The Central Paratethys palaeoceanography: a water circulation model based on microfossil proxies, climate, and changes of depositional environment

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Abstract: A comprehensive "model" of the semi-closed Central Paratethys Sea history was proposed for the entire time span of 25 Ma. Publications and datasets of the last decades were compiled and reviewed in the light of the Western Carpathian basins record, reflecting its changing palaeogeography, palaeoenvironment, and water circulation regimes. Moreover, a set of modified palinspastic palaeogeographic maps were reconstructed taking into account the enormous changes in depositional systems and sedimentary facies, reflecting the tectonic events from the Oligocene to Miocene. The study discusses possible gateways/straits assuming two way circulations, where the surface and deep water can be exchanged between the Mediterranean, Central Paratethys, and the Eastern Paratethys. It is suggested, that a good correlation between the regional Central Paratethys and the standard Mediterranean zonations based on planktic organisms can be achieved only in sequences deposited during anti-estuarine circulation regimes. Correlation of sequences deposited during the estuarine circulation of water masses becomes problematic, making hard to define above all the Early/Middle Miocene boundary. To solve problems of the regional biostratigraphy, and the definition of correlation levels between the Mediterranean, Central Paratethys, and Eastern Paratethys, the existing and new Sr-isotope dating was used as well. Dating of specific sediments of low oxic conditions or evaporitic events was also used as time horizons for the Central Paratethys.

Key words: Cenozoic; Central Paratethys; Western Carpathians; palaeogeography; palaeoclimate; regional biostratigraphy; Sr-isotope dating

1. INTRODUCTION

The Cenozoic convergence between the European and Adriatic (African) plates resulted in immense changes in the palaeogeography of the Alpine–Carpathian–Dinaride orogenic system. Many geodynamical processes formed the landscape at a regional scale, such as subduction, collision, folding, and thrusting of the Neo-Alpine complexes of accretionary wedges and/or vertical movements of the Eo-Alpine consolidated crustal segments of the orogen (e.g., Ratschbacher et al., 1991[a,b]; Csontos et al., 1992; Kováč et al., 1994, 1997, 1998, 2016, 2017[a]; Matenco, 1997; Matenco et al., 1997, 2013; Konečný et al., 2002; Schmid et al., 2008; Ustaszewski et al., 2008; Handy et al., 2010, 2015; Studencka et al., 2016).

Starting from the Oligocene, a system of sedimentary basins developed due to changing tectonic regimes, stretching from the Western Mediterranean coast across the Alpine forelands to regions in front of and behind the Carpathian chain, and from there, up to the Caucasus and Pontides (Black and Caspian seas) in the east (Fig. 1). For these basins, the term Paratethys was coined by Laskarev (1924). From the Neogene, the connection with the Mediterranean, as well as among the individual basins, became very unstable. Partial or total isolation of these basins often resulted in the appearance of endemic fauna in each basin system, enabling the division of this inland sea into three parts: the Western (Alpine), Central (Carpathian, Balkan) and Eastern (Crimean–Caucasian) Paratethys (e.g., Seneš, 1961[a,b]; Popov et al., 2004; Piller et al., 2007; Gozhyk et al., 2015; Neubauer et al., 2015).

The Central Paratethys Cenozoic palaeogeography was noticeably influenced by geodynamic development of the Eastern Alpine and Western Carpathian segments of the orogenic chain, in its fore- and hinterland areas (e.g., Kováč et al., 1994, 1998, 2003, 2007, 2016, 2017[a]; Meulenkamp et al., 1996; Oszczypko, 1998, 1999, 2006). The periods of compression leading to the uplift of mountain ranges alternated with periods of extension resulting in the collapse of their individual parts and basin formation. The plate tectonics and orogeny, along with the global sea level change influenced not only the palaeogeography of the area, but also controlled connections of this inland sea

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with the World Ocean across the Mediterranean or Eastern Paratethys (e.g., Hardenbol et al., 1998; Kováč et al., 1999; Sant et al., 2017).

The Central Paratethys' water masses and their circulation (salinity, temperature, estuarine, and anti-estuarine regimes, etc.) were heavily dependent on both global and local climatic changes. The exhumation of mountains often had an impact on precipitation, documented by dryer or more humid periods. Additionally, the Alpine mountain landforms led to the development of a latitudinal and altitudinal zonation (e.g., Kvaček et al., 2006). In some cases, these emerging high mountains represented a barrier to the movement of air masses, leading to

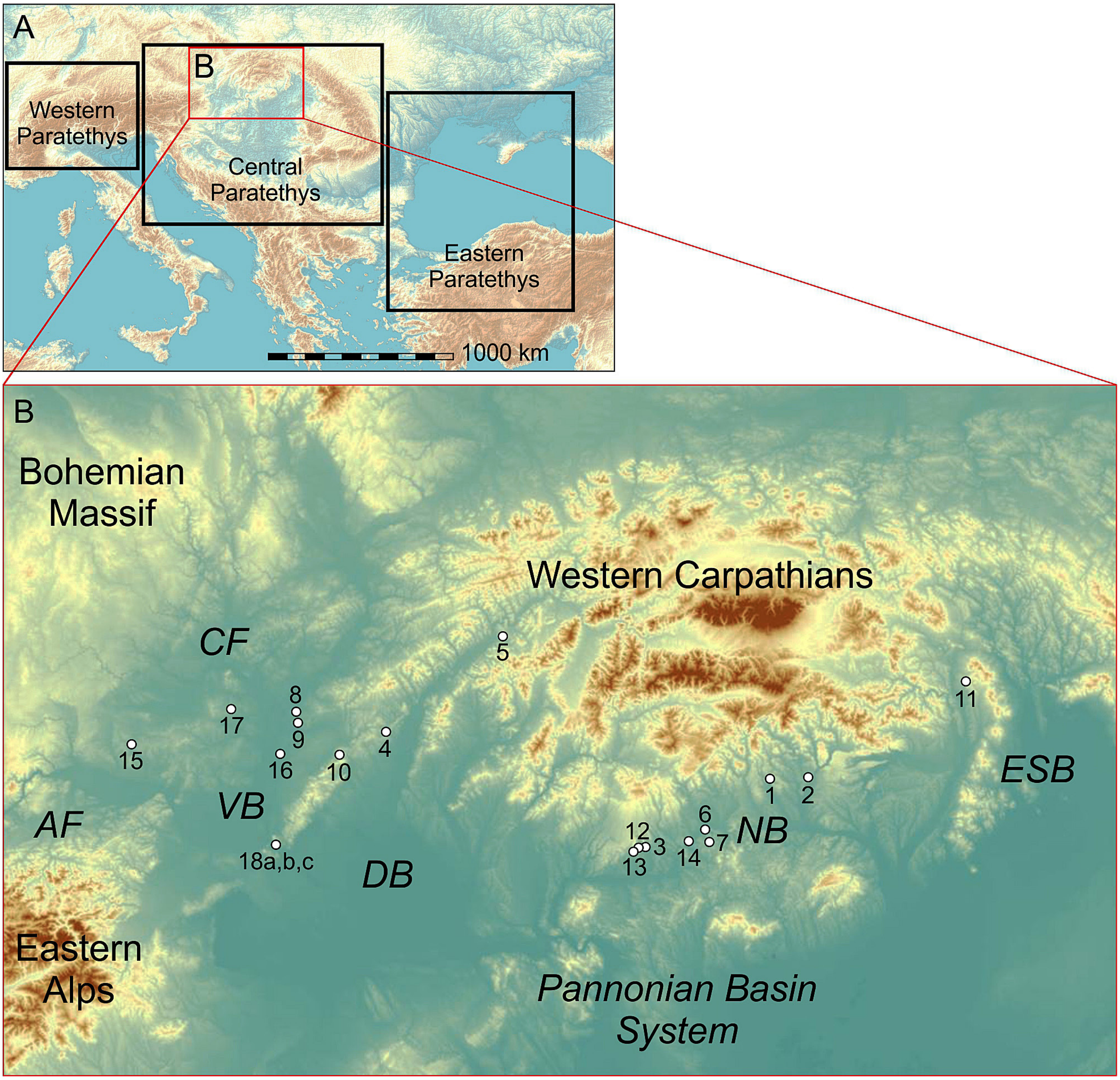


Fig. 1. (A) Position of the Western, Central, and Eastern Paratethys basins; (B) Sites with the Sr-dating.

Explanatory notes: **(1)** Budikovany; Novohrad-Nógrád Basin; Egerian; **(2)** Bretka; Novohrad-Nógrád Basin; Egerian; **(3)** Dolné Plachtince; Novohrad-Nógrád Basin; Egerian; **(4)** Podkylava; Váh River Valley; Eggenburgian; **(5)** Sverepec; Váh River Valley; Eggenburgian; **(6)** Rapovce; Novohrad-Nógrád Basin; Eggenburgian; **(7)** Lipovany; Novohrad-Nógrád Basin; Eggenburgian; **(8)** Cunín-21 borehole; Vienna Basin; Ottnangian; **(9)** Gbely-100 borehole; Vienna Basin; Karpatian; **(10)** Cerová–Lieskové; Vienna Basin; late Karpatian; **(11)** Mirkovce; Eastern Slovak Basin; Karpatian–Early Badenian; **(12)** Príbelce; Novohrad-Nógrád Basin; Early Badenian; **(13)** Čebovce; Novohrad-Nógrád Basin; Early Badenian; **(14)** Trenč; Novohrad-Nógrád Basin; Early Badenian; **(15)** Grund (Austria); Alpine Foredeep; Early Badenian; **(16)** Moravský Ján-3 borehole; Vienna Basin; Early Badenian; **(17)** Sedlec–HJ-2 borehole (Czech Republic); Vienna Basin; Early Badenian; **(18a)** Devínska Nová Ves–Sandberg; Vienna Basin; Early Badenian–Sarmatian; **(18b)** Devínska Nová Ves clay pit; Vienna Basin; early Late Badenian; **(18c)** Devínska Nová Ves–"Lingula Bed"; Vienna Basin; Early Badenian–Sarmatian; **(AF)** Alpine Foredeep; **(CF)** Carpathian Foredeep; **(DB)** Danube Basin; **(ESB)** Eastern Slovak Basin; **(NB)** Novohrad-Nógrád Basin; **(VB)** Vienna Basin.

a change in the local wind system causing coastal upwellings (e.g., Grunert et al., 2010[b]).

