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Towards better correlation of the Central Paratethys regional time scale with the standard geological time scale of the Miocene Epoch

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Abstract: Depositional sequences originating in semi-enclosed basins with endemic biota, partly or completely isolated from the open ocean, frequently do not allow biostratigraphic correlations with the standard geological time scale (GTS). The Miocene stages of the Central Paratethys represent regional chronostratigraphic units that were defined in type sections mostly on the basis of biostratigraphic criteria. The lack of accurate dating makes correlation within and between basins of this area and at global scales difficult. Although new geochronological estimates increasingly constrain the age of stage boundaries in the Paratethys, such estimates can be misleading if they do not account for diachronous boundaries between lithostratigraphic formations and for forward smearing of first appearances of index species (Signor-Lipps effect), and if they are extrapolated to whole basins. Here, we argue that (1) geochronological estimates of stage boundaries need to be based on sections with high completeness and high sediment accumulation rates, and (2) that the boundaries should preferentially correspond to conditions with sufficient marine connectivity between the Paratethys and the open ocean. The differences between the timing of origination of a given species in the source area and timing of its immigration to the Paratethys basins should be minimized during such intervals. Here, we draw attention to the definition of the Central Paratethys regional time scale, its modifications, and its present-day validity. We suggest that the regional time scale should be adjusted so that stage boundaries reflect local and regional geodynamic processes as well as the opening and closing of marine gateways. The role of eustatic sea level changes and geodynamic processes in determining the gateway formation needs to be rigorously evaluated with geochronological data and spatially-explicit biostratigraphic data so that their effects can be disentangled.

Keywords: Neogene, Central Paratethys, regional and standard time scales, sea-level changes, geodynamics.

Introduction

Since the definition of the Paratethys Sea realm nearly a century ago (Laskarev 1924) the correlation between the regional Paratethys stages and the standard (Mediterranean) stages remains poorly documented and validated (Magyar et al. 1999; Brzobohatý et al. 2003; Kováč et al. 2007; Hohenegger et al. 2014; Sant et al. 2017b). The geochronological and biostratigraphic definition of stage boundaries is one of the critical problems that limits understanding of the climatic, oceanographic, and ecosystem history of the Paratethys (Rögl et al. 2003; Piller et al. 2007; de Leeuw et al. 2013; Silye & Filipescu 2016). Although the regional time scale needs to be supported by absolute dating, geochronological point-based data used without sufficient knowledge of the local lithostratigraphic nomenclature and not accounting for forward and backward smearing of first and last appearances (Signor & Lipps 1982), sequence- and bio-stratigraphic correlations within the Paratethys and between the Paratethys and other regions can be misleading.

The Central Paratethys (CP) time scale was defined on the basis of lithostratigraphy of sedimentary sequences belonging to distinct transgressive–regressive cycles. Lithostratigraphic boundaries coincide with immigration or evolutionary events to some degree: the stages were defined on the basis of immigration of new taxa from other bio-provinces (Atlantic, Mediterranean, Indo–Pacific, and Eastern Paratethys) or by evolution of endemic biota. Initially, the stage boundaries were often represented by hiatuses or discordances that correspond to boundaries between lithostratigraphic formations (Cicha et al. 1967; Steininger & Seneš 1971; Papp et al. 1973, 1974, 1978, 1985; Báldi & Seneš 1975; Stevanović et al. 1989). Later, definitions of regional stages have begun to prefer biostratigraphic criteria (Piller et al. 2007). However, first appearances of marine organisms in isolated basins tend to be affected by significant delays relative to their first appearance in open ocean environments owing to geographical and environmental constraints (e.g., Kennett et al. 1985; Holcová et al. 2015; Jones & Murray 2017; Sant et al., 2017a). Therefore, the accuracy of correlations of stage boundaries based on palaeontological criteria in type sections, without any geochronological verification and without understanding of temporal changes in biogeographic distribution of index species, is unclear.

The most complete contemporary regional time scale presented by Krijgsman & Piller (2012), primarily based on biostratigraphic criteria, is still in use (Fig. 1). New point-based geochronological estimates were measured in order to determine the age of the regional stage boundaries and to allow broad-scale correlations (Vasiliev et al. 2010; de Leeuw et al. 2013; Roetzel et al. 2014; Zuschin et al. 2014; Palcu et al. 2015; Sant et al. 2017a). The essential condition of such estimates is that they should be derived from complete sedimentary sequences, with a relatively high sediment accumulation rate and conditions allowing preservation of index species. However, these two conditions are rarely met in the CP.

Extrapolating the age of a stage boundary gained by geochronological methods from a single site to a whole basin can be a major problem in intra-basinal correlations based on sedimentological criteria because depositional facies (e.g., deltas) and benthic biofacies are frequently diachronous, and the first and last appearances of planktic species can also be diachronous due to migration, environmental and sampling constraints (MacLeod 1996). The point-based geochronological data should thus be supported by biostratigraphic and biogeographic distribution of planktic organisms, with their first (FO) or last (LO) occurrences and vice versa. However, the usage of Atlantic biozonations of planktic foraminifers (Berggren et al. 1995) and calcareous nannoplankton (Martini 1971) is limited in the Mediterranean and adjacent CP realm because planktic foraminifers tend to be rare in isolated basins and some geochronological data indicate age offsets between times of their appearance in different basins (Iaccarino & Salvatorini 1982; Iaccarino 1985; Mărunţeanu 1999; Andrejeva Grigorovič et al. 2001; Turco et al. 2002, 2017; Lirer & Iaccarino 2005; Iaccarino et al. 2011; Paulissen et al. 2011; Gonera 2013; Bartol et al. 2014; Di Stefano et al. 2015; Holcová et al. 2015; Sant et al. 2017b). Therefore, it is necessary to date the FO of planktic taxa in the CP by geochronological methods and ensure that the extent of forward smearing will be assessed with taphonomic, palaeoecological, and palaeobiogeographic criteria. For example, if sample sizes are small, palaeoenvironmental conditions do not match preferences of index species closely, and/or if circulation barriers exist between provinces, the geochronological ages defined on the basis of FO in a given section are likely to be younger than the real timing of origination of a given species.

The calcareous nannofossil zonation (*sensu* Martini 1971; e.g., FO of *Helicosphaera ampliaperta* defines the base of the Eggenburgian (Burdigalian) in the NN2 Zone; LO of *Sphenolithus belemnos* defines the Ottnangian in the NN3 Zone, LO of *H. ampliaperta* defines the boundary between the NN4/NN5 zones, LO of *Sphenolithus heteromorphus* defines the termination of "Early Badenian" in the NN5 Zone, and FO of *Discoaster kugleri* occurs in the Sarmatian NN7 Zone), and planktic foraminiferal markers (*sensu* Piller et al. 2007; Filipescu & Silye 2008; *Catapsydrax* appears in the Ottnangian, *Trilobatus bisphericus* (=*Globigerinoides bisphericus*) appears in the late Karpatian, while *Praeorbulina glomerosa* and *Orbulina suturalis* appear in the Early Badenian) are used. The usage of benthic and/or endemic molluscs or foraminifera for the definition of regional stage boundaries can be unreliable (e.g., the Ottnangian/Karpatian boundary is marked by FO of *Uvigerina graciliformis*). Stratigraphic correlations based on the composition of benthic assemblages can be biased by diachronous occurrence of the benthic taxa temporally tracking their preferred environments, and can simply reflect an existence of environment that is optimal for a given taxon during a certain time span (e.g., delta, shelf, basin slope, basin floor).

Geochronological methods can accurately estimate the onset and duration of deposition of some specific sedimentary facies. This can be achieved by dating points in sections arrayed in vertical and horizontal transects across the basin — generally, multiple such point estimates are required. For example, Šujan et al. (2016) documented the diachronity of sedimentation of the Pannonian formations in the Upper Miocene infill of the Danube Basin. The spatial shift of facies types within a given depositional system was demonstrated on the basis of multiple point-based geochronological age estimates of sediments belonging to the same facies type (and lithostratigraphic formation) across the basin. The point-based ages showed that the sedimentation along a shelf-slope-basin transect lasted for more than 3 Ma, namely the time needed until the basin was filled up. These results thus documented the diachronity of sedimentation of the Pannonian formations, with a lower boundary equal to the base of the Pannonian regional stage. Therefore, the point-based age data can be reliably used in order to correlate sections within a basin, but also clearly show many inconsistencies in the correlation between several CP basins.

