35 ± 0.04</strong></td>
<td>14/7</td>
<td>0.52</td>
<td>12.35</td>
<td>99 k-feldspar</td>
<td>250–500</td>
<td>12.36 ± 0.04</td>
<td>292 ± 48 0.57</td>
<td>This study</td>
</tr>
<tr>
<td>Hâdârâni AL59, Câmpia Turzii VU78-22</td>
<td>Multiple total fusion AL56, Câmpia Turzii VU78-26</td>
<td>Multiple total fusion</td>
<td><strong>12.37 ± 0.05</strong></td>
<td>9/13</td>
<td>0.13</td>
<td>12.37</td>
<td>88 k-feldspar</td>
<td>250–500</td>
<td>12.37 ± 0.05</td>
<td>294 ± 4 0.13</td>
<td>This study</td>
</tr>
<tr>
<td>Ghirișu Român AL56, Câmpia Turzii VU78-17</td>
<td>Multiple total fusion AL58, Câmpia Turzii VU78-21</td>
<td>Multiple total fusion</td>
<td><strong>12.38 ± 0.04</strong></td>
<td>8/9</td>
<td>0.75</td>
<td>12.39</td>
<td>98 k-feldspar</td>
<td>200–250</td>
<td>12.38 ± 0.05</td>
<td>300 ± 53 0.87</td>
<td>This study</td>
</tr>
<tr>
<td>Oarba OM-C, Oarba-C VU59-5</td>
<td>Multiple total fusion AL58, Oarba-C VU59-5</td>
<td>Multiple total fusion</td>
<td><strong>11.62 ± 0.04</strong></td>
<td>11/13</td>
<td>1.71</td>
<td>12.39</td>
<td>98 k-feldspar</td>
<td>200–250</td>
<td>12.36 ± 0.05</td>
<td>315 ± 36 1.15</td>
<td>This study</td>
</tr>
</tbody>
</table>

*Fig. 5. Probability density diagrams for the multiple total fusion experiments for the investigated tuffs.*
around 1.56 m/kyr. This indicates that from the end of the Badenian Salinity Crisis to the final part of the Sarmatian deposition in the Transylvanian Basin was very rapid and occurred at a fairly constant rate. Lower sedimentation rates are obviously to be expected near the western basin margin and in the vicinity of rising diapirs that started their upward movement already in the late Badenian (Krészek and Filipescu, 2005).

It should be noted that the calculated Badenian, Sarmatian and Pannonian sedimentation rates do not permit direct quantitative conclusions on basin evolution without a proper subsidence history analysis taking into account factors such as paleo-bathymetry, compaction and sea-level variations. However, these sedimentation rates provide a first order qualitative indication of the sedimentary influx associated with processes such as subsidence or increases in input from the source areas.

### 6. Paleomagnetism

Sections were sampled for paleomagnetic investigation since their magnetostratigraphic polarity pattern can provide additional constraints for the correlation to the global timescale and the directional data can be used to unravel the amount as well as the partitioning in

![Fig. 6. Biostratigraphy, magnetostratigraphy and 40Ar/39Ar ages for the investigated upper Badenian to Sarmatian sections and tuffs, and their consequential correlation to the ATNTS (Krijgsman and Kent, 2004; Lourens et al., 2004a; Hüsing et al., 2010). Stratigraphic column and position of the sections therein based on the synthetic section in Fig. 4. Scaling and first order correlation of the stratigraphic succession to the ATNTS is based on the radio-isotopic ages for the top of the evaporites and the Oarba Tuff. Due the scale imposed by the high sedimentation rate detailed lithological changes in the depicted individual sections would be indiscernible. Detailed bio-, litho-, and magnetostratigraphic information on each of the outcrops is provided in the supplementary material.](image-url)
time of the Miocene rotation of the Transylvanian Basin (see Fig. 11 and Table 2 for locations and results).

We will discuss the magnetic polarity pattern of the investigated sections, taking the composite stratigraphic column (Figs. 6, 8) and the stratigraphic position of the sampled sections inferred from it as a starting point. In order to place the magnetostratigraphic pattern of the investigated sections into the composite stratigraphic column at the right scale, a scaling factor has to be determined. According to the geological map (Raileanu et al., 1968) and our own observations the stratigraphic distance between the Badenian–Sarmatian boundary and the Câmpia Turzii Tuff is around 400 m in the area where the respective sections were sampled. In the basin center, the corresponding interval is around 500 m and thus 1.25 times as thick. The upper Badenian to middle Sarmatian sections were scaled accordingly in the composite stratigraphic column. The stratigraphic distance between the Oarba Beta and A outcrop is, based on field data, approximately 530 m. In the composite stratigraphic column, the same stratigraphic interval is nearly 800 m, or 1.5 times as thick. The magnetostratigraphic pattern of Oarba Beta was thus scaled accordingly, which indicates that its earlier correlation to the ATNTS by Vasiliev et al. (2010a) does not agree with the correlation to our composite stratigraphic column. Its limited stratigraphic extent in combination with its position in the stratigraphy precludes the occurrence of two reversals (Figs. 6, 8).

Approximate scaling factors for the Pannonian sections can also be derived from the sequence of outcrops at Oarba de Mure. The stratigraphic distance between Oarba A and D is, based on field data, around 120 m. In the composite stratigraphic column, the same interval covers about 215 m and is thus 1.8 times as thick. All Pannonian sections are scaled accordingly.