The main aim of our study was to refine the Cenozoic palaeogeography of the Central Paratethys Sea in selected time intervals. Additionally, we summarized the palaeoclimatic evolution and interpreted the existing data in order to propose a "model of water circulation" and their impact on the regional biostratigraphy. This attempt to present a palaeoceanographic reconstruction of a semi-closed inland sea is based on an analysis of the changing depositional environment in the Western Carpathian basins and neighbouring areas controlled by tectonics and reflected by climatic and microorganism proxies.

The presented model is created by compiling data from various disciplines of geology, sedimentology, palaeontology, and geochronology collected by many authors from various regions during the last decades. Time horizons (correlation levels) and time intervals used for the reconstruction of palaeogeography and water circulation regimes are defined by planktic foraminifera together with calcareous nannoplankton biostratigraphy, and supplemented by $^{87}Sr/^{86}Sr$ isotope measurements.

2. PREVIOUS RESEARCH, METHODS, AND MATERIALS

The presented palinspastic-palaeogeographic maps concern the extent of the land and water-covered areas, as well as the opening and closing of various marine connections (straits/gateways) among the basins known from previous publications (e.g., Steininger et al., 1976; Rögl et al., 1978; Rögl & Steininger, 1983; Steininger & Rögl, 1984; Rögl, 1998; Popov et al., 2004; Harzhauser et al., 2007; Kováč et al., 2007, 2016, 2017[a]; Palcu et al., 2015). The extent of the Central Paratethys Sea and its bottom topography had to be reconstructed using characteristic changes of the environment of sedimentary depositional systems; taking into account processes induced by the collision of the Western Carpathians with the European Platform, as well as the subduction of the Magura and Krosno basins, both resulting in the gradual growth of the Outer Carpathian accretionary wedge (e.g., Kováč et al., 1994, 1998, 2003, 2007, 2016, 2017[a]; Meulenkamp et al., 1996; Matenco, 1997; Oszczypko, 1998, 1999; Kováč, 2000; Konečný et al., 2002; Oszczypko et al., 2006, 2015; Schmid et al., 2008; Oszczypko & Oszczypko-Clowes, 2009, 2014).

The reconstruction of the marine palaeoenvironment in selected time intervals was based on a number of publications with available datasets from the Western Carpathians and adjoining areas in the past (Fig. 2) based on calcareous nannoplankton, foraminifers, organic-walled dinoflagellates, pollen spectra (Lehotayová, 1978, 1982, 1984; Nagymarosy, 1982; Szczechura, 1982; Zlinská & Šutovská, 1991; Holcová, 1996, 2001[a,b]; Oszczypko-Clowes, 1999, 2001, 2008, 2012; Peryt, 1999, 2013[a,b]; Spezzaferri & Ćorić, 2001; Mandic et al., 2002; Oszczypko & Oszczypko-Clowes, 2002, 2003, 2006, 2009, 2014; Rögl et al., 2002, 2007; Spezzaferri et al., 2002, 2004, 2009; Andreeva-Grigorovich, et al., 2003; Bartakovics & Hudáčková, 2004; Ćorić et al., 2004; Gedl, 2004; Oszczypko-Clowes & Oszczypko, 2004, 2011; Spezzaferri, 2004; Kováčová & Hudáčková, 2005; Báldi, 2006; Tomanová Petrová & Švábenická, 2007; Báldi & Hohenegger, 2008; Kováč et al., 2008; Ozdínová, 2008[a,b]; Tóth & Görög, 2008; Zágoršek et al., 2008; Kováčová et al., 2009; Kováčová & Hudáčková, 2009; Grunert et al., 2010[b], 2013; Hudáčková & Zlinská, 2010; Jamrich & Halásová, 2010; Koubová & Hudáčková, 2010; Peryt & Gedl, 2010; Gedl & Peryt, 2011; Kopecká, 2012; Zlinská et al., 2010; Gonera, 2013; Bitner et al., 2014; Doláková et al., 2014; Ozdínová & Soták, 2014; Peryt et al., 2014; Holcová et al., 2015[a,b]; Kaczmarek et al., 2016; Nehyba et al., 2016), and VIENNDAT dataset (Hudáčková & Hudáček, 2004). Taphonomic analyses of foraminiferal assemblages were identified and evaluated according to methods described by Holcová (1997, 1999). Calcareous nannoplankton assemblages served as an excellent proxy for high-resolution palaeoenvironmental reconstructions (Grimalt & Lopez, 2007; Auer et al., 2014). Oxygen and nutrient content, salinity, temperature, and water energy of surface and bottom water were estimated using actuoecological data as well (e.g., Okada & McIntyre, 1979; Hemleben et al., 1989; Jorissen et al., 1992, 1995, 1998; Sjoerdsma & van der Zwaan, 1992; Gooday, 1993; Knappertsbusch, 1993; Rohling et al., 1993; Sen Gupta & Machain-Castillo, 1993; Kaiho, 1994; Wells & Okada, 1997; Schönfeld, 2001; Fontanier et al., 2002; Kameo, 2002; Schiebel & Hemleben, 2005; Murray, 2006; Kuhnt et al., 2013). Water depth was estimated using the methodology of Hohenegger (2005). We also studied the principles of origin and changes of marine currents, causes, and consequences identified by the fossil assemblages (Grimalt & Lopez, 2007; Saraswat & Nigam, 2013). The analysis of sedimentary records helped us to detect the direction of tidal currents (e.g., Vass et al., 1979, 2007; Faupl & Roetzel, 1990; Sztanó, 1994, 1995; Sztanó & de Boer, 1995; Bieg et al., 2008), the formation of the river and deltaic systems, and the deposition of coal or sedimentation of evaporites (Vass et al., 1979, 2007; Kováč et al., 1995, 2004, 2011; Andreyeva-Grigorovich et al., 2003[b]; Bukowski et al., 2003, 2007, 2010).

With the aim of compiling relevant data on palaeoceanography of the Central Paratethys, extended for the adjoining Mediterranean and Eastern Paratethys realms, a rigorous search was carried out (von Daniels & Ritzkowski, 1970; Steininger et al., 1976; Rögl et al., 1978; Rögl & Steininger, 1983; Steininger & Rögl, 1984; Por & Dimentman, 1985; Vetö, 1987; Martini, 1990; Rögl, 1998; Gebhardt, 1999; Seidenkrantz et al., 2000; Schulz et al., 2005; Kouwenhoven & van der Zwaan, 2006; Harzhauser et al., 2007; Kováč et al., 2007; Karami et al., 2011; Grunert et al., 2014; Palcu et al., 2015; Vasiliev, 2015). Water circulation regimes have been previously studied in detail in the Mediterranean (e.g., Gebhardt, 1999; Seidenkrantz et al., 2000; Kouwenhoven & van der Zwaan, 2006), while in the Paratethys, the studies were restricted mostly to discussing the connections and gateways among the Paratethys and nearby seas (von Daniels & Ritzkowski, 1970; Rögl et al., 1978; Rögl & Steininger, 1983; Steininger & Rögl, 1984; Por & Dimentman, 1985; Martini, 1990; Rögl, 1998; de la Vara, 2015; Palcu et al., 2015). However, we have found several papers also dealing with water circulation regimes (Fig. 3) in which the Oligocene palaeoceanographical aspects were discussed by Krhovský (1981[a,b]), Brukner-Wein et al. (1990), Krhovský et al. (1992), Oszczypko-Clowes (2001), Pekar et al. (2002), Schulz et al. (2005), and the Miocene by Kókay (1985), Brzobohatý (1987), Bicchi et al. (2003), Báldi (2006) and Kováčová et al. (2009). Moreover, the Paratethys mollusc extinction and build-up-events (Harzhauser, 2005) confirm the significant impact of the palaeogeography and water circulation regimes on the formation of the associations of marine biota.

The climate, as an important controlling factor for every aspect of changing precipitation (arid or humid climate), contributed to the water circulation changes in individual basins of the Paratethys. To estimate the mean annual temperature (MAT), the mean annual precipitation (MAP), the coldest month mean temperature (CMMT) in the Western Carpathians as well as changes in landforms and climate of the adjoining continental areas of the Central Paratethys, works on fossil flora were compiled (Ivanov et al., 2002, 2014; Bruch & Kovar-Eder, 2003; Erdei & Bruch, 2004; Bruch et al., 2006; Kvaček et al., 2006; Erdei et al., 2007; Kroh, 2007; Syabryaj et al., 2007; Utescher et al., 2007; Jiménez-Moreno et al., 2008; Kučerová, 2010; Kováčová et al., 2011; de la Vara et al., 2013; Doláková et al., 2014; Grunert et al., 2014). Temperature estimates were obtained by using the Coexistence Approach (e.g., Mosbrugger & Utescher, 1997; Mosbrugger et al., 2005) and the Climate Leaf Analysis Multivariate Program (CLAMP). Terrestrial palaeoclimate proxy data were compared with the global marine oxygen isotope records of Zachos et al. (2001) adapted to the time scale of the International Commission on Stratigraphy (Gradstein et al., 2012; Ogg et al., 2016).