In addition to biostratigraphy, the CP time scale published by Piller et al. (2007) and Krijgsman & Piller (2012) also comprised correlations with the global sea-level curve, and the individual stage boundaries were correlated with the boundaries of the 3rd order cycles of the global sequence stratigraphy (after Haq et al. 1988; Hardenbol et al. 1998). However, the research carried out in semi-enclosed basins has shown that the global sea-level change is captured by the sedimentary record only to some degree (Kováč et al. 2004; Krézsek & Filipescu 2005; Strauss et al. 2006). The active tectonics and/or a huge amount of material input can intensify, reduce, or completely hide the signatures of the global sea-level changes (Kováč & Zlinská 1998; Kováč et al. 1998, 1999a,b, 2004; Hlavatá-Hudáčková et al. 2000; Kováč 2000; Catuneanu 2006). In addition, the 3rd order sequence stratigraphic cycles recorded in the CP respond not only to the effects of the Mediterranean, but also to the Eastern Paratethys water masses.

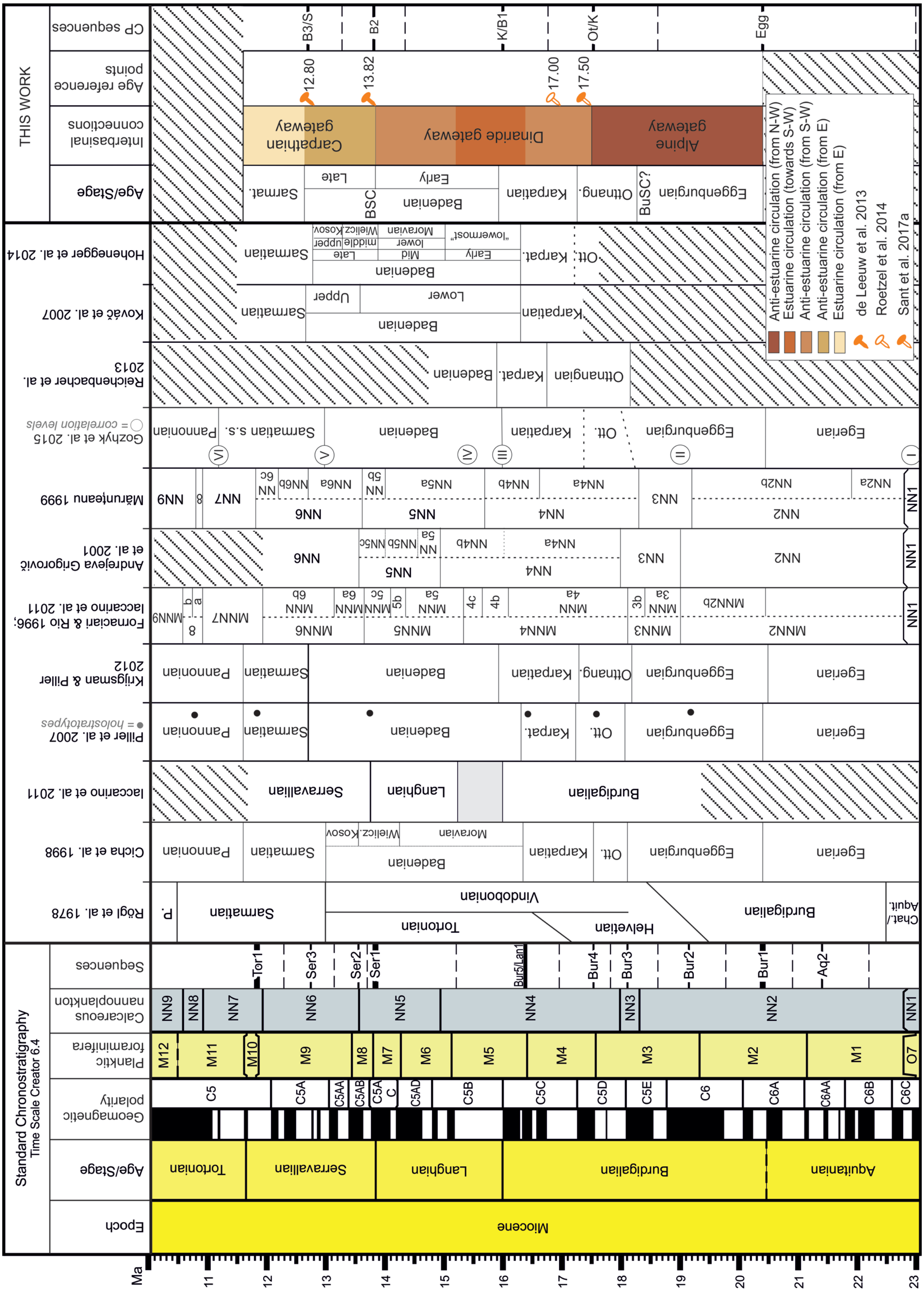

Fig. 1. Standard Neogene chronostratigraphy and biostratigraphy; selected regional Miocene time scales — an overview (for references see figure); proposed regional time scale and Central Paratethys marine gateways. Explanatory notes: hatched area — not studied; BuSC — Burdigalian Salinity Crisis.

Moreover, the relative sea-level curve of the Eastern Paratethys differs significantly from the global eustatic sea-level curve (Popov et al. 2010). Therefore, the impact of the Miocene global sea-level changes on sequence-stratigraphic architecture of basins in the CP should be re-evaluated.

In an inland sea, such as the CP, the global factors affected the palaeogeographical evolution only partially, while impacts of the local geodynamic processes were more critical (Kováč 2000; Kováč et al. 2016, 2017b; Sant et al. 2017b). The geodynamic development of the Alpine–Carpathian–Dinaride orogenic systems determined the distribution and extent of terrestrial and marine environments and significantly shaped the sequence architecture, palaeoclimate, and water masses circulation regime of the CP (e.g., Kováč et al. 2004, 2017a; Grunert et al. 2010, 2014; ter Borgh et al. 2013; Palcu et al. 2015). Intensity of marine currents and oceanic circulation patterns strongly impacts biogeographic distribution of plankton and benthos (e.g., Kennett et al. 1985; Peters et al. 2013; Holcová et al. 2015; Jones & Murray 2017; Kováč et al. 2017a; Sant et al., 2017a). The timing of faunal appearances thus principally depends on the opening of gateways towards the Mediterranean, or Eastern Paratethys as it is documented from the (sub)recent Mediterranean or Black Sea (e.g., Kouwenhoven & van der Zwaan 2006; Karami et al. 2011; Palcu et al. 2015; Kováč et al. 2017a; Sant et al. 2017b). Therefore, the onset of regional stages should correspond to conditions with a relatively high marine connectivity between the Mediterranean and the CP, or at least the connection with a substantially larger sea-covered area (Eastern Paratethys).

The present location of gateways that represent migration corridors for marine organisms between the Mediterranean, Central Paratethys and Eastern Paratethys realm, as well as the distribution of the individual CP basins does not correspond to their original position. The sedimentary fill of the Miocene basins forms part of fold and thrust belts, or is dissected by transform faults and the individual parts of basins were transported several hundreds of kilometres from their site of origin. These changes in the location and configuration of sedimentary basins were not taken into account in palaeogeographical reconstructions for more than decades (e.g., Hámor & Halmai 1988; Popov et al. 2004; Sant et al. 2017b).

The view that the Outer Carpathian thrust belt was shortened more than 150–200 km during the Miocene (e.g., Kováč et al. 2017b) can be used as an example. The wide marine realm in front of the moving orogenic wedge gradually shifted towards the European platform margins; basins on the top of the accretionary wedge were folded and thrust ahead (generally north- and east-ward). Basins on the platform margin (foredeep depocenters) were diachronously filled up (e.g., Meulenkamp et al. 1996). This shortening had a massive impact on the extent and distribution of marine and terrestrial environments. Similarly, in the orogenic hinterland system, the Upper Oligocene–Lower Miocene retro-arc basin was situated at least 200 km towards the southwest with respect to its recent position (e.g., Tari et al. 1992; Kováč et al. 2016, 2017b). The basin was later divided into two parts due to extrusion of the northern Pannonian crustal fragment from the zone between the Alps and Dinarides, and reached its present position in the late Early Miocene (e.g., Fodor et al. 1998; Kováč et al. 2016, 2017b). Therefore, geographic maps not accounting palaeogeographic shifts are misleading and cannot represent baselines for broader palaeogeographic reconstructions. To evaluate changes in the configurations of basins through time, an accurate palinspastic modelling based on interdisciplinary approach reflecting original position and extent of basins is needed.