6.1. Methods

All samples were collected with a hand-held electric drill. The orientation of the standard paleomagnetic cores and corresponding bedding planes were measured by means of a magnetic compass, and corrected for the local magnetic declination. In the laboratory, thermal as well as alternating field demagnetization techniques were applied to isolate the characteristic remanent magnetization (ChRM). The natural remanent magnetization (NRM) of the samples was measured after each demagnetization step on a 2G Enterprises DC Squid cryogenic magnetometer (noise level 3×10−12 Am2). Heating occurred in a laboratory-built, magnetically shielded furnace employing 10–30 °C temperature increments. AF demagnetization was accomplished by a laboratory-built automated measuring device applying 5–20 mT increments up to 100 mT by means of an AF coil interfaced with the magnetometer. The presence of iron sulfides in the studied samples was anticipated. In order to overcome the problem of gyromagnetic during alternating field demagnetization the specifically designed per component demagnetization scheme of Dankers and Zijderveld (1981) was applied. In addition, small (2–5 mT) field steps were taken in the 20–40 mT range. The ChRM was identified through assessment of decay-curves and vector end-point diagrams (Zijderveld, 1967). ChRM directions were calculated by principal component analysis (Kirschvink, 1980) and are based on at least four consecutive temperature or field steps.

6.2. Results

Several characteristic demagnetization diagrams, of mostly marls and clays, are depicted in Fig. 7. Their NRM intensity ranges between 1 and 1000 × 10−4 A/m. A viscous overprint is generally removed at 100 °C or 15 mT respectively. During progressive stepwise thermal demagnetization two and sometimes three components can be established. A first, low temperature component appears between 100 and 200 °C. The corresponding directions are generally of normal polarity and close to the present day field direction. A notable exception is the Câmpia Turzii section, where the 100–200 °C component bears both normal and reversed polarities. A second component appears between 200 and 300 °C. For the Badenian and Sarmatian sections, the corresponding directions are clearly different from the present-day field direction and indicative of a significant counterclockwise rotation. Approximately 90% of the NRM had been removed at 300 °C in samples from the Fâgâraș, Gherla and Câmpia Turzii sections. For these localities we interpret the 200–300 °C component as the ChRM. For samples of the oimuș and Viforoasa sections, a higher temperature (300–360 °C) component could be established. At 360 °C generally 99% of their NRM had been removed. We thus interpret the
300–360 °C as the ChRM for the Viforoasa and oimuș sections. In alternating field demagnetization diagrams of marly samples generally only a single component appears. The corresponding directions are established in the 20–40 mT interval. Above 40 mT the onset of gyroremanence often distorts the demagnetization diagrams.

The samples taken from the Pâglișa section consisted of volcanic tuff or tuftite, and displayed a different demagnetization behavior. Their NRM intensity was generally around $1000 \times 10^{-4}$ A/m. Two components were identified during thermal demagnetization. A first component appears between 100 and 260 °C. The corresponding directions are close to the present day field direction and we interpret this component as a present-day field overprint. A second component is isolated between 280 and 480 °C. Although also of exclusively normal polarity, the corresponding directions differ significantly from the present-day field direction. We interpret the 280–480 °C component as the ChRM of the Pâglișa section.

The ChRM temperature interval and gyroremanent behavior upon AF demagnetization indicate that the magnetic carrier in most samples from the Transylvanian Basin is an iron sulfide, most likely greigite. Iron sulfides are very commonly the main magnetic carrier.
in Middle to Late Miocene of the Paratethys (Vasiliev et al., 2010a). Correlation of the iron sulfide based magnetic polarity patterns of boreholes in the Pannonian Basin to the GPTS (Sacchi et al., 1997; Juhász et al., 1999; Magyar et al., 1999b; Sacchi and Horváth, 2002; Sacchi and Müller, 2004; Magyar et al., 2007) is often intricate. The reliability of greigite as a magnetic carrier has moreover frequently been questioned, because greigite is thermodynamically metastable and the timing of NRM acquisition by greigite is not well constrained due to its diagenetic formation (Vasiliev et al., 2008). The straightforward correlation of the magnetostratigraphy of greigite bearing sections in the Carpathian Foredeep (Vasiliev et al., 2007), and Mediterranean (Hüsing et al., 2009), however, shows that greigite may preserve its NRM for geological times and that reliable magnetostratigraphic results can be obtained if only a dedicated greigite specific demagnetization approach is taken.

6.3. Correlation to the timescale

The polarity of the sampled sections in general corresponds to the polarity predicted for the respective stratigraphic interval by the time-scale correlation based on $^{40}$Ar/$^{39}$Ar data and thus provides additional support (Fig. 6). The Făgăra outcrop, which has a reversed polarity, correlates to chron C5Ar.3r. The marls underlying the tuff exposed at Apahida were deposited during chron C5Ar.2n, and the reversed polarity interval exposed at Gherla correlates to the base of chron C5Ar.1r. At the top of the Turda section the magnetic polarity changes from normal to reverse. This reversal correlates to the C5An.2n to C5An.1r transition taking the acquired $12.37 \pm 0.04$ Ma radio-isotopic age for the Turda Tuff into account.

The magnetostratigraphic pattern of the Câmpia Turzii, however, requires additional explanation. The 150 m thick Câmpia Turzii section covers ~0.2 Ma. Its polarity is dominantly reversed, but a large number of single samples have a normal polarity. The demagnetization diagrams on which the magnetostratigraphic pattern is based are of high quality for both normal and reversed samples. They do, on the other hand, all display multiple and sometimes even up to four components. These can be of normal or reversed polarity independent of their demagnetization temperature or field interval. We consider it most likely that the primary magnetic component of the Câmpia Turzii section is reversed, but demagnetization behavior is so complex that reliable magnetostratigraphic information is hard to extract.