3. BIOSTRATIGRAPHY, SR-ISOTOPE DATING, TIME CORRELATION LEVELS AND TIME INTERVALS

The Cenozoic stratigraphy used in the time span from the Early Oligocene to Middle Miocene (Fig. 2) was based *(i)* on standard biostratigraphical events from the World Ocean (Martini, 1971; Berggren et al., 1995; Wade et al., 2011; Gradstein et al., 2012); *(ii)* on Mediterranean biostratigraphical events during communication of the Central Paratethys with the Mediterranean Sea (Fornaciari & Rio, 1996; Fornaciari et al., 1996, 1997[a,b]; Hilgen et al., 2005; Abdul Aziz et al., 2008; Iaccarino et al., 2011); *(iii)* on stratigraphical ranges of the endemic Paratethyan taxa (Mărunteanu, 1992, 1997; Cicha et al., 1998; Kováč et al., 2008). In order to range stratigraphically important taxa, the current status of the stratigraphical correlation between the Central Paratethys regional stages and the Mediterranean scale summarized by Kováč et al. (2007), Piller et al. (2007) and Hohenegger et al. (2014) was used. The stratigraphy of the Late Miocene strata was based on organic-walled dinoflagellate cysts (*sensu* Balteş, 1967, 1969; Hochuli, 1978; Sütő-Szentai, 1985, 1988, 1989, 1991, 2005; Sütő, 1994; Jiménez-Moreno et al., 2006; Bakrač et al., 2012), rare calcareous nannoplankton, and mammals (e.g., Kováč et al., 2006, 2011; Hilgen et al., 2012; Daxner-Höck & Höck, 2015), as well as on Be-isotope data (Šujan et al., 2016).

Additionally, Sr-isotope dating results from the selected sites were taken into consideration (Fig. 1). We were aware of the fact that the stratigraphic resolution of the method based on the global changes in the $^{87}Sr/^{86}Sr$ value of the open ocean through geological time (e.g., Palmer & Edmond, 1989; Andersson et al., 1992; Ravizza & Zachos, 2003; Allègre et al., 2010) is ultimately limited by the slope of the reference curve determined by the rate of change of marine Sr-isotope ratio, and by its accuracy expressed by the width of a 95 % confidence interval on the calibration curve (Ogg et al., 2016). Within these limits, the method helped us to solve the problem of correlation across different palaeobiogeographic provinces and between different marine palaeoenvironments (shallow to deep water). At the same time, it facilitated the calibration of biostratigraphic scales based on fossils that are rarely found together in the same rocks (e.g., planktic foraminifera, large benthic foraminifera, and nannoplankton). In the presented study, in addition to the existing data, samples of pristine, unaltered marine precipitates with preserved original Sr-isotope ratio of sea-water (McArthur, 1994; Veizer et al., 1997), and samples in the form of low Mg biotic calcite of compact shells were used.

To correlate the studied Western Carpathian basins´ fill with the sedimentary record of neighbouring marine realms of the Mediterranean and Eastern Paratethys, a definition of well-specified time levels was needed (Fig. 2). The correlation levels had to be recognizable in the studied sedimentary record based on standard biostratigraphical markers (Gradstein et al., 2012). The First Occurrence (FO), First Common Occurrence (FCO), Last Occurrence (LO) and Last Common Occurrence (LCO) of planktic foraminifera, calcareous nannoplankton and dinocysts were used (Martini, 1971; Sütő-Szentai, 1988; Berggren et al., 1995; Wade & Bown, 2006; Wade et al., 2011; Bakrač et al., 2012; Gradstein et al., 2012; Ogg et al., 2016). Six time correlation levels were allocated: *(i)* Oligocene/Miocene boundary (late Egerian; ~23 Ma); *(ii)* Aquitanian/Burdigalian boundary (early Eggenburgian; ~20.45 Ma); *(iii)* Middle Burdigalian (Ottnangian; ~18 Ma); *(iv)* Langhian (Early Badenian; ~15 Ma); *(v)* early Serravallian (Late Badenian; ~13.5 Ma); and *(vi)* the early Tortonian (early Pannonian; ~10.5 Ma). Most of these time levels/horizons were dated by previously published as well as new Sr-isotope data (Hudáčková et al., 2003; Fordinál et al., 2014; Less et al., 2015).

Estimation of the extent of the Central Paratethys realm was a very complicated task because of the geodynamic development of the area (basins and mountain chains evolution). Therefore, we selected broader time intervals which allowed us to roughly compare the time-consistent sedimentary sequences deposited in similar or different environments in the Western Carpathian basins with the sedimentary fill of neighbouring areas belonging to the rest of the Central Paratethys Sea. Moreover, those time intervals are shown in palinspastic-palaeogeographical maps (Figs. 4–10).

Time intervals dealt with in the following text (see Chapter 5), are biostratigraphically defined and supplemented by Sr-isotope dating carried out at the type localities. Their description contains comments on time-related palaeogeography, sea bathymetry, and interpretation of the possible

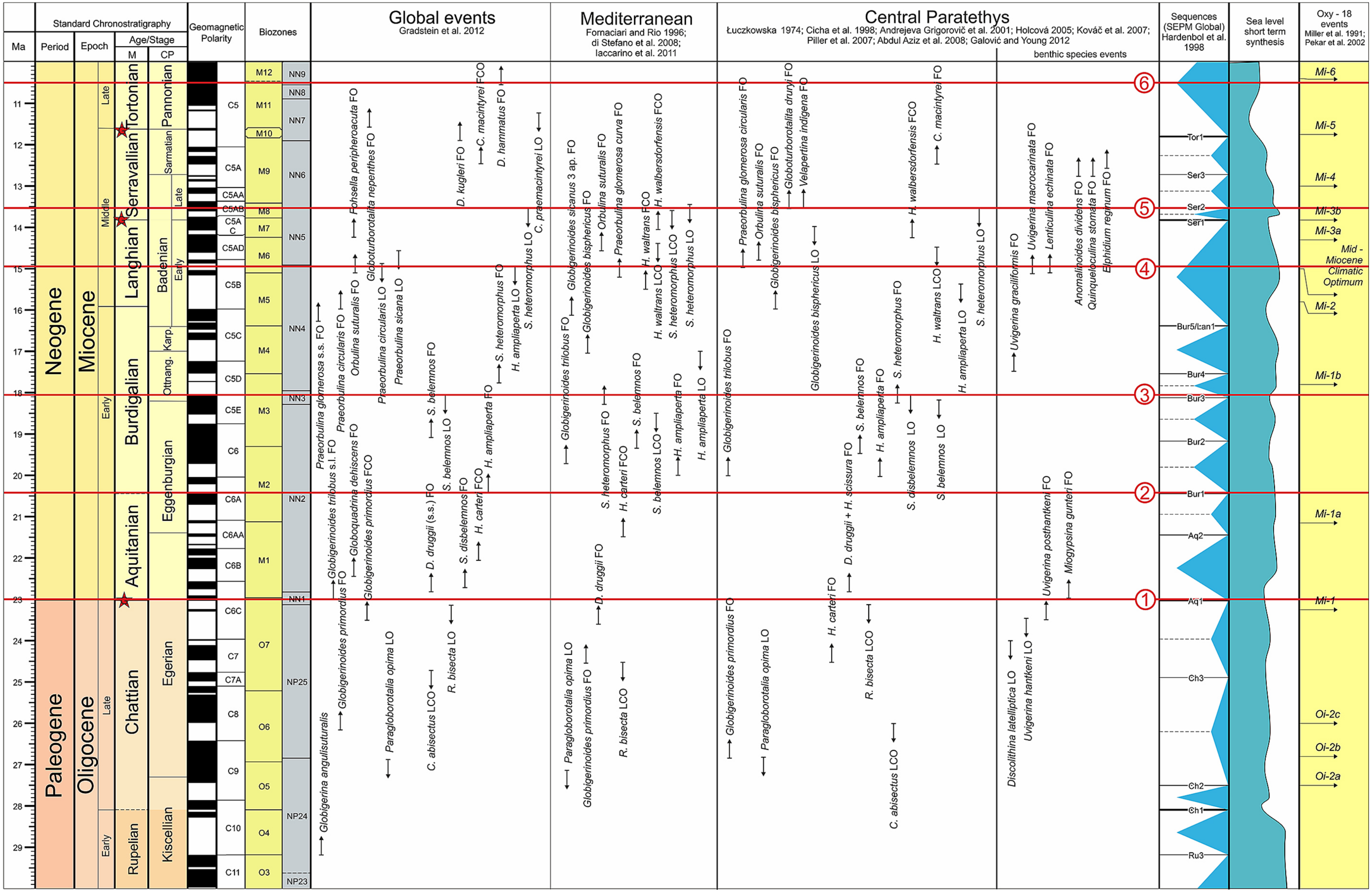


Fig. 2. Correlation chart: Standard planktic and benthic organism's biostratigraphy of the World Ocean, Mediterranean, and Central Paratethys. Time levels recognized in the studied sedimentary record are marked by red lines 1–6. *Explanatory notes:* **(asterisk)** Global Boundary Stratotype Section and Point; **(CP)** Central Paratethys; **(M)** Mediterranean. Chronostratigraphy, geomagnetic polarity, biozones, sequences and sea level short term synthesis and oxy-18 events according to Time Scale Creator (modified after Harzhauser & Piller, 2007; Ogg et al., 2016).

palaeoceanographical regime. Correlation of the local and global sequence stratigraphy of the 3rd order sea level change was evaluated after Hardenbol et al. (1998) and Kováč (2000).

4. REGIONAL PALAEOCLIMATE EVOLUTION

During the Rupelian (Kiscellian), in the southern realm of the Central Paratethys, at the foothill of the uplifting Eastern Alps, Transdanubian Range, and the pre-Neogene basement of the Danube Basin, the sea coast of the retro-arc basin was covered by swamps with thermophilous elements (e.g., Cupressaceae/Taxodioidea, Myricaceae, Nyssaceae, Mastixiaceae and *Sabal* palms). Mixed mesophytic forest with *Carpinus* and *Tsuga* indicates local cooling (Vass et al., 1979; Planderová, 1990; Erdei et al., 2012). The Chattian (Egerian) general palaeoclimate evolution in Central and Eastern Europe shows a temperature peak corresponding to the global warming known from the oceanic isotope record (Zachos et al., 2001).