The considerable problem of the regional CP time scale is that the individual stage boundaries are seldom supported by up-to-date geochronological data and by biostratigraphic data that would account for temporal changes in biogeographic distribution of index species, what makes correlation within individual basins of the CP area, as well as with the Mediterranean, troublesome. The use of regional stages without point-based geochronological age data and sufficient knowledge of local lithostratigraphic nomenclature, tectonics, and sequence stratigraphy can be misleading in interregional correlations at European scale. As we show below, the use of a standard time scale is more appropriate.

For example, in an inspiring paper by Sant et al. (2017b), the “Ottnangian Sea” at 18 Ma (see fig. 4A in Sant et al. 2017b) extends from the Alpine Foredeep (Molasse Basin) across the hinterland of the Central Western Carpathians (Novohrad–Nógrád Basin) to the area of the Eastern Slovakia Basin. However, this time slice should be referred to as the “early Burdigalian” because most of the Ottnangian strata are not formed by marine sediments in the Novohrad–Nógrád and Eastern Slovakia basins in the Western Carpathians. The Ottnangian sediments are represented by terrestrial deposits or by hiatuses in these basins (e.g., Vass et al. 1979; Rudinec 1989, 1990; Kováč et al. 1995; Vass 2002; Vass et al. 2007). We note that marine sediments of the age around 18 Ma are present in both basins, but they are assigned to the Eggenburgian stage (Vass et al. 1979, 2007; Vass 2002; Fordinál et al. 2014; Kováč et al. 2017a).

The “Karpatian Sea” with the age of ~16.5 Ma, and the “Badenian Sea” with the age of less than ~14 Ma depicted in figs. 4B and 4C (Sant et al. 2017b) represent a new period in the CP development, prior to the Badenian Salinity Crisis (BSC) and prior to the onset of the “Late Badenian Sea” transgression, respectively. However, the base of the Badenian regional stage is traditionally dated to ~16.4–16.3 Ma (e.g., Piller et al. 2007; Filipescu & Silye 2008; Hohenegger et al. 2014), whereas the “Karpatian Sea” in fig. 4B (Sant et al. 2017b) is dated to 16.5 Ma. Similarly, as in the previous case, the standard Miocene chronostratigraphic terminology should be used; this map captures remnants of the upper Burdigalian sediments deposited prior to the Langhian transgression.

Finally, the suggestion of Sant et al. (2017b) “*the establishment of the “Badenian Sea” (<15.2 Ma), triggered by subduction-related processes in the Pannonian and Carpathian domain, is significantly younger (by ~1 Myr) than usually* *estimated*" cannot be accepted. The onset of the NN5 Zone with *Orbulina suturalis* in the Early Badenian, coexisting with *Praeorbulina* ssp. around 15 Ma is well documented in most basins of the CP (e.g., Kováč et al. 2017a and references therein).

To conclude, the regional stage boundaries in the CP need to be dated by biostratigraphic approaches validated by geochronological methods, and the role of gateways should be evaluated as a function of local tectonics and not only as a function of broad-scale eustatic sea-level changes.

The definition of the Central Paratethys regional stages and their validity

The Miocene time scale applied for the CP area in the 19th century compiled Mediterranean and regional stages such as the Burdigalian and Helvetian for the Early Miocene, Tortonian for the Middle Miocene, and Levantian stage for the Pliocene (e.g., Mayer-Eimar 1858). In the 1970's, the currently used CP regional stages were defined in the series of books *Chronostratigraphie und Neostratotypen* (Cicha et al. 1967; Steininger & Seneš 1971; Papp et al. 1973, 1974, 1978, 1985; Báldi & Seneš 1975; Stevanović et al. 1989). However, the lower boundaries of these stages were not precisely geochronologically constrained, and in some cases the biostratigraphical definition of the stage boundaries was also insufficient. These stages are applied for the sedimentary record from the Alps across the Pannonian Basin System up to the Carpathians, Dinarides, and Balkans. The conversion from old to new stratigraphic nomenclature led to discrepancies in duration of the sedimentary record assigned to the same stage among different basins of the Eastern Alps and Western Carpathians. For example, the sediments formerly assigned to the Helvetian stage (Fig. 1) were partly correlated with the Ottnangian and partly with the Karpatian stage (e.g., Rutsch 1958; Cicha & Tejkal 1959; Rögl et al. 1978; Roetzel et al. 2006). The same problem holds true for the "Tortonian" which was ambiguously subdivided into sub-stages that did not correspond to the Badenian biozones defined previously by Grill (1943).

In the following text, the actual definition of CP regional stages is summarized for the time span from 20.4 to 11.6 Ma. Time scale modifications suggested over the last decades and the validity of chronostratigraphic estimates of the boundaries gained by point-based dating are discussed. Attention is also drawn to deficiency in definition of stages often caused by an ecostratigraphic approach.

The base of the Eggenburgian was situated by Piller et al. (2007) coevally with the base of the standard Burdigalian stage within the calcareous nannoplankton NN2 Zone, at the sequence boundary Bur1 (~20.4 Ma). The Eggenburgian transgression can be detected in the Alpine Foredeep and in the Vienna Basin, but not in the northern realm of the Pannonian Basin, where the NN1/NN2 boundary was in the past correlated with the Egerian/Eggenburgian boundary at 22.8 Ma (*sensu* Vass & Elečko 1989). However, the FO of *Helicosphaera ampliaperta* (correlated with the Aquitanian/Burdigalian boundary at 20.43 Ma; *sensu* Gradstein et al. 2012) can be recognized in sediments in most CP basins (Holcová 2002; Krijgsman & Piller 2012). Therefore, this biostratigraphical event at the Bur1 boundary can be accepted as a reliable level enabling correlation between the regional and standard zonation (Fig. 1).

The Ottnangian regional stage (~18.3–17.3 Ma; *sensu* Piller et al. 2007) lower boundary was placed in the NN3 Zone, while the upper boundary was situated within the NN4 Zone, bounded approximately by the Bur3 and Bur4 3rd order sequence boundaries (after Haq et al. 1988 and Hardenbol et al. 1998).

The Karpatian stage (~17.3–16.4 Ma; *sensu* Rögl et al. 2003) was situated inside the NN4 Zone as well, and its lower boundary was defined by the FO of endemic *Uvigerina graciliformis*. Nevertheless, the new magnetostratigraphic constraints provided by Sant et al. (2017a) dated the transition from the Ottnangian marine to brackish sediments in the south-German part of the Alpine Foredeep (Molasse Basin) to ~17.7–17.5 Ma. Termination of the brackish depositional environment in the Austrian part of the foredeep was dated to ~17.2 Ma (Roetzel et al. 2014), while the Karpatian marine sedimentation in the Korneuburg Basin was dated by astronomical tuning of the gamma ray record to the time interval from 17.0 to 16.3 Ma by Zuschin et al. (2014). These results, supported by $^{87}Sr/^{86}Sr$ isotope dating from the Vienna Basin (Hudáčková et al. 2003) point to an insufficiently defined boundary between the Ottnangian and the Karpatian. The foraminifera tests from deposits assigned to the Ottnangian in the Cunín-21 borehole provided the Sr-age of 17.01–16.9 Ma. Foraminifera from the Karpatian strata of the Gbely-100 borehole provided Sr-age of 16.3–15.9 Ma (Hudáčková et al. 2003). Moreover, Sr-age gained from the Cerová-Lieskové site assigned to the Karpatian is 16.26–15.47 Ma (Less et al. 2015; Kováč et al. 2017a).

Differences in age estimates of the Ottnangian/Karpatian boundary probably led to incorrect correlations of sedimentary successions in an interregional context. We assume that the sediments of the same age were in the Alpine Foredeep (Molasse zone) assigned to the Ottnangian and in the northern part of the Vienna Basin to the Karpatian, both assigned to these stages on the basis of NN4 Zone. This assumption can be supported by a distinct angular unconformity between the two "Karpatian" sedimentary formations in the northern Vienna Basin (fig. 9 in Kováč et al. 2004). The lower "Karpatian" complex was possibly deposited during the "Ottnangian" closing of the marine connection towards the Mediterranean in front of the Alps, and the overlying complex was deposited during the "Karpatian" opening of the new marine connection via the Trans-Tethyan-Trench Corridor (*sensu* Rögl 1998; Mandic et al. 2002; Kováč et al. 2007; Rasser et al. 2008). To test this assumption, geochronological data obtained from basins situated close to the gateways between the CP and the Mediterranean can be used.