For the uppermost Sarmatian to Pannonian part of the basin infill we do not have $^{40}$Ar/$^{39}$Ar dates that can guide first order estimates on sedimentation rates and correlation to the ATNTS. We thus have to rely on the magnetostratigraphic information acquired (Fig. 9). For the correlation of the magnetostratigraphic pattern of the Oarba sections we follow (Vasiliev et al., 2010a), except for Oarba Beta. The polarity of the newly sampled Pannonian sections and sites is generally normal. We thus correlate the Târnăveni, Viforoasa, and oimuş sections to the long normal Chron C5n.2n (Fig. 9). The short reversed intervals of the Viforoasa section might represent small-term fluctuations in the geomagnetic field (Krijgsman and Kent, 2004). The uppermost part of oimuş, which bears a reversed polarity, in our view correlates to C5n.1r. This correlation implies that in the Pannonian the sedimentation rate is much lower than during the upper Badenian to Sarmatian time-interval and amounts to 0.36 m/kyr (Fig. 10). This is in agreement with seismic interpretations that demonstrate a gradual change from subsidence to uplift during the Pannonian stage (Matenco et al., 2010). Our synthetic seismic stratigraphy shows that the Sighi oara and Mediaş sections are slightly younger than oimuş. They have been correlated to the timescale based on the calculated sedimentation rate for the Pannonian part of the basin infill (Fig. 9).

Along the course of our study it has become clear that the quality of demagnetization results for and the preservation of NRM in the sediments of the Transylvanian Basin is highly variable. Sections for which demagnetization results were of low quality were immediately excluded from further analysis. The polarity of most sections with high quality demagnetization diagrams corresponds to the one expected based on correlation to the GPTS (Lourens et al., 2004a) according to our $^{40}$Ar/$^{39}$Ar based chronostatigraphic framework. This consistency provides independent support for the reliability of the acquired paleomagnetic results. The intricate polarity pattern of the Câmpia Turzii section, however, points out that magnetostratigraphic interpretation in iron-sulfide dominated sediments remains delicate and great care has to be taken.
7. Biostratigraphy

7.1. Methods

All sections were investigated in detail for micropaleontological analyses primarily focusing on foraminifera and ostracods. Sediment samples of approximately 1 kg were processed by standard micropaleontological methods, sieved on 63 μm mesh and hand-picked under the microscope. A detailed description of the micropaleontological assemblage characteristic for each section or part thereof is provided in the supplementary material.

Each section is attributed to one or more particular biozones (Fig. 3). The biostratigraphic zonation of the Badenian and Sarmatian with foraminiferal assemblages is based on Popescu (2000), Filipescu and Silye (2008), and Beldean et al. (2010). The biostratigraphic zonation for the Pannonian is based on ostracods and follows (Jiříček, 1975, 1985), who referred to the standard Pannonian Biozones of Papp (1951) and Papp et al. (1985).

During the early Badenian the Paratethys and Mediterranean were well connected (Rögl, 1999; Popov et al., 2004) and biostratigraphic events can be correlated. Endemism hampers direct correlation of the middle Badenian, Sarmatian and Pannonian sections. Moreover, relatively deep water environments characterized the larger part of the Transylvanian Basin at that time. Most of the specifically designed biostratigraphic zonations for the Middle Miocene of the Central Paratethys rely on shallow water benthic taxa (e.g. Görög, 1992). They are therefore not always suitable for the deep-water sediments of the Transylvanian Basin (Filipescu and Silye, 2008). This sometimes complicates attribution of the sampled sections to a specific biozone. Fitting the ostracod assemblages from the Transylvanian Basin in the chronostratigraphic framework of the Pannonian stage is rather difficult because of pertinent differences regarding the Pannonian ostracod biozonation between authors (Pokorný, 1944; Brestenská, 1961; Krstić, 1973; Jiříček, 1975, 1985; Krstić, 1985; Rundić, 1997, 2006).

Assignment of the late Badenian to Pannonian sections to a specific biozone is for all these reasons often tentative.

7.2. Results

Three successive biozones can be identified in the Ciceu Giurgești section (Fig. 8). The assemblage of the lowermost part of the section is dominated by biseriolar planktonic foraminifera of the Early Miocene *Streptochilus pristinum* Biozone (Beldean et al., 2010). The second part of the outcrop, which starts with a conglomerate bed with clasts up to 15 cm in diameter, contains foraminifera of the early Badenian *Praeorbulina glomerosa* Biozone. Samples from the remaining part of the section belong to the *Orbulina suturalis* Biozone.

Nannoplankton data by Mészáros and Şuraru (1991) indicate that the Dej Tuff in the Pâglișa area pertains to the NN5 calcareous nannoplankton zone. The foraminiferal assemblage identified at Pâglișa belongs to the interval characterized by the *S. pristinum* to *O. suturalis* biozones.

The Făgăraș outcrop exposes sands and marls that belong to the middle to upper part of the late Badenian *Velapertina* Biozone. SEM investigations of samples from the Făgăraș 2 outcrop revealed frequent microperforate globigerinids belonging to the genera *Tenuitella* and *Tenuitellinata* assigned to the top Badenian *Tenuitellinata* Biozone (Filipescu and Silye, 2008).

The Faţăraş outcrop exposes sands and marls that belong to the middle to upper part of the late Badenian Velapertina Biozone. SEM investigations of samples from the Făţăraş 2 outcrop revealed frequent microperforate globigerinids belonging to the genera *Tenuitella* and *Tenuitellinata* assigned to the top Badenian *Tenuitellinata* Biozone (Filipescu and Silye, 2008).