The Egerian-Eggenburgian time span yielded an ideal environment for flora preferring warm and humid climate. On the sea coast of the Hungarian-Slovenian retro-arc Basin this climate is documented by a high ratio of the tropical sporomorphs and macroflora in the hinterland of the Alpine-Western Carpathian mountain chain (Nagy & Pálfalvy, 1963; Šikić, 1966, 1968; Nagy, 1979, 1992, 1999, 2005; Planderová, 1990; Kovar, 1982; Erdei & Bruch, 2004; Kováčová & Sitár, 2007; Kováčová et al., 2011). Humid tropical forests with *Triplanosporites*, *Cyathidites*, *Saxosporites* ferns and *Triatripollenites excelsus*, *Slowakipollis*, *Engelhardia* and evergreen Fagaceae covered the slopes of the mountains (Vass et al., 1979; Planderová, 1990).

After the Mi-1 cooling event (~ 23 Ma), the changes of precipitation in the southern realm of the Central Paratethys are indicated by xerophytic taxa, together with the appearance of deciduous trees in palynofloras (Planderová, 1990; Nagy, 1999, 2005; Erdei et al., 2007) from the northern coast of the retro-arc basin (southern Slovakia and northern Hungary area). The identical vegetation type is also described from the northern part of Dinarides (Báldi & Seneš (Eds.), 1975), representing during this time interval the southern margin of the same retro-arc basin (Fodor et al., 1998). The warming had its imprint on the local climate, noted in the marine plankton by increasing of warm-water planktic foraminiferal and calcareous nannoplankton taxa (Ozdínová & Soták, 2014). Appearing of hyposaline benthic foraminiferal assemblages indicates seasonal precipitation increase as well (Holcová, 2017).

The high percentage of tropical taxa does not necessarily indicate a tropical climate since most tropical taxa live under humid subtropical climates as well. On the contrary, the areas on the eastern margin of the Central Paratethys Sea, with decreasing percentages of the laurel forest-type vegetation, indicate a slight aridification of climate (Syabryaj et al., 2007). This fact can suggest a strong influence by the dry continental climate of the Volhynian High and the adjoining areas of the Eastern European Platform (Popov et al., 2004).

During the Early Miocene, an overall trend of increasing temperature is observed, persisting into the Middle Miocene. However, this warming seems to be more stepwise, and the curves show several short-term cooling (Planderová, 1990), which can be correlated with the Mi cooling events (Miller et al., 1991). In the Burdigalian (Eggenburgian–Ottnangian–Karpatian) a large number of thermophilous elements suggests a humid subtropical climate. The estimated parameters indicate MAT to be around 18–20°C and MAP between 1200 and 1300 mm based on the Ottnangian flora of the Oberdorf Mine in the Styrian Basin (Meller, 1998; Meller et al., 1999; Kovar-Eder et al., 2001; Bruch & Kovar-Eder, 2003) which coincides with MAT values of 16.5–18.8°C calculated by Hably (1985) and Erdei et al. (2007) on macroflora from northern Hungary. The Eggenburgian–Ottnangian flora at the southern foothill of the Western Carpathians is characterised by thermophilous species (not tropical). The absence of swampy and riverside elements in the fossiliferous strata suggests that the vegetation thrived under a humid subtropical climate (Sitár & Kvaček, 1997; Kučerová, 2009). In contrast to that, at the coeval locality of Ipolytarnóc situated more to the south (Hably, 1985; Mai, 1995), the quantitative estimates expect MAT of less than 20°C, which would mean tropical-paratropical conditions. These more realistic values would vary between 15–18°C, with CMMT above 1°C. The discrepancy suggests an altitudinal difference among the aforementioned sites, due to the uplift of the neighbouring Western Carpathians, representing during this time interval the coasts of the Central Paratethys southern realm.

The Early Miocene uplift of the axial part of the Eastern Alps and partly also the axial portion of Central Western Carpathians led to the formation of a natural barrier for winds from the south. At the foothill of the mountain range, coal deposition began in swamps environment (Ottnangian/Karpatian) in the Styrian and Novohrad basins (e.g., Vass et al., 1979, 1999; Vass & Elečko (Eds.), 1989; Sachsenhofer, 1996). In the Novohrad Basin, such type of sedimentation on the coast of the sea lasted until the beginning of the Middle Miocene (Early Badenian; Ádám, 2006).

From the eastern margin of the Central Paratethys, documented Lower Miocene pollen spectra refer to MAT ranging between 16–17.5°C, therefore indicating Burdigalian warming as well (Syabryaj et al., 2007). The Eggenburgian–Ottnangian mixed coniferous and broad-leaved forests with subtropical species of the Juglandaceae, sclerophyllous oaks and chestnut forests were most common. These forests are similar to extant evergreen forests growing at present in the lowland areas of the Azores. Later, during the Ottnangian–Karpatian, swamp forests occupied smaller regions of the Carpathian Foredeep Basin located in front of the growing accretionary wedge of the Outer Carpathians, with altitudinal zonality and habitats differentiation (e.g., Syabryaj et al., 2007).

The Middle Miocene climate is characterised by the global oxygen curve with a maximum caused by reduced ice volume and a warming trend of the deep sea-water (Zachos et al., 2001). The continental record shows a similar pattern with temperature rapidly increasing from 17 Ma (Utescher et al., 2002). The succeeding warm time span persisted through the whole Langhian (Early Badenian), up to the early Serravallian

(Late Badenian), corresponding to the globally observed Mid-Miocene Climatic Optimum (e.g., Piller & Harzhauser, 2005; Harzhauser et al., 2007). However, records of CMMT indicate regional differences for the first time. Very high CMMTs of 9–13°C are calculated for flora from the platform regions surrounding the Central Paratethys Sea (e.g., Lausitz and Lower Rhine basins, and Ukraine; Utescher et al., 2000, 2002, 2012; Syabryaj et al., 2007; Doláková et al., 2014). In the Central Western Carpathians, the macroflora from the Nováky coal mine in the Horná Nitra Basin documents CMMT at 7.9°C (Takáč, 1966; Sitár, 1976; Kučerová, 2010), which we assume is an effect of the uplifted mountain ranges and rising volcanoes surrounding the swamps. Similarly, in the case of the Alpine Molasse Basin, orographic differences caused by the uplift of the Alps are reflected in a palaeoclimate with the onset of the Miocene cooling between 13.0 and 14.0 Ma, and a rapid decrease of CMMT at the end of Middle Miocene, when CMMT dropped below 4°C (Mosbrugger et al., 2005). Development of a high mountain relief in the Central Western Carpathians is documented by the Badenian pollen spectra composition, presented in the studies by Sitár & Kováčová-Slamková (1999) and Kvaček et al. (2006).

In contradiction to the early Serravallian (Late Badenian), when palaeoflora documents a humid climate (mangrove and swamp habitats) in the southern realm of the Central Paratethys, the overlying late Serravallian (Sarmatian) strata indicate clear evidence of aridification. It is documented by xerophytes and halophytes in the pollen spectra from northern Hungary during this time (e.g., Nagy, 2005). The occasional hypersaline facies (Piller et al., 2007) which deposited in the Vienna and Danube basins, probably originated due to local rain shadows. A decrease in humidity was identified not only from the southern realm of Central Paratethys, but also from the Alpine Foredeep (Böhme & Vasilyan, 2014).

Along the eastern margin of the Central Paratethys the Late Badenian vegetation is characterised by humid coastal and riparian forests with thermophilous ferns (Syabryaj et al., 2007). Within the Sarmatian, a change was registered. Along the seashore, fresh-water lagoons formed with the accumulation of peat and were followed by the development of swamp forests with abundant *Taxodium*. Altitudinal zonality and high mountain relief of the Carpathian mountain chain is marked by pollen from coniferous forests with spruce and spruce-fir assemblages together with various *Tsuga* species (Syabryaj et al., 2007). The lower Pannonian (lower Tortonian) vegetation reflect dominant swamp habitats developed in a subtropical and humid climate predominantly within the Pannonian Basin System (Kvaček et al., 2006; Erdei et al., 2007; Doláková & Kováčová, 2008; Kováčová et al., 2011; Utescher et al., 2017).

5. THE CENTRAL PARATETHYS PALAEOGEOGRAPHY AND WATER CIRCULATION REGIMES

After its birth, the Central Paratethys Sea covered extensive areas in front of, as well as behind the Western Carpathian orogen. In the Oligocene, the northern realm of this sea (fore-axis basin system; *sensu* Kováč et al., 2016) covered areas from the European Platform in the north to the uplifted axial part of the Central Western Carpathians. The basin system included the European Platform shelf and its slopes (passive margin), the remnant flysch troughs with oceanic or thinned continental crust, and a variety of basins located along an actively deforming margin of the overriding plate. Towards the southwest, a retro-arc basin was located behind the uplifting axial part of the orogen (back-axis basin system; *sensu* Kováč et al., 2016) and represented the southern realm of the Central Paratethys. Starting with the Miocene, the palaeogeography was significantly changed. The northern realm was restricted to the Carpathian Foredeep situated on the platform margins and to several piggy-back basins on top of the Outer Carpathians accretionary wedge. However, in the southern realm of the Central Paratethys, several pull-apart basins and extensional hinterland back-arc basin system were developed, characterised by their initial rifting, synrift, and thermal postrift stages of the evolution (Kováč et al., 2017[a]).