The Karpatian/Badenian boundary, set within the NN4 Zone, was initially correlated with the boundary between the Burdigalian and the Langhian, thus, with the boundary between the Early and Middle Miocene (*sensu* Blow 1969; Rögl et al. 1978; Piller et al. 2007). Currently, there is no consensus on the placement of the Early/Middle Miocene boundary in the CP. Piller et al. (2007) correlated it with the Bur5/Lan1 sequence boundary, while Hohenegger et al. (2014) shifted the base of the Badenian into the Burdigalian stage, corresponding to the FO of *Praeorbulina* at ~16.4 Ma in the Styrian Basin (Hohenegger et al. 2009). De Leeuw et al. (2013) placed the FO of *Praeorbulina glomerosa* on 16 Ma in the Transylvanian Basin. Krijgsman & Piller (2012) placed the Karpatian/Badenian boundary at 15.97 Ma (Fig. 1). The *Globigerinoides–Praeorbulina* lineage is continuously recorded in the Styrian, Sava, and Transylvanian basins (Krézsek & Filipescu 2005; Hohenegger et al. 2009, 2014; Premec Fućek et al. 2017). The FO of *Praeorbulina* cannot always be estimated in the Western Carpathian basins because they are extremely rare or absent (e.g., Andrejeva Grigorovič et al. 2001; Kováč et al. 2007; Rögl et al. 2008). The definition of the Badenian stage lower boundary was designated by the onset of *Praeorbulina sicana* (currently accepted as *Trilobatus sicanus*; erroneous synonym "*Globigerinoides sicanus*" is used by some authors either for *G. bisphericus* or for *Pr. sicana*) within the NN4 Zone (16.303 Ma at the top of chron C5Cn.2n) at the Wagna site in the Styrian Basin by Rögl et al. (2003) and confirmed by Hohenegger et al. (2009, 2014).

The original sub-division of the Badenian regional stage into the Early (Moravian), Middle (Wielician) and Late (Kosovian) sub-stages (Papp et al. 1978; Piller et al. 2007) remains problematic as well. According to Hohenegger et al. (2014), the Wielician sub-stage, namely the evaporite sequence at/below the base of the NN6 Zone, cannot be simply coeval with the "Middle Badenian" zone with agglutinated foraminifera in the western part of the CP because this foraminifera zone covers a much longer time span (upper part of the NN5 and the lowermost part of the NN6 zones; Andrejeva Grigorovič et al. 2001). Therefore, instead of referring to the "Wielician sub-stage", it is more appropriate to use the term Badenian Salinity Crisis (BSC).

The BSC is a reasonable correlation interval, with the duration of approximately 500 kyr between ~13.8–13.3 Ma, which is well documented in the eastern part of the CP (e.g., Filipescu & Gîrbacea 1997; Krézsek & Filipescu 2005; Peryt 2006; de Leeuw et al. 2010, 2013). This interval has also been detected in the sediments of the Pannonian realm (Báldi et al. 2017) and in the wider Mediterranean area (Ied et al. 2011). The base of the BSC (when dated by the geochronological methods) can thus be a reliable correlation level for the CP because it seems to be synchronous with the Langhian/Serravallian boundary, corresponding to a major glacioeustatic sea-level drop (*sensu* Gradstein et al. 2012).

Hoheneggers' et al. (2014) attempt to solve the "Badenian conundrum" brought even more confusion into the CP stratigraphy (Fig. 1). Although the "Middle Badenian" sub-stage Wielician was not accepted by Hohenegger et al. (2014) and the BSC range was assigned to the base of the Late Badenian (Hohenegger et al. 2014), the term "Wielician sub-stage" is still used in studies from the Eastern Carpathian region (e.g., de Leeuw et al. 2013; Gonera et al. 2014; Palcu et al. 2015). It is, however, improper to consider the Moravian sub-stage within the NN5 Zone, introduced for the "Early Badenian" by Papp et al. (1978), as the (re)established "Mid Badenian" (Hohenegger et al. 2014).

Another attempt to correlate the Badenian regional stage with the standard Mediterranean time scale resulted in the division of Badenian into lower and upper parts, roughly corresponding to the Langhian and early Serravallian (Kováč et al. 2007). This definition, placing the BSC at the top of the Early Badenian, led to a shift of the Late Badenian lower boundary to 13.63 Ma (instead of 13.82 Ma).

The Sarmatian stage defined as Sarmatian *s.s.* and Sarmatian *s.l.* is difficult to correlate even between the western and eastern part of the CP (e.g., Suess 1866; Papp et al. 1974). In the eastern part, the Sarmatian *s.l.* is divided into sub-stages Volhynian, Bessarabian, and Khersonian, thus a subset of the Sarmatian *s.l.* corresponds to the regional Pannonian stage in the west (e.g., Piller et al. 2007; Popov et al. 2010; Gozhyk et al. 2015). For the subdivision of the Sarmatian *s.s.* (12.7–11.6 Ma; *sensu* Piller et al. 2007), four successive zones (*Anomalinoides dividens*, large elphidia, *E. hauerinum*, *Porosononion granosum*) are used (*sensu* Grill 1941, 1943; Papp 1951; Harzhauser & Piller 2004). However, preliminary analyses of foraminiferal assemblages from boreholes cores positioned in a 3D seismic model in the northern Vienna Basin indicate that these assemblages track temporally shifting environments and their temporal distribution depends strongly on the former basin topography (Hudáčková et al. 2013). In the Transylvanian Basin the Badenian/Sarmatian boundary was dated by the $^{40}Ar/^{39}Ar$ method to 12.80±0.05 Ma (de Leeuw et al. 2013). This datum is similar to the one brought by Harzhauser & Piller (2004) on the basis of sequence stratigraphy from the western margin of the Central Paratethys, and correlates with the magnetostratigraphic results of Paulissen et al. (2011) from the Vienna Basin.

The lower boundary of the Sarmatian *s.s.* was set by Harzhauser & Piller (2007) to the extinction event at the Badenian/Sarmatian boundary (BSEE). However, it seems that the BSEE timing is diachronous due to complex tectonic evolution of the Carpathian–Pannonian region, reflecting the final isolation of the CP from neighbouring basins — the Mediterranean and Eastern Paratethys (e.g., Magyar et al. 1999; de Leeuw et al. 2013; Palcu et al. 2015; Kováč et al. 2017a,b and references therein). Therefore, point-based data are necessary to support this hypothesis, as already suggested by Silye & Filipescu (2016).

Similarly, the extinction event at the top of the Sarmatian *s.s.* (SPEE) has been placed at different levels. Harzhauser & Piller (2004) placed it at 11.6 Ma. Similar results for the Sarmatian/Pannonian boundary were brought by Paulissen et al. (2011) and ter Borgh et al. (2013) by magnetostratigraphy from the Vienna Basin and the southern Pannonian Basin. Dating of two volcanoclastic layers located approximately 40 m below the Sarmatian–Pannonian transition (Transylvanian Basin) yielded $^{40}Ar/^{39}Ar$ ages of 11.62±0.12 Ma and 11.65±0.13 Ma (de Leeuw et al. 2013). Based on the sequence stratigraphic correlations to global sea level curves in the Vienna Basin Lirer et al. (2009) estimated the Sarmatian/Pannonian boundary at 11.4 Ma. The Sarmatian/Pannonian boundary in the Transylvanian Basin was dated to an age of 11.3±0.1 Ma (Vasiliev et al. 2010). The correlation of the NN8 Zone in the Paratethys domain (based on the presence of *Catinaster coalitus*; Galović & Young 2012) with magnetostratigraphic data points to the Sarmatian–Pannonian transition in the Croatian Basin around 11.2 Ma.

The perspectives of regional and standard time scale correlation

On the one hand, the correlation between sedimentary sequences is relatively simple if only one time scale is used for a single basin. On the other hand, it is difficult to compare time scales of basins characterized by their partial isolation from the World Ocean (WO) with the standard time scale (GTS). First and last appearances of species represent one of the major biostratigraphic tools. However, discrepancies in the timing of first appearances of particular species exist even between the Atlantic and Mediterranean, and such discrepancies can be expected to be more pronounced between the CP and other basins.