The faunal assemblage of the marls of the Apahida section is dominated by small-sized foraminifera which are mostly in their juvenile stage. Specimens of *Tenuitellids* occur together with *Anomalinoides dividens*. The former are very common planktonics in the late Badenian and rare in the lowermost Sarmatian, whereas a bloom of latter marks the lowermost Sarmatian. This situation suggests a transitional zone positioned probably just below the Badenian–Sarmatian boundary. A similar situation occurs at Unguraş where the micropaleontological assemblage included several *Tenuitellinata* specimens and one specimen of *A. dividens*. The Gherla section clearly belongs to the lower
transgressive part of the Sarmatian corresponding to the A. dividens Biozone. Dilution of the assemblage in combination with reworking, and a biozonation based on benthic foraminifera which strongly depend on the local environment (see Piller et al., 2007) makes biozonation of the Hădăreni, Ghiroși, Câmpia Turzii 2 and Turda sections difficult. The diluted micropaleontological assemblage from the Hădăreni outcrop contains a few specimens of small rotaliids (Elphidium spp., Nonion spp.,) and miliolids (Quinqueloculina spp., Sinuloculina spp., Varidentella spp.). These characterize the lower half of the Sarmatian, but the exact biozone remains unclear. The Câmpia Turzii section exposes 150 m of clays, marls, and sandstones that belong to the A. dividens Biozone. The Câmpia Turzii 2 section, situated stratigraphically 250 m above the Câmpia Turzii section, exposes some 30 m of marls and subordinate sandstones with on top of it an 8 m thick tuff bed, named the Câmpia Turzii Tuff here. Biostratigraphic samples from this section contain frequent A. sarmatica in combination with rare A. dividens and most likely belongs to the Articulina Zone (middle-upper Volynnian), in which Anomalinaoides is rare but still occurs. The 150 m of marls and sandstones overlying the Turda Tuff might belong to the Elphidium reginum Biozone.

The biostratigraphy of five successive upper Sarmatian and lower Pannonian outcrops at Oarba de Mureș was studied in detail by Sztanó et al. (2005), Vasileiv et al. (2010a), and Filipescu et al. (2011). Vasileiv et al. (2010a) attribute the Oarba Beta section to the Poroconion aragviensis Biozone. Renewed inspection of the biostratigraphic samples, however, pointed out that the whole section contains species of Elphidium, but no specimens of Dogtelina sarmatica were encountered. This demonstrates that it is difficult to apply the zonations established in shallow marine settings to the deeper areas. We thus refrain from exact biozone determination for the Oarba Beta section. The larger part of outcrop A belongs to the upper Sarmatian Poroconion aragviensis Biozone. Samples from the top part of outcrop A contain typical Pannonian deep-water ostracods and thus Filipescu et al. (2011) place the Sarmatian–Pannonian boundary 2.3 m below the top of the section. This is in good agreement with the results of Sütő and Szegö (2008), who found massive numbers of Meceksia ultima 3.4 m below the top and the index taxon for the base of the Pannonian (Spiniferites bentonii pannonicus) 1.4 m below the top. Outcrops B, C and D are Pannonian in age and contain a rich ostracod association specific for low salinities. Sztanó et al. (2005) identified several mollusks indicative of the Lymnocardium propeunctum Zone of the early Pannonian in the lower part of outcrop B.

The micropaleontological assemblage of the Viforoasa and Șoimis sections is dominated by Pannonian ostracods. The larger part of the ostracod assemblages can be attributed to Zone 5 (Amlocypris abscessa Zone) and the lower part of Zone 6 (Hemicytheria croatica Zone) of the basal part of the upper Pannonian (lower Serbian) as defined by Krsčić (1973, 1985) and mentioned by Rundić (1997, 2006). They can also be correlated with Zone D and the first part of Zone E (Subzone E1) according to Jiříček (1975, 1985), who referred to the standard Pannonian biozones of (Papp, 1951). The lower part of Viforoasa section seems to be older than the Șoimis section and might correlate to the uppermost Slovenian (Propontiella candeo Zone) (Krsčić, 1973, 1985). The Târnăveni, Mediaș and Sighetșoara sections were exclusively sampled for rotation research and only few biostratigraphic samples were taken for age determination.

![Fig. 11. PDF, present day field direction; Dexp, declination expected based on the 10 Ma reference pole for Europe (Torsvik et al., 2008). Observed paleomagnetic declination for each section and uncertainty thereof according to results in Table 2. Section ages follow the correlation in Figs. 6, 8 and 9. Average directions for the Badenian to Pannonian outcrops. Rotation: vertical axis rotation with 95% confidence calculated applying (n) the number of sites used to calculate mean direction, (pol) the polarity of the site. Reference pole: age, latitude and longitude, and A 95 (95% confidence) according to Torsvik et al. (2008). Rotation: vertical axis rotation with 95% confidence limit (rotation and flattening: derived from observed direction minus expected direction at locality calculated from reference pole).