Geodynamic processes that significantly influenced palaeogeography of the Central Paratethys Sea primarily controlled the connection of the sea with the World Ocean, with the partial influence of the sea level change. During the Cenozoic, several occasional gateways/straits developed and disappeared between the Mediterranean and Eastern Paratethys (e.g., Rögl, 1998; Kováč, 2000; Popov et al., 2004; Sant et al., 2017). If the gateway reached a depth of less than 1 000 m, then the circulation regime was influenced by local climate (de la Vara, 2015). The opening and closure of these gateways had such a considerable impact on the exchange of water masses, but also on the possible immigration of marine organisms from surrounding basins. On a regional scale, closure favoured a nearly landlocked position, resulting in a decrease of the basin's buffering capacity and increasing its sensitivity to climate changes (Gladstone et al., 2007). Thus, the isolation often led to deposition of evaporites, anoxic events, or in the case of reducing salinity led to the development of a brackish environment with specific endemic taxa.

Assuming that through geological time the volume of the Earth's water body is more or less stable, communication of water masses between a smaller inland sea-water body and a larger water body of the neighbouring seas (Mediterranean and Eastern Paratethys) can be principally two ways, in most cases depending on the local climate. Based on the balance between evaporation and precipitation, the following thermohaline circulation regimes can be distinguished if the seaways are deep enough:

(i) Anti-estuarine regime (e.g., the present-day Mediterranean) occurs when evaporation exceeds the fresh-water input from precipitation and river discharge. The surface water salinity might increase enough to make it heavy, and the relatively more saline water sinks to the bottom of the basin and flows out at the strait connecting the basin with the neighbouring sea/ocean. The great mass of deep water leaving our inland basin (Central Paratethys) has to be replenished by surface water entering from the, e.g., Mediterranean or Eastern Paratethys.

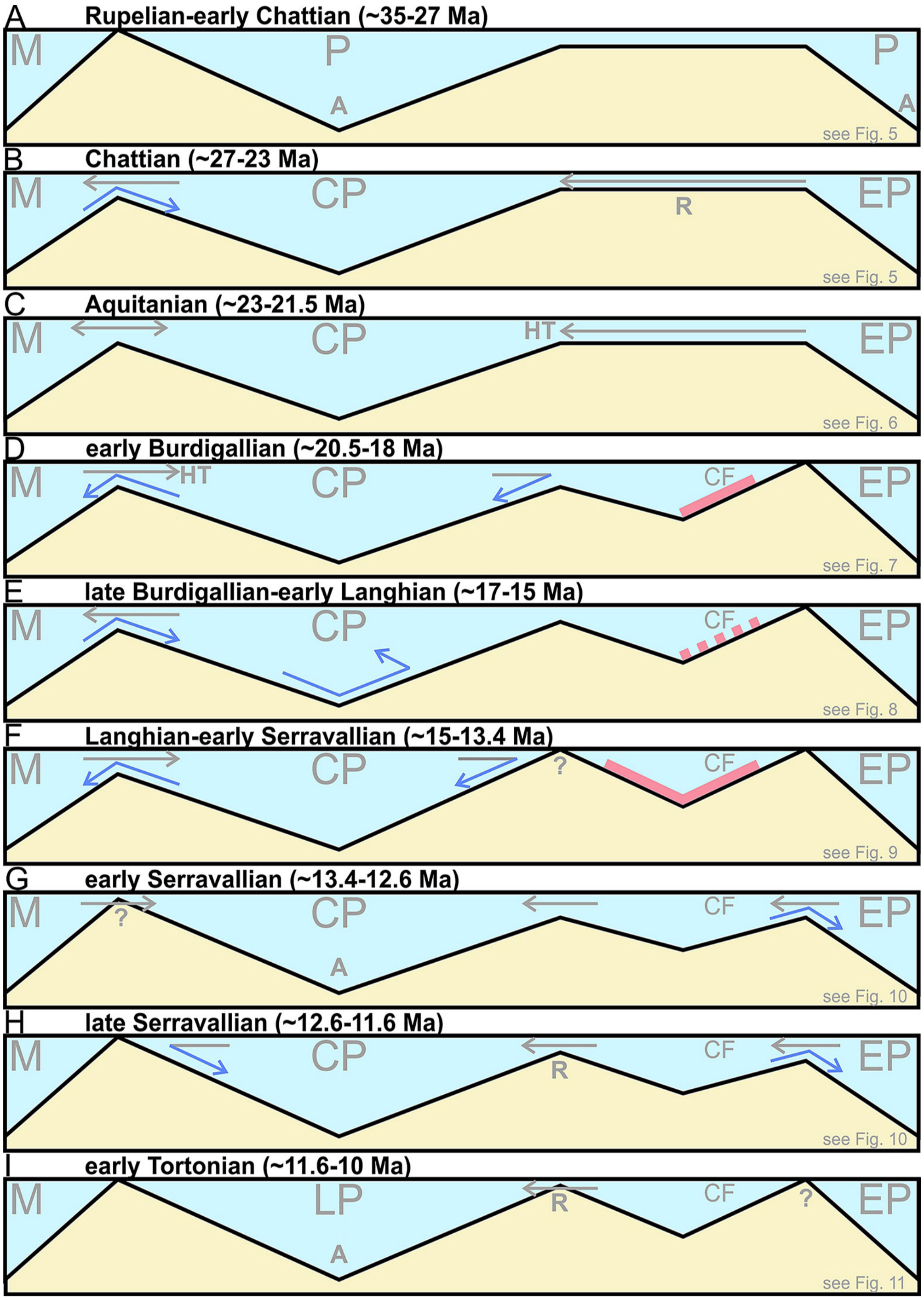


Fig. 3. Scheme of the Cenozoic Central Paratethys circulation regimes (A–H). **(A)** Rupelian–early Chattian interval – euxinic regime; **(B)** Chattian interval – estuarine regime; **(C)** Aquitanian interval – open marine regime; **(D)** early Burdigalian interval – anti-estuarine regime; **(E)** late Burdigalian–early Langhian – estuarine regime; **(F)** Langhian–early Serravallian interval – anti-estuarine regime; **(G)** early Serravallian interval – low oxic with occasional anti-estuarine regime from the east; **(H)** late Serravallian interval – occasional anti-estuarine regime from the east; **(I)** early Tortonian interval – basin isolation. *Explanatory notes:* **(A)** bottom anoxia; **(CF)** Carpathian Foredeep; **(CP)** Central Paratethys; **(EP)** Eastern Paratethys; **(HT)** tidal currents; **(LP)** Lake Pannon; **(M)** Mediterranean; **(P)** Paratethys; **(R)** restricted connection; **(blue arrow)** deep-water currents; **(grey arrow)** surface-water currents; **(pink line)** evaporites; **(pink dashed line)** semi-arid conditions with local salt deposition. For details see Figs. 4–10.

In the case of the absence of evaporites, we might assume that the deep water is not a hypersaline brine, but relatively more saline or colder. Therefore, this circulation enables the vertical motion of the water and keeps the bottom well-oxygenated. In the extreme case of hypersaline deep water supersaturated by salts, deep water evaporites can form.

(ii) Estuarine regime (e.g., the present-day Baltic Sea and the Black Sea) occurs when a fresh-water influx in the basin exceeds evaporation, due to a humid climate and/or rivers. A layer of less-saline surface water will flow out of the inland sea, and will be replaced by normal saline deep water from the World Ocean (Stewart, 2008). In the case of this circulation, a strong halocline can form, restricting vertical mixing and leads to a deficiency of oxygen at the bottom of the sea. If communication is hampered, or even total isolation occurs, as long as fresh-water influx exceeds evaporation, similarly to estuarine regime, a less saline surface water layer will form with enhanced halocline. Restriction of vertical mixing leads to bottom water hypoxia or anoxia results in the sedimentation of the euxinic facies. The incoming cold and high-nutrient deep water masses from the ocean, if sill depth allows the entering of such deep water, might well up in case of appropriate winds

and coastal configuration during an upwelling regime. From the Central Paratethys, this regime is well-documented by Roetzel et al. (2006) and Grunert et al. (2010[b]) from the middle Burdigalian (Ottnangian).

The regimes discussed above can develop if the strait is deep enough for two-way communication between the Central Paratethys and the adjoining basins. In this case, water masses can in-flow from one basin and out-flow to another one. The necessary water depth of the strait also influenced the high tide regime, which can be widely documented in periods with restricted shallow platforms along the coastal line. A high-tide regime supported the existence of surface currents from the basin centre to its coastal parts. During the Cenozoic, all of the aforementioned circulation regime types could alternate several times (Fig. 3), resulting in changes of environmental conditions for deposition and marine life, expressed as sedimentary facies and in their fossil content.

5.1. The Late Oligocene

The Paratethys Sea came into existence during the Early Oligocene (Fig. 4). This event is associated with a complete isolation of this intra-continental sea from the World Ocean (Báldi, 1980; Rusu, 1988; Rögl, 1999). The water column stratification in the Hungarian-Slovenian Palaeogene Basin, in the southern realm of the Central Paratethys, was documented by Brukner-Wein et al. (1990), in the same way as Schulz et al. (2005) regard the fixation of nitrogen and low-organic carbon accumulation rates in the northern realm of Central Paratethys. In both settings, the vertical water stratification was