For example, the Early/Middle Miocene boundary is apparently correlated with the Burdigalian/Langhian stage boundary (Hilgen et al. 2012; Turco et al. 2017). Using the latest Time Scale Creator database (Fig. 1) the base of Langhian stage is correlated with the base of magnetic polarity chron C5Br (15.97 Ma) and with the FO of *Praeorbulina circularis* which is in accordance with the scale of Ogg et al. (2016). The boundary is not officially established, so the reliability of such “praeorbulina datum” can be doubtful (Lirer & Iaccarino 2011). In this context, the FO of *Praeorbulina glomerosa glomerosa* at 15.2 Ma is the key event in the Mediterranean (Iaccarino et al. 2011; Turco et al. 2017) while it occurs at 16.4 Ma in the WO (Wade et al. 2011). *Orbulina suturalis* appears at 14.6 Ma in the Mediterranean (Abdul Aziz et al. 2008; Di Stefano et al. 2008) and at 15.1 Ma in the WO (Wade et al. 2011), as reviewed by Sant et al. (2017b). The Langhian delay of the FO of the planktic foraminifera in the Mediterranean compared to the Atlantic Ocean can be explained by: *(i)* circulation patterns that did not allow immediate migration of planktic species to the Mediterranean and/or *(ii)* the establishment of conditions for survival of these species in the Mediterranean, which was influenced by the inflow of CP water masses into the Mediterranean realm at that time (Kováč et al. 2017a; Sant et al. 2017b). Moreover, in the Mediterranean, the last common occurrence (LCO) of *Helicosphaera ampliaperta* is dated to 16.1 Ma (Iaccarino et al., 2011). The LO of *H. ampliaperta* (~14.9 Ma in WO) defining the top of the NN4 Zone cannot be properly recognized in the Mediterranean (Di Stefano et al. 2008, 2015). Therefore, using this event for the NN4/NN5 boundary accompanied by the FO of *Orbulina suturalis* at 14.6 Ma in the Mediterranean (Abdul Aziz et al. 2008) while in the WO it appears at 15.1 Ma (Gradstein et al. 2012) is not satisfactory.

Inconsistencies generated by converting the regional stages to standard ones are partly caused by the lack of multiple point-based geochronological data, by inadequate biostratigraphic data that do not account for temporal shifts in geographic ranges of index species, and by problems with local nomenclature in lithostratigraphy. Therefore, the correlation of individual basins within the CP as well as with the Mediterranean or Eastern Paratethys realms without accurate geochronological data remains problematic.

Reflection of eustatic sea-level changes in the Central Paratethys time scale

The problems of CP sequence stratigraphy are well documented in the Vienna, Danube, Transylvanian, and other basins (e.g., Kováč 2000; Kováč et al. 2004, 2007, 2008; Krézsek & Filipescu 2005). The Miocene depositional sequences reveal several 3rd and 4th order cycles that were generated by eustatic sea-level changes, tectonic evolution of basins, and local sediment supply delivered by deltas. However, the global sequence boundaries *sensu* Hardenbol et al. (1998) were tied to the regional stage boundaries (Piller et al. 2007). This sequence–stratigraphic definition of stage boundaries partly contrasts with the local sequence stratigraphy following local geodynamic events as demonstrated by Kováč et al. (2004). Moreover, due to active tectonics and rapid palaeogeographic changes in the Alpine–Carpathian–Dinaride domain, it is difficult to discriminate between the 3rd and 4th order cycles during the Miocene (Fig. 2).

Around the Aquitanian–Burdigalian transition, the marine connections of the CP with Mediterranean probably led through a strait between the Alps and Dinarides. The connections between the Central and Eastern Paratethys (and possibly up to Indo–Pacific) via a strait between the Volhynian High and Moesia were gradually closing (e.g., Popov et al. 2004; Kováč et al. 2017a,b).

The following early Burdigalian sea-level changes in the CP were probably influenced by the sea-level rise or fall transferred from the Mediterranean through a new connection in front of the Alps (e.g., Rögl 1998; Harzhauser & Piller 2007; Kováč et al. 2017a,b). The gateway opened at the Bur1 boundary (~20.4 Ma) and closed ~17.7–17.5 Ma, as constrained by magnetostratigraphic data (Sant et al. 2017a). According to GTS, the gateway was closed at the Bur4 boundary (17.5 Ma after Hardenbol et al. 1998; Piller et al. 2007). During this time interval two 4th order cycles (Eggenburgian and Ottnangian; Fig. 2) were documented in the Vienna Basin and adjacent

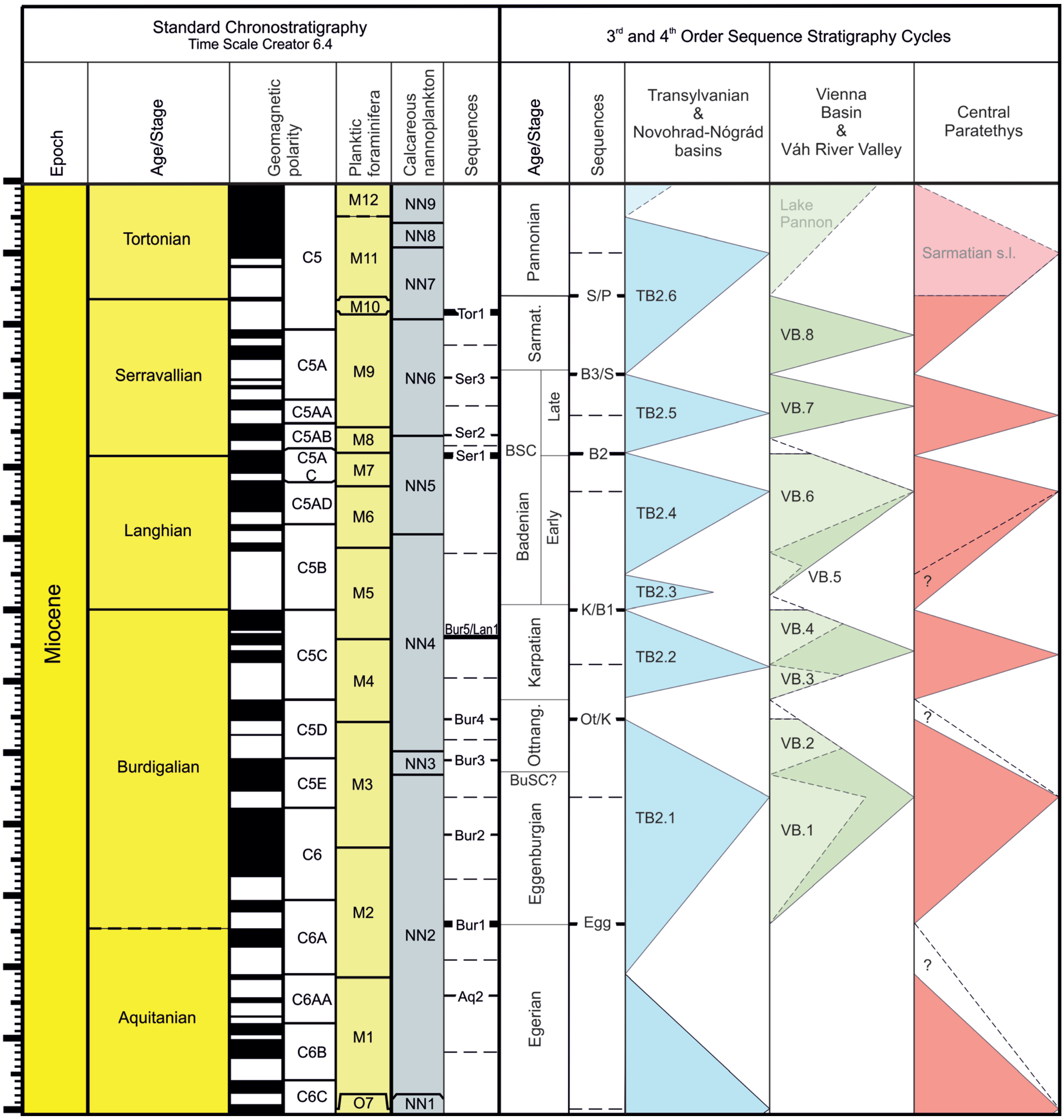

Fig. 2. Central Paratethys sequence stratigraphy; 3rd and 4th order cycles of the entire Central Paratethys (this work), as well as the Novohrad–Nógrád, Transylvanian and Vienna basins (after Kováč 2000; Krézsek & Filipescu 2005; Kováč et al. 2007; Pezlej et al. 2013).

basins to the north-east (Kováč 2000; Kováč et al. 2004). The regression of the Ottnangian Sea and development of the "Rzehakia Lake" (e.g., Rögl 1998) is a significant marker horizon preceding the Karpatian full marine transgression in the Alpine–Carpathian junction.

The Early Miocene eustatic changes in the Novohrad–Nógrád Basin represent two 3rd order cycles (Egerian–Eggenburgian and Eggenburgian–Ottnangian; Kováč 2000) and thus cannot be directly compared with the Vienna Basin 4th order cycles (Fig. 2). A similar situation can be deduced from the lithostratigraphy and micropalaeontology of the Transylvanian Basin (Filipescu 2011), where an Egerian–Eggenburgian and Eggenburgian–Ottnangian cycle can be distinguished as well. The deep-sea equivalent of the Eggenburgian onshore formations corresponds to the lower part of mid-fan turbidites of the Hida Formation. The upper part of this formation displays a regressive trend towards the boundary with the Middle Miocene. The nannoplankton assemblages indicate Early Miocene age (NN2 to NN4 zones; Mészáros 1991; Beldean et al. 2010) while the agglutinated and planktic foraminifera point to a wider span (Iva et al. 1971; Beldean et al. 2010; Beldean & Filipescu 2011). Therefore, the correlation between the Early Miocene western and eastern CP basins sequence stratigraphy remains unclear (Fig. 2).