### Table 2

<table>
<thead>
<tr>
<th>Section</th>
<th>Site location</th>
<th>Observed direction</th>
<th>Reference pole</th>
<th>Expected I</th>
<th>Expected D</th>
<th>Rotation</th>
<th>Flattening</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mediș</td>
<td>46.17 24.36</td>
<td>54.4 19.0 4.1 141 4.9 10</td>
<td>10 87.2 125.0 2.5</td>
<td>44.4 63.9 ± 2.0</td>
<td>3.9 ± 3.6</td>
<td>15.1 ± 0.3</td>
<td>9.3 ± 3.6</td>
</tr>
<tr>
<td>Sighetșoara</td>
<td>46.24 24.81 58.4 10.8 59 8.9 9</td>
<td>10 87.2 125.0 2.5</td>
<td>44.3 64.0 ± 2.0</td>
<td>3.9 ± 3.6</td>
<td>8.3 ± 0.8</td>
<td>5.6 ± 5.7</td>
<td></td>
</tr>
<tr>
<td>oimuș</td>
<td>46.36 24.96 51.4 6.3 36 78 3.6 21</td>
<td>10 87.2 125.0 2.5</td>
<td>44.2 64.1 ± 2.0</td>
<td>4.0 ± 3.6</td>
<td>2.3 ± 5.4</td>
<td>12.7 ± 3.3</td>
<td></td>
</tr>
<tr>
<td>Viforoasa</td>
<td>46.43 24.81 53.2 10.4 27 84 2.9 34</td>
<td>10 87.2 125.0 2.5</td>
<td>44.1 64.1 ± 2.0</td>
<td>4.0 ± 3.6</td>
<td>4.6 ± 4.6</td>
<td>10.9 ± 2.7</td>
<td></td>
</tr>
<tr>
<td>Târnăveni</td>
<td>46.34 24.30 41.0 14.2 7.8 27 6.9 14</td>
<td>10 87.2 125.0 2.5</td>
<td>44.3 64.0 ± 2.0</td>
<td>3.9 ± 3.6</td>
<td>10.3 ± 0.8</td>
<td>23.0 ± 6.4</td>
<td></td>
</tr>
<tr>
<td>Orba D</td>
<td>46.46 24.28 66.8 6.9 39 114 5.9 13</td>
<td>10 87.2 125.0 2.5</td>
<td>44.1 64.1 ± 2.0</td>
<td>4.0 ± 3.6</td>
<td>2.9 ± 8.5</td>
<td>– 2.7 ± 3.5</td>
<td></td>
</tr>
<tr>
<td>Orba B</td>
<td>46.46 24.29 58.5 355.3 7.2 61 9.2 8</td>
<td>10 87.2 125.0 2.5</td>
<td>44.1 64.1 ± 2.0</td>
<td>4.0 ± 3.6</td>
<td>8.7 ± 11.5</td>
<td>5.6 ± 6.0</td>
<td></td>
</tr>
<tr>
<td>Average</td>
<td>46.25 24.50 53.4 10.1 20 47 2.1 106</td>
<td>10 87.2 125.0 2.5</td>
<td>44.3 64.3 ± 2.0</td>
<td>4.0 ± 3.6</td>
<td>6.2 ± 18.9</td>
<td>106.2 ± 18.9</td>
<td></td>
</tr>
<tr>
<td>Turda</td>
<td>46.57 23.83 46.1 4.2 42 48 2.5 28</td>
<td>10 87.2 125.0 2.5</td>
<td>44.0 64.2 ± 2.0</td>
<td>4.0 ± 3.6</td>
<td>17.2 ± 5.9</td>
<td>181.1 ± 3.9</td>
<td></td>
</tr>
<tr>
<td>Apașida</td>
<td>46.82 23.76 52.6 35.1 4.9 155 5.9 7</td>
<td>10 87.2 125.0 2.5</td>
<td>43.8 64.4 ± 2.0</td>
<td>4.0 ± 3.6</td>
<td>11.1 ± 7.1</td>
<td>11.8 ± 4.2</td>
<td></td>
</tr>
<tr>
<td>Unguraș</td>
<td>47.09 24.09 57.2 0.3 58 70 7.7 10</td>
<td>10 87.2 125.0 2.5</td>
<td>43.5 64.6 ± 2.0</td>
<td>4.0 ± 3.6</td>
<td>3.7 ± 9.1</td>
<td>7.4 ± 4.9</td>
<td></td>
</tr>
<tr>
<td>Făgărăș</td>
<td>45.87 25.00 51.8 24.8 120 51 15.0 4</td>
<td>10 87.2 125.0 2.5</td>
<td>44.7 67.3 ± 2.0</td>
<td>3.9 ± 3.6</td>
<td>20.9 ± 16.0</td>
<td>119.9 ± 7.9</td>
<td></td>
</tr>
<tr>
<td>Păgîlța</td>
<td>47.01 23.66 54.3 20.1 5.3 68 6.5 12</td>
<td>10 87.2 125.0 2.5</td>
<td>43.6 64.5 ± 2.0</td>
<td>4.0 ± 3.6</td>
<td>16.1 ± 7.8</td>
<td>102.4 ± 4.5</td>
<td></td>
</tr>
<tr>
<td>Average</td>
<td>46.25 24.50 51.5 30.0 3.3 33 4.0 57</td>
<td>10 87.2 125.0 2.5</td>
<td>44.3 64.0 ± 2.0</td>
<td>3.9 ± 3.6</td>
<td>26.1 ± 5.1</td>
<td>125.3 ± 3.1</td>
<td></td>
</tr>
</tbody>
</table>
from a single stratigraphic level in each of the quarries. The faunal assemblage indicates a Pannonian age.

8. A new chronostratigraphy for the Middle to Upper Miocene of the Transylvanian Basin

Our new radio-isotopic, magnetostratigraphic and biostratigraphic results allow, in combination with the high resolution seismic correlation of surface outcrops, the establishment of a new chronostratigraphy for the Middle to Upper Miocene of the Transylvanian Basin.

8.1. Late Karpatian and early Badenian: the S. pristimum, P. glomerosa and O. suturalis biozones and deposition of the Dej Tuff complex

The Ciceu Giurgeşti, Dej and Pâglişa outcrops (Fig. 8) reflect the early Badenian evolution of the last megasequence deposited in the Transylvanian Basin. The lower part of the Ciceu Giurgeşti, Dej, and Pâglişa sections comprise the sediments that accumulated before the onset of volcanic activity related to the Dej Tuff complex. The presence of the late Karpatian and two Badenian biozones in an interval of only 24 m corroborates extremely low sedimentation rates. The transition from Karpatian to Badenian (i.e. the Early to Middle Miocene transition) occurs after 7 m (measured from the base of the section) and before 11.5 m where P. glomerosa is first registered. In the Mediterranean the FO of P. glomerosa was dated to occur at 15.97 Ma (Lourens et al., 2004b). A second biostratigraphically correlatable horizon (Fig. 8) is the FO of O. suturalis, indicative for the start of the O. suturalis Biozone. It was dated to occur at 14.56 Ma in the Western Mediterranean (Abdul Aziz et al., 2008).