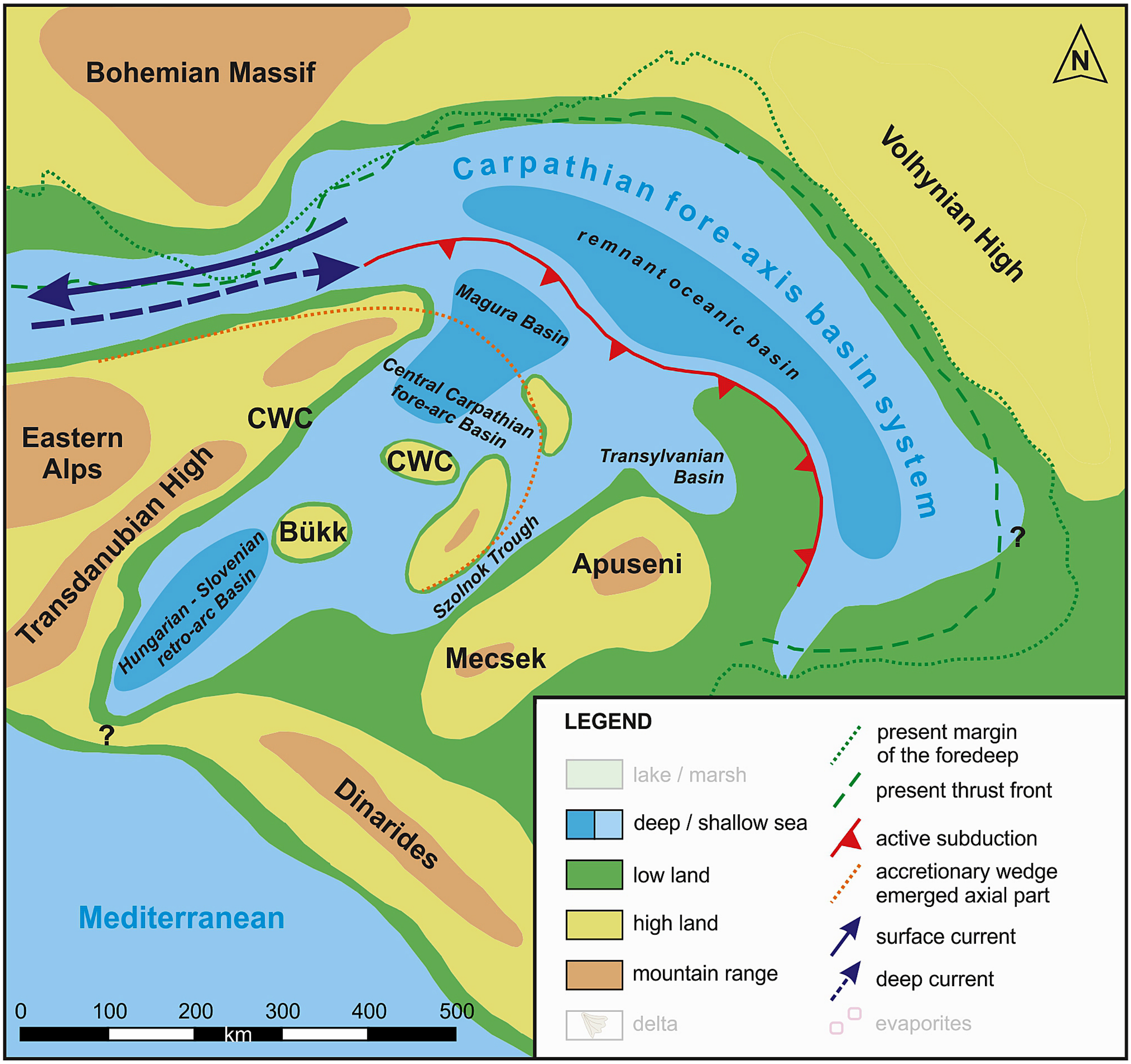


Fig. 4. Palaeogeography of the Central Paratethys: late Rupelian–Chattian. ***Explanatory notes:*** **(CWC) Central Western Carpathians; (?) assumed short living sea connection.**

accompanied by bottom anoxia and by anaerobic methane oxidation (e.g., Bessereau et al., 1996). The low oxic-euxinic conditions (Figs. 3A) on the sea floor reflected the restricted circulation (Tari et al., 1993; Krhovský, 1995; Leszczyński, 1997; Oszczypko-Clowes, 2001; Schulz et al., 2005; Soták, 2010) and can be traced in the Central and Eastern Paratethys (Chert Member and Dynów Marl of the Menilite Formation in the Ždánice Unit and Pouzdřany Unit, see Krhovský, 1981[a,b]; Krhovský et al., 1992; Bessereau et al., 1996; Oszczypko-Clowes, 2001, 2008; Oszczypko-Clowes & Oszczypko, 2004, 2011; Oszczypko-Clowes & Żydek, 2012; Studencka et al., 2016; or the Polbinian Horizon, see Nagymarosy & Voronina, 1992; Nagymarosy, 2000). The global Oligocene cooling was accompanied by lowering of CCD (Gradstein et al., 2012).

The characteristic feature of the Oligocene palaeogeography was the extensive flooding in front of the developing Western Carpathian orogen (Fig. 4). During this time, prominent geodynamic and palaeogeographic changes took place, especially in the basins of the future Outer Carpathians, which were transformed from a remnant oceanic basin into a flexural foreland basin (Oszczypko, 1999). The remnant oceanic basin (flysch troughs) covered the area represented mainly by the Krosno–Moldavides realm with a deep water sedimentary environment. Not only the active trench in front of the accretionary wedge (built by the Magura fold and thrust belt) but also a great part of the Central Western Carpathian margin was covered by the sea, where turbidite currents filled the Magura Basin and the Central Carpathian Palaeogene fore-arc Basin (Fig. 4).

In the hinterland of the Eastern Alps and Central Western Carpathians, the push of the Adriatic plate led to shortening and piling up of the Southern Alpine nappes. These processes are associated with the Kiscellian rapid deepening of the sedimentary environment in the Hungarian–Slovenian Palaeogene retro-arc Basin, in the Central Paratethys southern realm (Tari et al., 1993). The retro-arc basin with deep water turbidite sedimentation was still connected with the Carpathian fore-axis basin system across the partly flooded Central Western Carpathians, and probably along the Szolnok flysch Trough (Fig. 4).

Increased run-off, starting during the Rupelian nannoplankton zone NP 22 (Krhovský, 1981[a,b]; Krhovský et al., 1992; Oszczypko-Clowes, 2001), led to a distinct drop in salinity of the semi-closed Central Paratethys during the NP 23 Zone. The sea represented a palaeobiogeographical province defined by a brackish-water environment and by the appearance of first endemic assemblages, belonging to the Kiscellian stage. The appearance of new nannofossil species: *Reticulofenestra lockeri*, *R. ornata*, *Pontosphaera latelliptica*, *P. latoculata*, *Transversopontis fibula*, as well as the presence of another tolerant species *Cyclicargolithus floridanus - abisectus, Reticulofenestra bisecta*, and the blooms of *R. lockeri*, *R. ornata*, *Braarudosphaera bigelowii* reflect this event (Báldi-Beke, 1984; Nagymarosy, 2000; Švábenická et al., 2007; Švábenická, 2014).

Open-marine, calcareous nannofossil assemblages developed again near the NP 23/24 boundary. The NP24 Zone is characterised by the presence of species *Cyclicargolithus abisectus*, *Helicosphaera recta*, *Reticulofenestra lockeri* and *Sphenolithus distentus* (Nagymarosy, 2000; Garecka, 2005; Oszczypko-Clowes, 2008; Oszczypko-Clowes & Żydek, 2012). Change of the colour of fine sediments from dark to light, as well as the change of the fossil record composition, suggests the heterochronous change of circulation regime from stratified to mixed water during this time (Figs. 3B and 4).

The Oligocene part of the regional stage Egerian comprises the time span mostly of the NP25 and a part of NN1 zones (Piller et al., 2007). Generally, the correlation with standard planktic foraminiferal and nannoplankton zones is difficult, especially around the Oligocene/Miocene boundary because (many) index species did not migrate to the Paratethys (Rögl, 1998). However, the large benthic foraminifera biostratigraphical correlation can be used (Fig. 2).

After total isolation, since the Late Oligocene, an occasional communication with the Dinaridic Foredeep within the Mediterranean domain is assumed (Kováč et al., 2016). In the southern realm of the Central Paratethys Sea (Fig. 4), dark grey claystones were substituted by light grey siltstones around the Rupelian/Chattian boundary (Vass et al., 1979, 2007; Vass & Elečko (Eds.), 1989). Both the Egerian benthic and the planktic assemblages of the southern realm indicate a high-nutrient environment without oxygen decrease. Local upwelling cannot be excluded. However, due to bad preservation of foraminiferal tests unsuitable for stable isotope analysis, it is impossible to verify upwelling by geochemical proxies (Holcová, 2017). High humidity and precipitation are confirmed by the finding of hyposaline facies in marginal settings in the Kováčov site (Fig. 1) at the end of this period (Kováč, 1977), with Sr-age of 23.11–23.03 Ma (Fordinál et al., 2014). The opening of a new sea connection is assumed with some uncertainty, and an estuarine circulation regime is reconstructed during the regional stage Egerian (Fig. 3B). The following shallow inflow of water from the Mediterranean (Fig. 3C) is based on the immigration mainly of large benthic foraminifera (BouDagher-Fadel & Price, 2013) which invaded the southern realm of the Central Paratethys during the Egerian warming (Váňová, 1975; Less, 1991).

5.2. The Early Miocene

The Aquitanian time interval (~23–20.5 Ma), correlated in the Central Paratethys with the late Egerian-earliest Eggenburgian, includes the NN1 Zone and the lower part of the NN2 Zone; starting from LO of *Reticulofenestra bisecta* (Gradstein et al., 2012) to FO of *Helicosphaera ampliaperta*. During this time, several FOs of plankton (e.g., *Discoaster druggi, Helicosphaera scissura, Globigerinoides trilobus*), as well as benthos (large foraminifera; Váňová, 1975) were recorded (Fig. 2). The MV-1 borehole near the village of Dolné Plachtince with Sr-age 22.23–22.07 Ma (Fordinál et al., 2014) and the Bretka site (Báldi & Seneš (Eds.), 1975) with Sr-age 22.4–21.09 Ma (Less et al., 2015) were proposed as type localities of this time interval (Fig. 1).