The late Burdigalian eustatic sea-level rise was transferred from the Mediterranean into the CP via a new marine gateway — the Trans-Tethyan-Trench Corridor (*sensu* Rögl 1998; Mandic et al. 2002; Kováč et al. 2007; Rasser et al. 2008). The early Karpatian full marine transgression (NN4 Zone with *Uvigerina graciliformis*) represents the next pronounced sea-level change in the Vienna Basin. The base of this 4th order cycle can be coeval with Bur4 boundary (*sensu* Hardenbol et al. 1998). The next late Karpatian 4th order cycle inside the NN4 Zone is situated above the lower Karpatian depositional sequence (Kováč et al. 2004). In the Novohrad–Nógrád Basin, the "Karpatian" transgression started at the top of Ottnangian "Oncophora beds" (Vass & Elečko 1992; Holcová 2001) and was followed by offshore deposition inside the NN4 Zone (Kováč 2000). The local 3rd order cycle is capped by sediments of the Early Badenian transgression (Fig. 2).

The Karpatian stage was terminated by a regression causing large-scale erosion in the northern Vienna Basin around the sequence boundary of the local VB4/VB5 3rd order cycles (Kováč et al. 2004). The sedimentation above this boundary continued in the form of lobes of deltaic and alluvial sediments, followed by the Early Badenian transgression in the entire basin and in the junction towards the Alpine Molasse Zone (*sensu* Kováč et al. 2004). Strauss et al. (2006) correlated this lower boundary of the Badenian cycle with the Bur5/Lan1 3rd order cycle and placed it at 16.4 Ma (after Hardenbol et al. 1998).

The erosion and deposition around the Karpatian/Badenian boundary fits well with the regressive phase during the late Karpatian and the Early Badenian transgression. The geochronological point-based data from the Vienna and Novohrad–Nógrád basins (Hudáčková et al. 2003; Fordinál et al. 2014; Kováč et al. 2017a) show that the age of the top part of the Karpatian eustatic cycle (<16 Ma) does not coincide with the global Bur5/Lan1 boundary (16.4 Ma; *sensu* Hardenbol et al. 1998; Krijgsman & Piller 2012), and also does not coincide with the base of the "earliest" Badenian (*sensu* Hohenegger et al. 2014). The relative sea-level fall (prior to the Early Badenian transgression) in the CP estimated as up to 200 m was enhanced by the mountain uplift (compiled from Krézsek & Filipescu 2005; Dellmour & Harzhauser 2012; Filipescu 2011; Kováč et al. 2017a). Therefore, the sea-level fall probably began after ~16.5 Ma (Fig. 2) and the sea-level low-stand probably lasted until ~15.5–15.1 Ma, when the Badenian transgression was initiated. In this case, the absolute age of the Early Badenian sequence boundary does not simply coincide with the Bur/Lan1 boundary.

The termination of the Vienna Basin initial rifting led to a decrease in subsidence rates and to a very indistinct reflection of the global TB2.3 cycle (16.5–15.5 Ma; *sensu* Haq et al. 1988; Haq 1991; Hardenbol et al. 1998). Conglomerates at the base of the local 4th order Early Badenian cycle (defined as the 3rd order VB5; Kováč et al. 2004) are overlain by pelites dated by Kroh et al. (2003) and placed to the uppermost part of the "Lower Lagenidae Zone" (*sensu* Grill 1943) based on co-occurrence of *Praeorbulina glomerosa circularis*, *Orbulina suturalis,* and *Trilobatus bisphericus*. The nannoplankton assemblage with *Helicosphaera waltrans*, *Sphenolithus heteromorphus*, *Calcidiscus premacintyrei*, *Reticulofenestra pseudoumbilicus*, *Coccolithus miopelagicus*, rare *Discoaster deflandrei* and *D. variabilis* indicates that the pelites belong to the NN5a Zone of the "Early Badenian" (*sensu* Kováč et al. 2007). Strauss et al. (2006) documented the local "Early" Badenian cycle as an equivalent of the 3rd order TB2.3 cycle (15.97–14.4 Ma; *sensu* Hardenbol et al. 1998) in the SE Vienna Basin. This local cycle is situated below the next "Middle" Badenian cycle, as an equivalent of the 3rd order TB2.4 cycle (14.4–13.65 Ma; *sensu* Hardenbol et al. 1998). A similar, local 3rd order cycle was documented in the southern Pannonian Basin System (Pezelj et al. 2013).

The age of the Early Badenian transgression can be deduced from borehole cores in the eastern Danube Basin (Kováč et al. 2018). The Badenian basal conglomerates and silts, both without volcano-clastics, are overlain by siliciclastics with tuffites. These sediments contain nannofossils of the NN5a Zone with common *Orbulina suturalis* (FO of *O. suturalis* at 14.56 Ma; Abdul Aziz et al. 2008). The volcanic activity related to the basin opening started at 15 Ma (Pécskay et al. 2006) and points to the age of the marine flooding with *Praeorbulina* spp. prior to the deposition of volcanoclastic sequences (Fig. 2). Similarly, in the Transylvanian Basin, the onset of Dej Tuff volcanism dated by $^{40}Ar/^{39}Ar$ method to 14.38±0.06 Ma is also preceded by the FO of *Praeorbulina* spp. and *Orbulina suturalis* (de Leeuw et al. 2013).

The boundary between the "Early" Badenian (VB5) and "Middle" Badenian (VB6) 4th order cycles (3rd order; *sensu* Kováč et al. 2004) corresponds to the sequence boundary within the "Upper Lagenidae Zone" proposed by Weissenbäck (1996) in the southern Vienna Basin. The "Middle" Badenian cycle (uppermost Lagenidae Zone–lower Bulimina–Bolivina Zone; *sensu* Grill 1943) covers the time span of the NN5 Zone upper part, and the maximum flooding surface was identified by Weissenbäck (1996) within the Spiroplectammina carinata Zone. The high-stand system deposits are capped with the base of the Late Badenian (VB7) local 3rd order cycle (Harzhauser et al. 2018).

In the Transylvanian Basin, the sedimentary record at the Early–Middle Miocene transition offers similar proofs in the form of changes in sedimentary facies and microfossil assemblages. The upper part of the Lower Miocene sediments contains foraminifera assemblages dominated by planktic *Streptochilus pristinum* associated with rare benthics (*Bulimina, Bolivina, Cibibicidoides*) and calcareous nannoplankton, probably indicating a late Burdigalian age (Beldean et al. 2010, 2013). The sea-level drop (100–200 m) is documented by several deep incised valleys (Krézsek & Filipescu 2005). The overlying Middle Miocene deposits comprising tuffites and fall-out tuffs interbedded with low density fine siliciclastics contain typical Lower Badenian planktic foraminifera (species belonging to genera *Praeorbulina, Orbulina,* *Globigerinoides, Globorotalia*). These deposits can be linked to the main phase of marine transgression that started at the beginning of the Middle Miocene. The "Early" and "Middle" Badenian depositional cycles correlated with the 3rd order TB2.3 and TB2.4 cycles of global sea-level change (*sensu* Hardenbol et al. 1998) terminated prior to the BSC (Krézsek & Filipescu 2005). The plankton bloom in the Praeorbulina glomerosa Biozone (M5a) was followed by bloom in the Orbulina suturalis Biozone (M5b) Krézsek & Filipescu (2005). The lower boundary of the Late Badenian 3rd order cycle VB7 in the Vienna Basin (*sensu* Kováč et al. 2004) is represented by subaerial erosion in its NE part.

The Late Badenian depositional systems in the SW Vienna Basin were considered to be of regressive origin (Kreutzer & Hlavatý 1990; Weissenbäck 1996), while Kováč et al. (2004) defined a complete 3rd order cycle (VB7) in the NE Vienna Basin. Later on, this cycle was defined for the CP (13.65–12.7 Ma; Kováč et al. 2007) and correlated with TB2.5 global cycle (13.8–12.6 Ma; Haq et al. 1988). The latest research in the northern Vienna Basin has confirmed the sequence boundary between the "Middle" and "Late" Badenian (Harzhauser et al. 2018). Moreover, a huge sea-level drop is correlated with the base of the BSC and thus with the base of the Serravallian at 13.8 Ma (Harzhauser et al. 2018). This actually indicates that the base of the TB2.5 is captured by the Vienna Basin sequences (Fig. 2). The TB2.5 cycle in the Transylvanian Basin is correlated with two local 4th order depositional cycles: the Badenian and the early Sarmatian lasting from BSC base to top of the Anomalinoides dividens Biozone (Krézsek & Filipescu 2005).