The 14.38 ± 0.06 Ma 40Ar/39Ar age for the Dej Tuff from Ciceu Giurgeşti provides the third tie-point to the global timescale. Deposition of this tuff slightly postdates the FO of O. suturalis (Fig. 8) and provides an independent minimum age constraint on the first occurrence of O. suturalis in the Central Paratethys. Since this dated tuff horizon is one of the basal horizons of the Dej Tuff complex, the acquired age should be indicative of the onset of related volcanism. The statistically indistinguishable 14.37 ± 0.06 Ma age determined for the Dej tuff at Dej provides additional support for this conclusion. This age contrasts strongly with the much older age of between 15.4 and 15 Ma previously attributed to the Dej Tuff based on K-Ar and fission track measurements (Szakács et al., 2000). Since the potassium-argon method is restricted to whole rock measurements, the strong reworked component that our experiments demonstrate to be present in the Dej Tuff, might have induced seemingly higher ages. Since multiple single fusion 40Ar/39Ar experiments are performed on only a small selected fraction of minerals in the rock, detection of, and correction for reworking is possible. We thus conclude that the here calculated 14.37 ± 0.06 Ma and 14.38 ± 0.06 Ma crystallization ages provide more reliable ages for the Dej Tuff.

The part of the tuff complex sampled at Pâglişa also pertains to the O. suturalis Biozone and is exclusively of normal polarity. It was thus deposited during chron C5A.Dn. Our results imply that accumulation of the over 80 m thick volcanoclastic complex took much shorter than previously thought. Tuffs and tuffites belonging to the Dej Tuff complex disappear from the sedimentary record before the onset of the BSC at 13.82 Ma. Tuff deposition can thus have lasted a maximum of 0.6 Ma.

8.2. Late Badenian and Sarmatian biostratigraphy

The base of Velupertina Biozone corresponds to the end of the salinity crisis and is consequently 13.36 Ma. The boundary between the Velupertina and Tenuitellinata Biozones is located in the Fâgărăș 2 section. Our results also provide new age constraints on the Badenian–Sarmatian boundary and the Velupertina s.l., Tenuitellinata and A. dividens Biozones. The upper Badenian Fâgărăș 2 outcrop, described by Filipescu and Silye (2008), falls into the latest Badenian Tenuitellinata Biozone. The micropaleontological assemblage of the marls of the Apahida Tuff reflects the very top part of the same biozone and suggests it is just below the base of the Sarmatian. The Gherla section belongs to the lower Sarmatian Anomalinooides Biozone and therefore provides an upper limit for the position of the Badenian–Sarmatian boundary, which is consequently positioned between 2.73 and 2.93 km, and is dated between 12.70 and 12.83 Ma (Fig. 6). The observed micropaleontological assemblages suggest that the boundary is stratigraphically closer to the Apahida than to the Gherla section and is thus likely close to 12.8 Ma old (the Apahida Tuff is traditionally placed at the boundary — Vancea, 1960). This age for the transition from Badenian to Sarmatian in the Transylvanian Basin is very similar to the age attributed to this transition along the western margin of the Central Paratethys by Harzhauser and Piller (2004) on the basis of sequence stratigraphic arguments, and in good agreement with the down-well magnetostratigraphic results of Paulissen et al. (2011) for the Vienna Basin. It is remarkably younger than the 13.32 Ma age found by Lirer et al. (2009) through cyclostratigraphic calibration of well-data in the same basin. Many wells in the Vienna Basin, including the Spannberg-21 studied by Paulissen et al. (2011) are, however, known to have a hiatus spanning the interval of the Badenian–Sarmatian boundary. A hiatus in the Eichorn-1 well cannot be excluded (Lirer et al., 2009), which may explain the discrepancy between our results and those of Lirer et al. (2009).

The Gherla and Câmpia Turzii 1 outcrops belong to the A. dividens Biozone (Fig. 6). The transition between the A. dividens, Varidentella reussi, and A. sarmatica biozones is hard to pinpoint since it occurs gradually. A. dividens continues to be present in the A. sarmatica Biozone although it becomes less abundant. Some taxa characteristic for the Articulina Zone moreover already occur at the top of the A. dividens zone. However, at the level of the Câmpia Turzii 2 and Hâdâreni outcrops the assemblage becomes more diverse than in the underlying strata. The abundance of Anomalinooides decreases and the assemblage is enriched in specimens of miliolids (Articulina, Varidentella, Sinuloculina) and rotaliids (Elphidium spp.), most likely in response to environmental changes in the basin. This suggests it is adequate to place the zone-boundary at this level. The A. dividens Biozone should, in our opinion, therefore end at the stratigraphic level of the Hâdâreni Tuff. The Câmpia Turzii Tuff is the first tuff level above the Câmpia Turzii 1 section, which pertains to the A. dividens Biozone. The faunal assemblage is, in analogy with the Hâdâreni Tuff, enriched at the level of the Câmpia Turzii Tuff. This suggests that the Hâdâreni and Câmpia Turzii Tuffs represent a single tuff level that coincides with the top of the A. dividens Biozone. The top of the Anomalinooides is consequently around 12.4 Ma old. Differentiation between biozones in the investigated deep water sections becomes more problematic above this level and we thus refrain from age estimates for the A. sarmatica, Dogteliina sarmatica and the base of the P. aragviensis biozones, which were separated using benthiic taxa, which strongly depend on the local environment.