Around the Oligocene/Miocene boundary, the culmination of compression led to considerable N–S shortening of the orogenic system. The frontal part of the Outer Carpathian accretionary wedge began to rise above sea level. The emerged axial zone of the Eastern Alps, western part of the Central Western Carpathians (pre-Neogene basement of the Danube Basin), together with the Transdanubian Range complexes were uplifted (Fig. 5). The palaeogeography of the Central Paratethys Sea remained similar to the previous period. In the sea's northern realm (western part), the Andrychów-6 borehole penetrated the autochthonous Lower Miocene to Upper Oligocene deposits beneath the Carpathian nappes. The basal portion of this sequence is composed of fan-delta pebbly mudstones and conglomerates overlain by dark marine mudstones of the Egerian age (Jurková et al., 1983). North-eastward, these shelf mudstones are followed by the Eggenburgian-Ottnangian grey-greenish marine clays (Zebrzydowice Formation; Garecka et al., 1996) that reflect progressive flooding of the foreland basin. At the same time, this marine basin was linked with the deep residual flysch basins of the Outer Carpathians (Oszczypko et al., 2015). The deep trenches in the Silesian-Krosno zone were bordered by an emerging archipelago of islands built up from the rising front of the Outer Western Carpathian accretionary wedge (Fig. 5). The marine seaways connecting the northern and southern realms of Central Paratethys provided sufficient water exchange. Concerning the palaeogeography, it has to be pointed out that the two parts of the retro-arc basin – the Hungarian and Slovenian Palaeogene basins were not separated at this time (Fodor et al., 1998; Kováč et al., 2016). Hence, the northern retro-arc basin depocenter shifted eastwards (Hungarian part) and the so-called Pétervására Basin started to develop (Fig. 6).

The disappearance of large reticulofenestras in the southern realm of the sea probably reflects the pronounced Mi-1 cooling

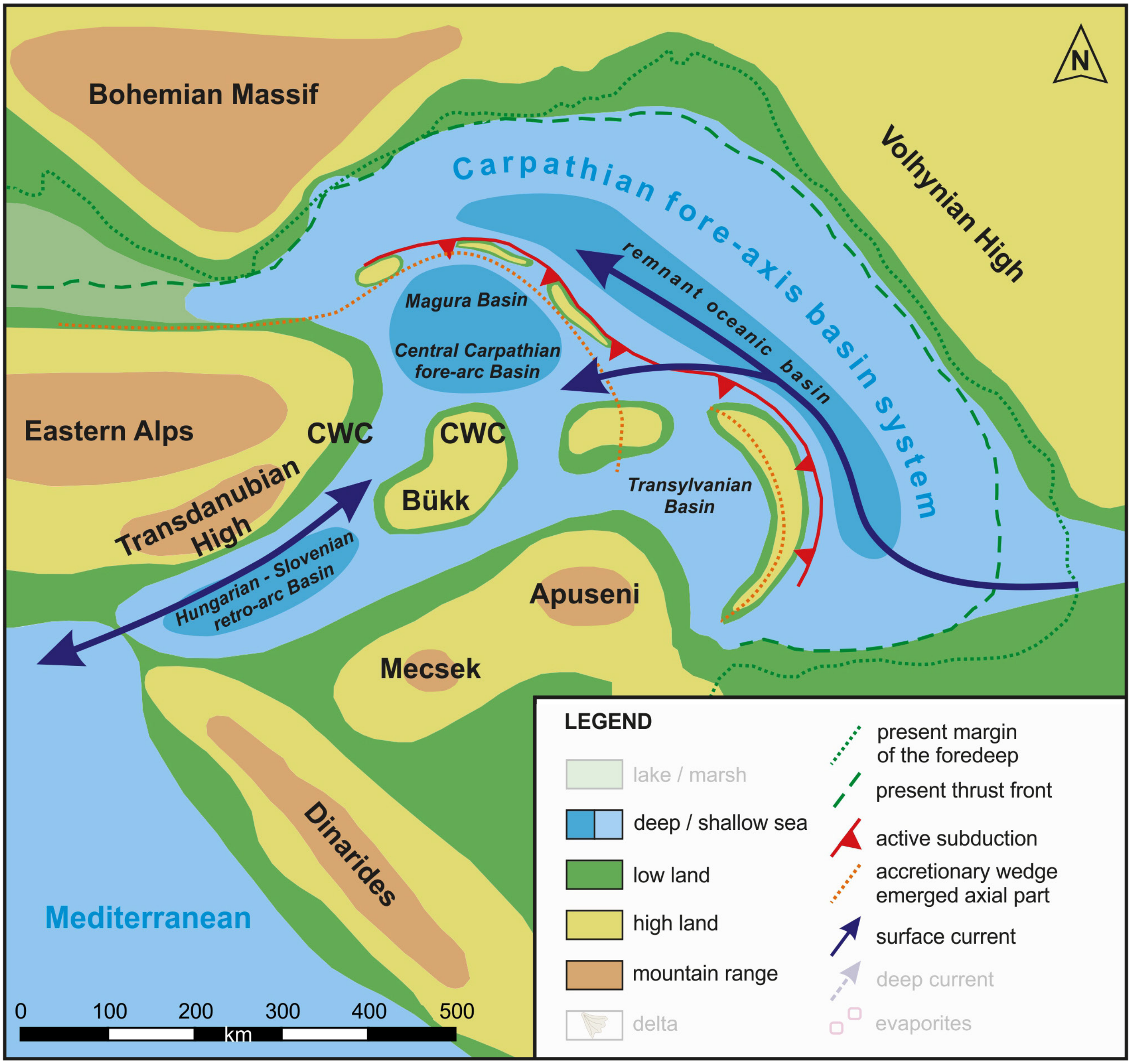


Fig. 5. **Palaeogeography of the Central Paratethys around Oligocene/Miocene boundary: Aquitanian.** *Explanatory notes:* **(CWC)** Central Western Carpathians

event correlated with aridification at the Oligocene/Miocene boundary in the Mediterranean (Sun et al., 2011; Dong et al., 2013). Aridification was also recorded in the terrestrial environment of the Western Carpathians hinterland (Erdei & Bruch, 2004; Erdei et al., 2007). Due to this climatic event, the terrigenous input decreased thereby enabling carbonate production in the marginal parts of the retro-arc basin (Fig. 5). After the Mi-1 cooling event, a warm and humid period followed, indicated by palynological data (Planderová, 1990).

The connection of the Central Paratethys with the Mediterranean along the Alpine Foredeep was closed (Rögl, 1999; Steininger & Wessely, 2000; Rasser et al., 2008). Nevertheless, occasional connections of the retro-arc basin with the Mediterranean cannot be excluded in the south, across the northern edge of Dinarides (Fig. 3C). This assumption is supported by the presence of *Miogypsina gunteri* and *Nephrolepidina morgani* at the Bretka locality (Váňová, 1975), possibly migrating from the foredeep of the Dinarides in the south (Rögl & Steininger, 1969). The sedimentary structures in the Pétervására retro-arc Basin fill were influenced by a high-tidal regime which acted from the east (Sztanó, 1995). This can indicate a possible connection to the Eastern Paratethys (Figs. 3C and 5). Therefore, we can assume the existence of two seaways towards the World Ocean through the Mediterranean and Eastern Paratethys. This assumption can be supported by the fact that the initial Early Miocene transgression in the Eastern Paratethys was of eustatic nature (Popov et al., 2004, 2010; Gozhyk et al., 2015). With the aim of reconstructing the circulation regime, the findings of the mixed Atlantic and Indo-Pacific mollusc species (Báldi, 1979) document for a short period of time the open

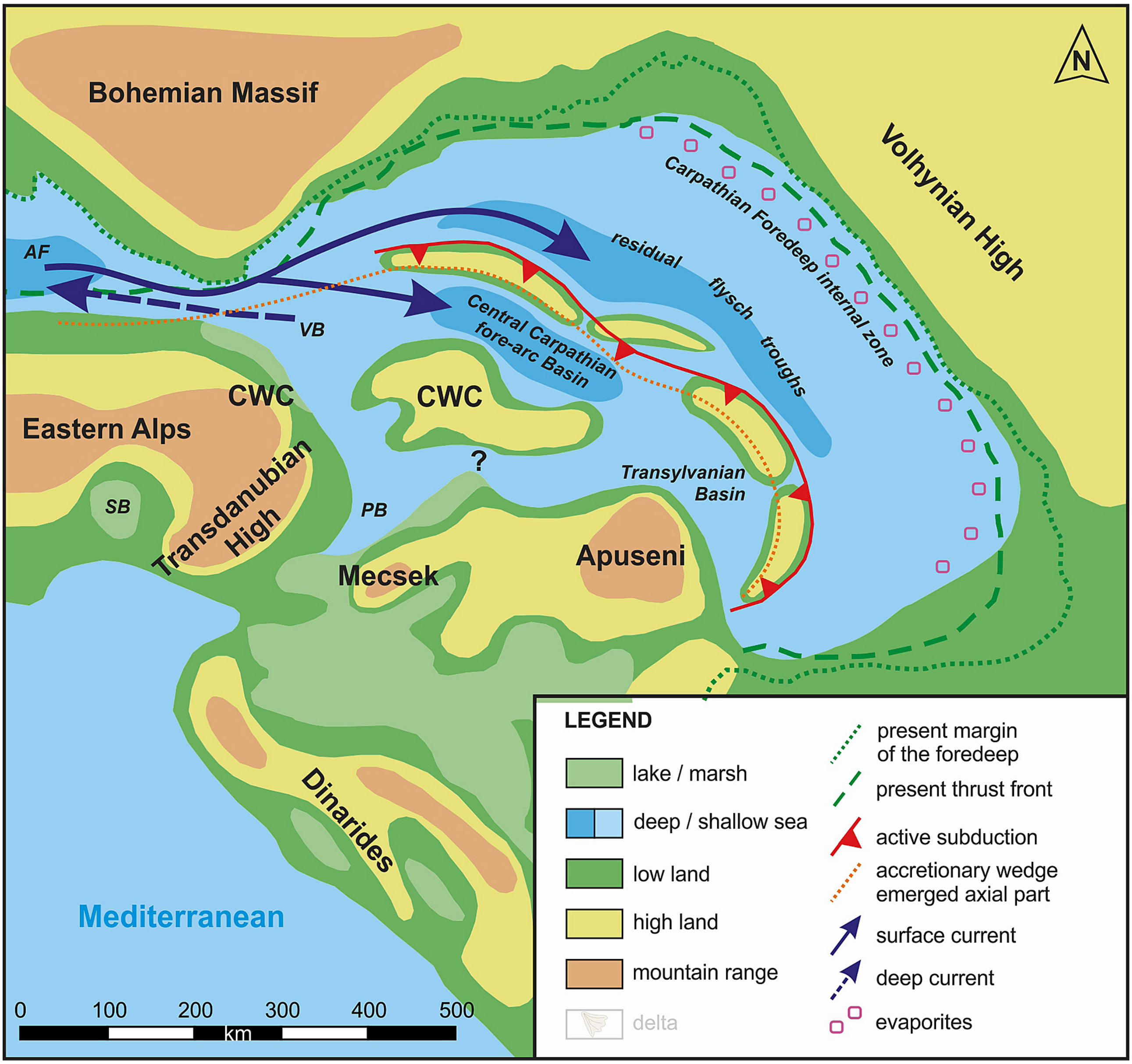


Fig. 6. Palaeogeography of the Central Paratethys: early Burdigalian. *Explanatory notes*: **(AF) Alpine Foredeep; (CWC) Central Western Carpathians; (PB) Pétervására Basin; (SB) Styrian Basin; (VB) Vienna Basin; (?) assumed short living sea connection.**