The Badenian/Sarmatian sequence boundary in the Vienna Basin is placed at the biostratigraphic boundary defined by molluscs and foraminifera turnover (*sensu* Harzhauser & Piller 2007) affected by salinity decrease (Kováč & Hudáčková 1997). On the other hand seismic lines and well-logs show overlap of the VB7 into the earliest Sarmatian sediments (Harzhauser & Piller 2004). The base of the local Sarmatian VB8 3rd order cycle (*sensu* Kováč et al. 2004) is well recorded by a transgressional overlap on the Upper Badenian sediments. The falling sea-level in the terminal Sarmatian (uppermost Porosononion granosum Zone = "pauperization" Zone; *sensu* Papp 1956) caused a shift of the littoral zone far into the basin, indicated by littoral potamidid-bearing sand with scattered coal in the basin drillings (Harzhauser & Piller 2004). The regression at the end of the Sarmatian is also indicated by local erosions and incision of deltaic feeding channels.

In the Transylvanian Basin, the Sarmatian deposits represent a single 3rd order depositional cycle. In contrast, two Sarmatian 4th order cycles consisting of parasequences were documented the Vienna Basin (e.g., Harzhauser & Piller 2004). These parasequence sets are present in the entire basin, as well as in other basins of the Carpathian–Pannonian region (Styrian and Transylvanian basins), suggesting that they were governed by orbital impulses — a common feature of different basins in CP realm (Kováč et al. 2008).

To summarize, extensive erosion characterized the Burdigalian–Langhian transition due to sea-level drop in the CP at ~16–15.5 Ma. The Early Badenian 3rd order eustatic cycle ended prior to the BSC (Figs. 1 and 2). The younger 3rd order cycles are marked by the Late Badenian and Sarmatian transgressions. The three Middle Miocene Central Paratethys 3rd order cycles of sea-level changes can be only partly correlated with the Langhian and Serravallian global sea-level changes (*sensu* Kováč 2000; Krézsek & Filipescu 2005; Strauss et al. 2006; Kováč et al. 2007). Additional geochronological data are needed to improve correlation of depositional sequences between the CP basins and to untangle the effects of regional tectonics from the effects of global eustatic changes.

The Central Paratethys time scale adjusted to geodynamic development

(i) The geochronological definition of regional stage boundaries, (ii) the appropriate application of the point-based data supported by well-defined biostratigraphic correlation levels, (iii) the refined CP sea-level changes, (iv) the interpretation of the plankton and benthos migration driven by opening and closing of gateways between the Mediterranean, the Central Paratethys, and the Eastern Paratethys, associated with taphonomic and palaeoecological inferences on the role of reworking, preservation and habitat suitability in determining FO and LO in individual sections, (v) and the palinspastic approach should result in the reappraisal tuning of the CP time scale in respect to geodynamic processes, enabling better correlation with the standard chronostratigraphy of the Miocene period (GTS). Below, we propose three intervals of the CP evolution with respect to geodynamic development of the area and different positions of sea gateways; more likely as a reflection of geodynamically induced changes and only partially corresponding to changes in the global sea-level.

The Burdigalian transgression represents the onset of pronounced 3rd order sequence stratigraphy cycle in the northern CP, correlating with the Bur1 boundary (Hardenbol et al. 1998; Piller et al. 2007; Krijgsman & Piller 2012). The base of the Eggenburgian is dated by the FO of *Helicosphaera ampliaperta,* like the base of the Burdigalian stage in the Mediterranean (20.4 Ma; Piller et al. 2007; Ogg et al. 2016).

During the Eggenburgian and Ottnangian, after the closing of the earliest Miocene connections towards the Eastern Paratethys and Mediterranean (e.g., Popov et al. 2004; Kováč et al. 2017a,b and references therein) a new marine flooding from the Mediterranean went through the foredeep basin in front of the Alps. The time span of this connection is documented by the presence of calcareous nannofossil NN2, NN3, and a part of NN4 zones (*sensu* Martini 1971). The gateway opened around the Aquitanian/Burdigalian boundary and the sea (Fig. 3A) flooded the foreland and hinterland of the developing Carpathian mountain chain (e.g., Kováč et al. 2017b and references therein). In the distal part of the CP, the isolation led to development of hypersaline facies, later also to hyposaline facies, probably due to spatio–temporal shifts in rainfall distribution (e.g., Kováč et al. 2017a). In the eastern segment of the Carpathian Foredeep, evaporites of the Vorotyshche Formation were deposited during the Eggenburgian (e.g., Gozhyk et al. 2015), while a system of brackish and freshwater lakes with endemic *Rzehakia* fauna developed in the late Ottnangian (Harzhauser & Piller 2007; Harzhauser & Mandic 2008). In the foreland and hinterland of the developing Carpathian mountain chain, as well as in some parts of the Pannonian domain, the terrestrial (lake) sedimentation prevailed between 18 and 17 Ma. In this area, situated along junction of the Central Western Carpathians and Northern Pannonian domain a continuous Early Miocene terrestrial sedimentation is documented by radiometric dating at 17.4–17.02 Ma (Pálfy et al. 2007). However, along the northern margin of the Pannonian domain (Novohrad–Nógrád Basin), the "Ottnangian" sediments containing *Rzehakia* fauna are occasionally accompanied by the Karpatian index species *Uvigerina graciliformis* (Holcová 2001).

The gateway in front of the Alps disappeared before the end of the Early Miocene (~17.5 Ma). Roughly at the same time a new marine strait between the Mediterranean and CP opened in the hinterland of the Eastern Alps, following the northern edge of Dinarides (Fig. 3B). The so-called Trans-Tethyan-Trench Corridor (e.g., Rögl 1998; Piller et al. 2007; Sant et al. 2017b) was active during the upper part of NN4 and NN5 zones (*sensu* Martini 1971). The base of the local early Karpatian 4th order cycle in the CP can be approximately correlated with the Bur4 boundary, or slightly above it. The deposition of the "upper Karpatian–lowermost Badenian" sediments was associated with significant changes in geomorphology, especially with the uplift of mountain chains, accompanied by local fluctuations in humidity. During the relative sea-level fall by up to 200 m, a huge erosion of mostly Karpatian strata and development of a pronounced 3rd order sequence boundary (placed above Bur5/Lan1 boundary of GTS) is assumed. The development of a new river network caused the input of voluminous masses of fresh water into the sea which probably triggered the switch of circulation regime, at least in the western part of the CP during this time (Fig. 1). The shift from an anti-estuarine to an estuarine circulation regime (during the latest Burdigalian and early Langhian) delayed plankton immigration into the CP (Kováč et al. 2017a) and probably also influenced the marine environment in the adjacent Mediterranean area (e.g., problem with the Burdigalian/Langhian boundary definition in the Mediterranean; Iaccarino et al. 2011; Lirer & Iaccarino 2011).

During the Early Badenian transgression, an anti-estuarine regime between the CP and the Mediterranean was re-established again (Kováč et al. 2017a). The base of this Central Paratethys 3rd order cycle can be placed inside the Langhian 3rd order sequence of GTS (below the maximal flooding on the global sea-level curve) bordered by the Bur5/Lan1 and Ser1 boundaries. This assumption is supported by the Lower Badenian sediments with the FO of *Orbulina suturalis* together with the NN5 Zone at ~14.6 Ma. However, the occurrence of the *Praeorbulina* spp. in several CP basins could suggest the Middle Miocene transgression around 16–15 Ma (Fig. 2).

From what has been discussed above, a couple of questions arise: where is the boundary between the Lower and Middle Miocene in the sedimentary record, and how should we understand the Karpatian regional stage? In other words: Does the sedimentary sequence assigned to the Karpatian belongs to the Early Miocene time span? We suggest that the lower part of the deposits assigned to the Karpatian regional stage belongs to the Early Miocene, whereas the upper part belongs to the Middle Miocene.