8.3. The Sarmatian–Pannonian Boundary

In the Oarba de Mure section, situated near the centre of the Transylvanian Basin, the Sarmatian–Pannonian boundary is located 46 m above the 11.62 ± 0.04 Ma old Oarba Tuff. If the average rate of deposition of 1.5 m/kyr calculated for the upper Badenian to Sarmatian infill of the basin were applicable to the Oarba de Mure A section, then the Sarmatian–Pannonian boundary would be around 11.59 Ma old. This would be in good correspondence with the results of Paulissen et al. (2011) and ter Borgh et al. (2013-this volume), whose magnetostratigraphic correlations show that the Sarmatian–Pannonian boundary is also around 11.6 Ma old in the Vienna Basin and the southern Pannonian Basin respectively. Based on calibration of the magnetostratigraphic pattern of outcrop A–D of the Oarba de Mure
section to the ATNTS, Vasilev et al. (2010a) have nevertheless determined an age of 11.3 ± 0.1 Ma for the Sarmatian–Pannonian boundary in the Transylvanian Basin. This implies significantly lower sedimentation rates for the middle and upper part of the Oarbe de Mure A outcrop.

Biostratigraphic evidence for a younger Sarmatian–Pannonian boundary in Transylvania is unequivocal. The basal Pannonian ostracod zone A is missing at Oarbe de Mure (Filipescu et al., 2011), suggesting that the marine conditions characteristic for the Sarmatian stage persisted longer in the Transylvanian Basin than in the Pannonian Basin and inhibited colonization by the fresh-water ostracods that characterize the base of the Pannonian there. This would provide additional support for a younger (i.e. 11.3 Ma) age for the Sarmatian–Pannonian boundary in Transylvania, were it not that Dino-flagellate and mollusc-records (Sztanó et al., 2005; Sütő and Szegó, 2008) characteristic for the basal part of the Pannonian are present.

We here choose to adopt the magnetostratigraphic correlation of the Oarbe de Mure sections by Vasilev et al. (2010a) and attribute an age of 11.3 Ma to the Sarmatian–Pannonian boundary. This age implies that the Transylvanian Basin became isolated from the Eastern Paratethys 0.3 Ma after the Pannonian Basin became isolated, as discussed in detail by ter Borgh et al. (2013–this volume).

8.4. The Pannonian

Marked differences exist regarding the Pannonian ostracod biozonation between different authors (Pokorný, 1944; Brestenská, 1961; Krsť, 1973; Jiříček, 1975, 1985; Krsť, 1985; Rundić, 1997, 2006). Moreover, ostracod assemblages are commonly environment sensitive. Pannonian ostracod associations from the central part of the Transylvanian Basin, including the Viforoasa and Šomuș sections, reflect a more deep, basinal and possibly fresher water environment (Krêszek and Filipescu, 2005) than in the other parts of Pannonian Basin where the ostracod bioclines were defined. The micropaleontological assemblage of the Viforoasa and Šomuș sections can, however, be correlated with the Middle Pannonian Zone D and the lower part of Zone E (Subzone E1) according to jiříček (1975, 1985). According to our magnetostratigraphic correlation of these outcrops and their position in the synthetic seismic stratigraphic column (Fig. 9), they correspond to the time-interval between 10.6 and 9.9 Ma. This is in good agreement with results by Harzhauser et al. (2004) who attribute an age of 10.56 Ma to the base of Zone D. These authors do, on the other hand, attribute an age of 10.36 Ma to the top of Zone E1. Adoption of a 10.36 Ma maximum age for the Šomuș section would lead to an even larger difference in sedimentation rates for the Sarmatian and Pannonian intervals in the Transylvanian Basin. Moreover, the polarity of the uppermost part of this section warrants correlation to a reversed Chron. We therefore prefer correlation of the reversal in the Šomuș section to the transition from C5n.2n to C5n.1r and infer an age of 10.0 Ma for this level. This is in good agreement with the age of faunal assemblages similar to those of Matenco et al. (1997, 2006). More-
difference in exhumation can be related to a larger amount of shortening in the Eastern Carpathian areas adjacent to the northern part of the Transylvanian Basin and therefore explain the observed 6° Pannonian to post-Pannonian rotations.

Our results are in agreement with the near 30° CW rotation recorded by sediments directly underlying the Sarmatian strata of the Southern Carpathians (Dupont-Nivet et al., 2005). Sediments younger than 6.2 Ma are, according to these authors, on the other hand, not rotated. These findings demonstrate that Tisza–Dacia (including the Transylvanian Basin and surrounding orogens) rotated as a rigid block around the Moesian indenter during the Middle to Late Miocene, without the requirement of internal differential strain partitioning structures.

The observed CW rotation of the Transylvanian Basin is driven by the eastward roll-back of the European slab during its subduction below the Eastern Carpathians (Royden, 1988) and provides new constraints on the concurrent emplacement of the Tisza–Dacia upper plate into the Carpathian embayment.

11. Conclusions

Our new radio-isotopic, magnetostratigraphic and biostratigraphic results provide a new chronostratigraphy for the Middle to Upper Miocene of the Transylvanian Basin when stratigraphically calibrated to a synthetic seismic section in the basin center through subsurface tracing. This improves insight in the basin’s evolution and moreover provides new time constraints on the Middle Miocene Central Paratethys regional stages. The magnitude and timing of paleomagnetically determined tectonic rotations furthermore give new constraints on the emplacement of the Tisza–Dacia plate into the Carpathian embayment.