The temporal span of the Karpatian regional stage remains unclear. The Karpatian marine transgression is documented from the southern Vienna Basin at ~17 Ma (Zuschin et al. 2014), while the base of the Badenian is placed at ~16.4 Ma (e.g., Piller et al. 2007; Hohenegger et al. 2014), thus the Karpatian stage is just limited to 600 ky. We note that the lower boundary of the Badenian stage, as suggested by Papp et al. (1978), should be placed at the beginning of the NN5 Zone, which means ~15 Ma, whereas Hohenegger et al. (2014) considered the time interval between 16.3 and 15.1 Ma as the "lowermost" Badenian. Following Papp et al. (1978), the resulting time interval would last ~2 Ma (i.e. "Karpatian–lowermost Badenian"; *sensu* Hohenegger et al. 2014). Significant changes in CP palaeogeography took place between 17 and 15 Ma, controlled predominantly by geodynamic development of the Alpine–Carpathian–Dinaride orogenic systems (Kováč et al. 2017b), and therefore it would be appropriate to define a new regional stage between the Ottnangian and (re)defined Badenian on the basis of geochronological dating and constrained by the Central Paratethys 3rd order sequence stratigraphy.

The Late Badenian and Sarmatian *s.s.* sub-stages represented a period when the connection to the Mediterranean was gradually closed (or at least its existence has not been sufficiently proved). The connection to the Eastern Paratethys most likely became opened (e.g., Popov et al. 2004; Bartol et al. 2014; Palcu et al. 2015; Silye & Filipescu 2016; Kováč et al. 2017a; Harzhauser et al. 2018). The view that the marine connection from the east was opened is also induced by a sea-level rise in the Eastern Paratethys during this time (Popov et al. 2010). The presence of the NN6 Zone is common in all CP basins due to marine connection with the Mediterranean in the west until the base of the Sarmatian (*sensu* Bartol et al. 2014) and with the Eastern Paratethys during the Late Badenian and Sarmatian (e.g., Popov et al. 2004). The NN7 Zone was identified only in several basins (Palcu et al. 2015; Kováč et al. 2017a).

Sedimentary sequences of the western part of the CP can therefore be roughly correlated with the early and late Serravallian (Fig. 3C), with the base at ~13.82 Ma (e.g., Hilgen et al. 2009; Iaccarino et al. 2011; de Leeuw et al. 2013) and the top at ~11.6 Ma (Hilgen et al. 2005; Vasiliev et al. 2010; de Leeuw et al. 2013). The connection with the Eastern Paratethys in front of the Carpathians persisted even longer

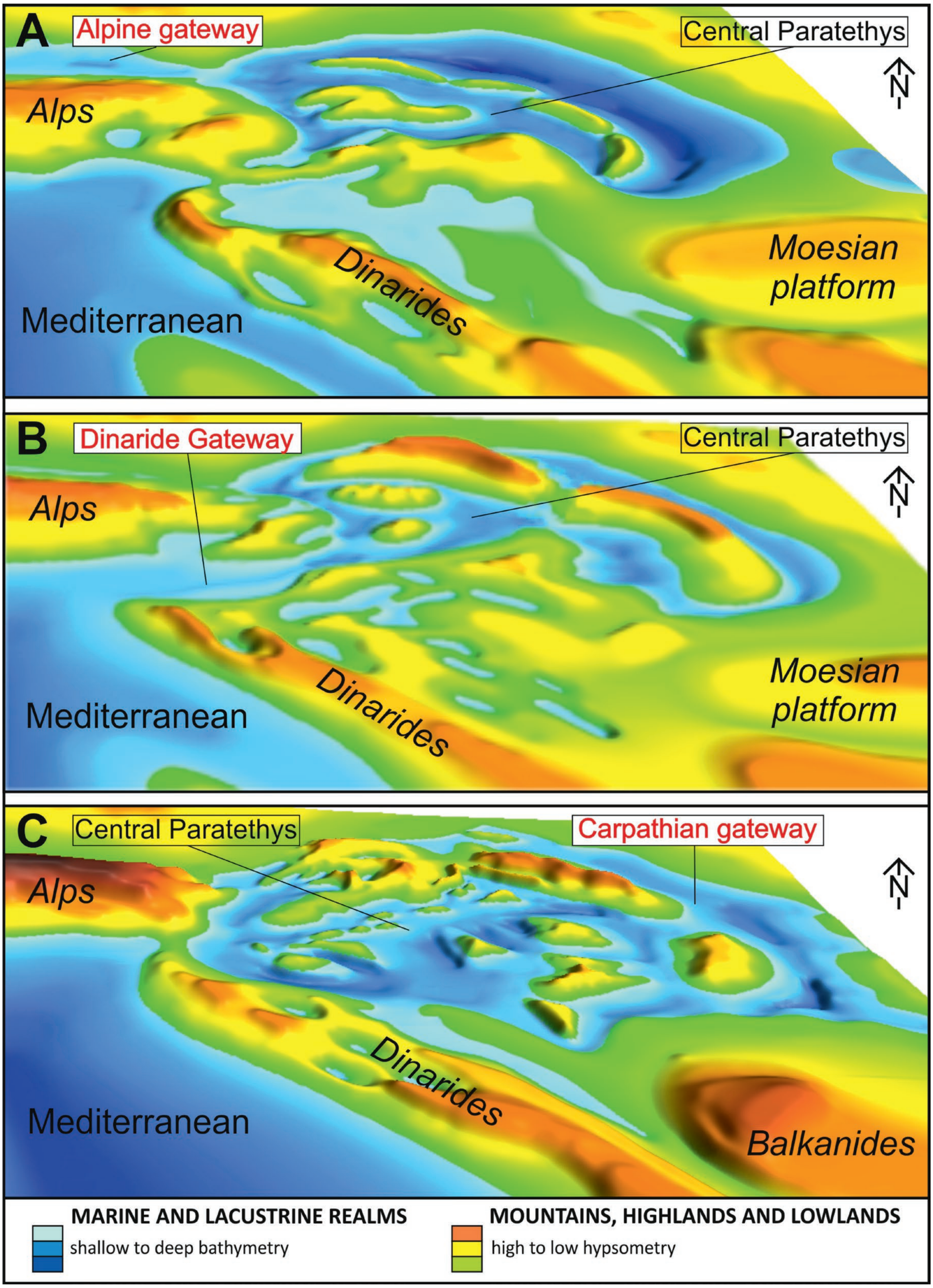

Fig. 3. Palinspastic–topographic maps of the Central Paratethys; modified after Hámor & Halmai 1988; Rögl 1998; Kováč 2000; Popov et al. 2004; Kováč et al. 2017a,b; Sant et al. 2017b): **A** — the Burdigalian CP with gateway in front of the Alps; **B** — the late Burdigalian–Langhian CP with gateway between the Alps and Dinarides; **C** — the Serravallian CP with gateway towards the Eastern Paratethys. Data were handled using the PostgreSQL Server v. 9.4; topology and spatial geometry using the GRASS-GIS v. 7.2.1; GRASS-GIS software was used to compute the location of each cell of the DTM using regularized spline with tension for approximation from vector data (module v.surf.rst; GRASS Development Team 2017).

and it is thus possible to correlate this time interval with the upper part of the Eastern Paratethyan Sarmatian *s.l.* (e.g., Popov et al. 2004; Gozhyk et al. 2015).

Conclusions

The CP time scale defined by biostratigraphic data remains poorly constrained by geochronological and spatially-explicit biostratigraphic methods, making the correlation with the standard GTS problematic (Fig. 1). The increase in spatial and temporal coverage of point-based geochronological data is therefore an essential task. In broad-scale palaeogeographic analyses requiring correlation of the CP with the Mediterranean, the standard geological time scale should be used as a reference to avoid problems with the definition of regional stages.

The rise or fall of the sea-level, as well as climate changes in the semi-enclosed CP realm often have a local character and were influenced by global sea-level changes only to some degree. The differences between the global, Mediterranean or the Eastern Paratethys sea-level curves indicate that the 3rd order sea-level cycles in the CP need to be further validated and the climate evolution should be better resolved (Fig. 2). The complex geodynamic evolution of the Alpine–Carpathian–Pannonian and Dinarides domains causes difficulties in correlation with GTS, even between individual CP basins. It would be beneficial to revise the regional time scale in respect with the geodynamics of the orogenic system, as well as the opening of gateways between the CP, Mediterranean, and Eastern Paratethys (Fig. 3A–C). The palaeogeographical reconstructions should reflect the original position and extent of basins which fill was later deformed by folding and thrusting in front of the orogenic system or by the movement of crustal fragments along several hundred km long transform boundaries. These observations were not taken into account for decades leading to palaeogeographical misconceptions on a European scale. To understand changes in the layout of sedimentary basins during distinct time spans an improved palinspastic model based on an interdisciplinary approach is needed in the future.

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