In the early Badenian (16 to 13.82 Ma), when marine connections between the Central Paratethys and the Mediterranean still existed, up to 100 m of relatively deep water sediments accumulated under a slight extensional regime (Křezek et al., 2010). Intense volcanism led to the deposition of the Dej Tuff complex, an important lithostratigraphic marker for the lower Badenian. The onset of Dej Tuff volcanism is constrained between the first occurrence (FO) of O. suturealis at 14.56 Ma (Abdul Aziz et al., 2008) and 14.38 ± 0.06 Ma, based on new 40Ar/39Ar ages for two of the lowermost tuff levels. Tuff accumulation ceased before the onset of the Badenian Salinity Crisis (BSC) at 13.82 Ma (de Leeuw et al., 2010). The BSC was triggered by a glacio-eustatic restriction of the connection between the Central Paratethys and the Mediterranean (de Leeuw et al., 2010) and lead to the accumulation of up to 300 m of salt in the central part of the Transylvanian Basin. The establishment of a marine connection to the Eastern Paratethys at the onset of the late Badenian ends salt deposition restores normal marine conditions. The strongly endemic nature of late Badenian faunal assemblages suggest the marine connection to the Mediterranean ceased to exist (see Rögl, 1999; Piller et al., 2007). Accumulation rates strongly increased in comparison with the early Badenian, and a marked W–E thickening of the upper Badenian in the basin suggests this relates to an increase in tectonic activity in the Eastern Carpathians. The transition from Badenian to Sarmatian is marked by a strong faunal turnover (Harzhauser and Piller, 2007). Most foraminifera, apart from some Tenuitellina and Tenuitellinata (see Filipescu and Silye, 2008), and over 500 gastropod species disappear. We date the Badenian–Sarmatian Boundary in the Transylvanian Basin at 12.80 ± 0.05 Ma. The upper Badenian consequently lasts from 13.36 to 12.80 ± 0.05 Ma. At 12.4 Ma, a second pulse of intense volcanic activity occurred, leading to deposition of the Chişirş, Hădăreni, Turda and Ciâmpia Turzi tuffs that are collectively called the middle Sarmatian tuff complex. The tuffs accumulated in a short time as indicated by their statistically indistinguishable ages of 12.38 ± 0.04 Ma, 12.37 ± 0.05 Ma, 12.37 ± 0.04 Ma, and 12.35 ± 0.04 Ma. The interval between the onset of the Sarmatian and the middle Sarmatian tuff complex belongs to the Anomalinoides dividers Biozone which accordingly covers the 12.80 ± 0.05 Ma to 12.4 Ma time interval. The basin is struck by a second major environmental change with a heavy impact on its biota at the Sarmatian–Pannonian boundary, in Transylvania dated at 11.3 Ma by Vasiliy et al. (2010a). The complete disappearance of foraminifera from the faunal record and a subsequent proliferation of ostracods mark a transition to freshwater conditions registered throughout the whole Central Paratethys. This was triggered by tectonic uplift of the Eastern and Southern Carpathians which isolated the Central Paratethys from the Eastern Paratethys (Křezek et al., 2010). Concomitant enhanced uplift in the Carpathians is corroborated by the timing of paleomagnetically determined tectonic rotation of the Tisza–Dacia block. The average rotation of sampled Badenian and Sarmatian sites is 26.1 ± 5.1°. The average rotation of the Pannonian sites is 62.2 ± 3.9°. This indicates the Transylvanian Basin rotated approximately 20° CW from the late Sarmatian to early Pannonian. This rotation was accommodated sinistral strike slip along the Drago Vodă–Bogdan Vodă (DVBV) fault system and differential shortening observed in the Eastern and Southern Carpathians. Magnetostratigraphic correlation of the investigated Pannonian sections, which belong to middle Pannonian Zone D, and the lower part of Zone E (Subzone E1) of Jiřiček (1975, 1985), to the ATNIS (Lourens et al., 2004a) indicates that these cover the 10.6–9.9 Ma time interval. This suggests an 8.4 Ma age for the uppermost Pannonian deposits in the depository of the Transylvanian Basin. Our new chronostratigraphic results confirm that during the late Badenian and Sarmatian, coincident with intensive nappe stacking in the neighboring East Carpathians (Matenco et al., 2007), sediment accumulation rates in the center of the Transylvanian Basin were much higher than during the Pannonian, when a gradual change from subsidence to uplift occurred (Matenco et al., 2010). In the Eastern Carpathians, deformation at this time migrated towards the interior of the thrust wedge. The resulting antiformal geometries were significantly larger in the NE Carpathians than in the SE Carpathians (Matenco et al., 2010) which can explain the observed 6° Pannonian to post-Pannonian rotation.

Supplementary data to this article can be found online at http://dx.doi.org/10.1016/j.gloplacha.2012.04.008.

Acknowledgments

We would like to express great appreciation for the commitment of Andrei Bricas, Arnoud Slootman, and Jaap Verbaas who have contributed to the conceptual and practical advancement of this research. This study would not have been possible without the benevolence of multiple brick factories that have provided access to their quarries. We thank Mr. Vasilie for his hospitality, Mr. Moldovan for logistic support, Roel van ELSA van help for mineral separation, Jan Wijnbrans, Guillaume Dupont-Nivet, Douwe van Hinsbergen for discussion, and Cor Langereis and four anonymous reviewers for critically reviewing the manuscript. This study was supported by the Netherlands Research Centre for Integrated Solid Earth Sciences (ISES) and by the Netherlands Organization for Scientific Research (NWO/ALW).

References


Seghedi, I., Szakács, A., 1991. The Dej Tuff

Popov, S.V., et al., 2006. Late Miocene to Pliocene palaeogeography of the Paratethys