marine circulation as well. The beginning of the featured interval in Fig. 5 corresponding to the Aq1 boundary (23.05 Ma) and the Aq2 boundary (21.45 Ma) can be correlated with the end of this time (Hardenbol et al., 1998; Gradstein et al., 2012).

The early Burdigalian interval (~20.5–18 Ma), which can be correlated with the late Eggenburgian–early Ottnagian of the Central Paratethys' regional scale, is defined by FO of *Helicosphaera ampliaperta* at the bottom and by LO of *Sphenolithus belemnos* at the top, corresponding to the upper part of the NN3 Zone (Lehotayová, 1984; Andrejeva Grigorovič & Halásová, 2000; Andrejeva Grigorovič et al., 2001; Holcová, 2002). Besides *H. ampliaperta,* other new planktonic taxa immigrated as well (e.g., *Sphenolithus disbelemnos*). The Eggenburgian transgression with new Mediterranean taxa such as the benthic meiofauna of bryozoans (Vávra, 1979), echinoids (Kroh & Harzhauser, 1999), and pectinid bivalves (Mandic & Steininger, 2003) in the NN2 Zone was documented.

The beginning of the chosen time interval can be approximately connected with the Aquitanian/Burdigalian boundary in the Mediterranean (20.4 Ma; Gradstein et al., 2012). For the Western Carpathians type localities, for the lower boundary of the early Burdigalian, we chose the Eggenburgian site Sverepec (Steininger & Seneš, 1971) with Sr-age 20.41–20.29 Ma (Fordinál et al., 2014). For the upper boundary, the Rapovce site with Sr-age 18.57–18.42 Ma can be considered (Fordinál et al., 2014) (Fig. 1).

The Eggenburgian marine connection in front of the Alps has been documented very well (e.g., Rögl, 1999; Steininger & Wessely, 2000; Grunert et al., 2010[b]). It was caused by the proceeding collision of the orogenic system with the platform, producing a deep subsurface load of the underthrusted plate, which accelerated the subsidence in the foredeep basin. The (re)opening of the western gateway led to transgression across this narrow strait and the sea currents entered the northern realm of the Central Paratethys (Fig. 6). The NN2/NN3 zones boundary was recognized in the sediments of the deep residual flysch troughs and the Carpathian Foredeep internal zone (Andreeva-Grigorovich, 2005; Gozhyk et al., 2015). Simultaneously, development and uplift of the forebulge on the western margin of the Volhynian Massif definitively closed the marine connections toward the Eastern Paratethys (Popov et al., 2004). Besides the residual flysch troughs, submerged internal zone of the accretionary wedge (the Magura folds and thrust belt) and the residual fore-arc basin at the edge of the Central Western Carpathians were still covered by the Eggenburgian deep sea (Fig. 6). This basin, filled with synorogenic turbidites containing nannoplankton of the NN1–NN2 zones, was supplied with detritic material from the SE, as well as from uplifted parts of the Pieniny Klippen Belt and Magura Nappe (Cieszkowski, 1992; Oszczypko et al., 1999; Maťašovský & Andreeva-Grigorovich, 2002; Oszczypko & Oszczypko-Clowes, 2002, 2003, 2006, 2009, 2014). During the Ottnangian, the whole Magura Nappe pile emerged above the sea level together with the north-eastern margin of Central Western Carpathians, and the Central Carpathian fore-arc Basin disappeared (e.g., Kováč et al., 1995, 2017[a]; Soták, 2010).

The Eggenburgian marine flooding of the southern realm of Central Paratethys affected the residual retro-arc Pétervására Basin in the south-west, and the Transylvanian Basin in the south-east (Fig. 6). The marine sedimentation in the Pétervására Basin gradually changed to a shallow water and finally to continental in the Ottnangian (Vass et al., 1979, 2007; Halásová et al., 1996). The Late Oligocene-earliest Miocene connection towards the Mediterranean located between the Transdanubian High and the Dinarides was closed.

The changes of water circulation regimes in the Central Paratethys were influenced after the Mi-1a cooling event by a shift to a warmer climate (Gradstein et al., 2012). An anti-estuarine circulation developed in the west, in the Alpine Foredeep (Fig. 3D). The recorded high tidal regime indicates an unrestricted circulation (Faupl & Roetzel, 1990; Bieg et al., 2008) between the Mediterranean water body and the western part of the Central Paratethys during the Eggenburgian-Ottnangian time (Fig. 6). In the east, along the shelf of the Eastern European Platform closing of the marine gateways can be assumed (Fig. 3D). This fact is proved by subsequent evaporite deposition in the basin of Carpathian Foredeep due to isolation (Vorotyshche Fm.; Popov et al., 2004). In the eastern sector of the Carpathian Foredeep deposition of evaporites continued across the NN3/NN4 zones boundary in Ukraine (Gozhyk et al., 2015).

The end of this time interval is marked by the upwelling regime documented in the Ottnang-Schanze locality in the Alpine Foredeep Molasse Basin with a magnetostratigraphic age of 18.056–17.95 Ma (Grunert et al., 2010[a,b]). The upwelling regime foreshadowed the change from an anti-estuarine to estuarine regime in the western part of the Central Paratethys during the following time interval (Fig. 3E). The Eggenburgian-Ottnangian sedimentary record can be correlated with two global sea-level cycles bounded by the Bur1 (20.5 Ma), Bur2 (19.5 Ma), and Bur3 (18.15 Ma) boundaries (Hardenbol et al., 1998).

The following transition from the marine to continental sedimentary environment in the Alpine Foredeep, approximately correlated with the global sea-level change, bordered by the Bur3 and Bur4 boundaries (Hardenbol et al., 1998), interferes with the base of the NN4 Zone (~18–17 Ma). The change terminated at 16.0 Ma in the Swiss part of the Alpine Foredeep Molasse Basin, while the change from brackish to fresh-water sedimentary environment in the German part of the Molasse Basin is limited by ~16.6 Ma (Reichenbacher et al., 2013; Sant et al., 2017). In the Alpine Foredeep, Western Carpathian Foredeep, Vienna Basin (Holcová, 2001[a]), and Novohrad Basin positioned in the orogene hinterlands (Vass et al., 1979, 2007), it is characterised by the appearance of the *"Oncophora beds"* at the base of the Karpatian strata.

The late Burdigalian–early Langhian interval (~17–15 Ma) can be correlated with the Karpatian–early Badenian interval in the Central Paratethys, and it is placed inside the NN4 Zone (Gradstein et al., 2012). Biostratigraphically, the interval was defined by, e.g., LO of *Helicosphaera ampliaperta*, or by local biostratigraphical events (e.g., FO of *Uvigerina graciliformis* and FO of *Globigerinoides bisphericus* (Fig. 2.). However, in the

Central Paratethys, the *Globigerinoides-Praeorbulina* lineage is not continually recorded, with the exception of the Styrian and Sava basins (Hohenegger et al., 2009; Premec Fućek et al., 2017).

Taking into consideration the confusing information from the existing Ar/Ar and Sr-dating from the Central Paratethys area (see chapter Discussion), the aforementioned time interval should then be correlated with the regional stages: latest Ottnangian–Karpatian–earliest Badenian in terms of the concept of Piller et al. (2007). The radiometric dating by U-Pb method between 17.42 ± 0.04 Ma and 17.02 ± 0.14 Ma and additionally by $^{40}Ar/^{39}Ar$ age 16.99 ± 0.16 Ma from tuff that overlies the track bearing sandstone on the famous "Ottnangian" locality Ipolytarnóc in northern Hungary (Pálfy et al., 2007) confirmed that the terrestrial to lacustrine deposition continued from the "Ottnangian" to the "earliest Karpatian". In the Western Carpathians hinterland, the terrestrial deposition of coal seams continued from the Ottnangian to the earliest Karpatian (Figs. 6 and 7). Later, the region was flooded by the "Karpatian sea" (Vass et al., 1979, 2007; Sant et al., 2017).

The Ottnangian marine sediments in the stratotype locality in Austria had a Sr-age 17.15–16.25 Ma (Grunert et al., 2010[b]). The Ottnangian foraminifera from the Cunín-21 borehole in the Vienna Basin (Fig. 1) provided similar Sr-age 17.01–16.9 Ma (Hudáčková et al., 2003), while foraminifera from the Karpatian strata in the Gbely-100 borehole in the Vienna Basin (Fig. 1) provided Sr-age 16.3–15.9 Ma (Hudáčková et al., 2003). The Karpatian locality Cerová–Lieskové (Schlögl et al., 2011)

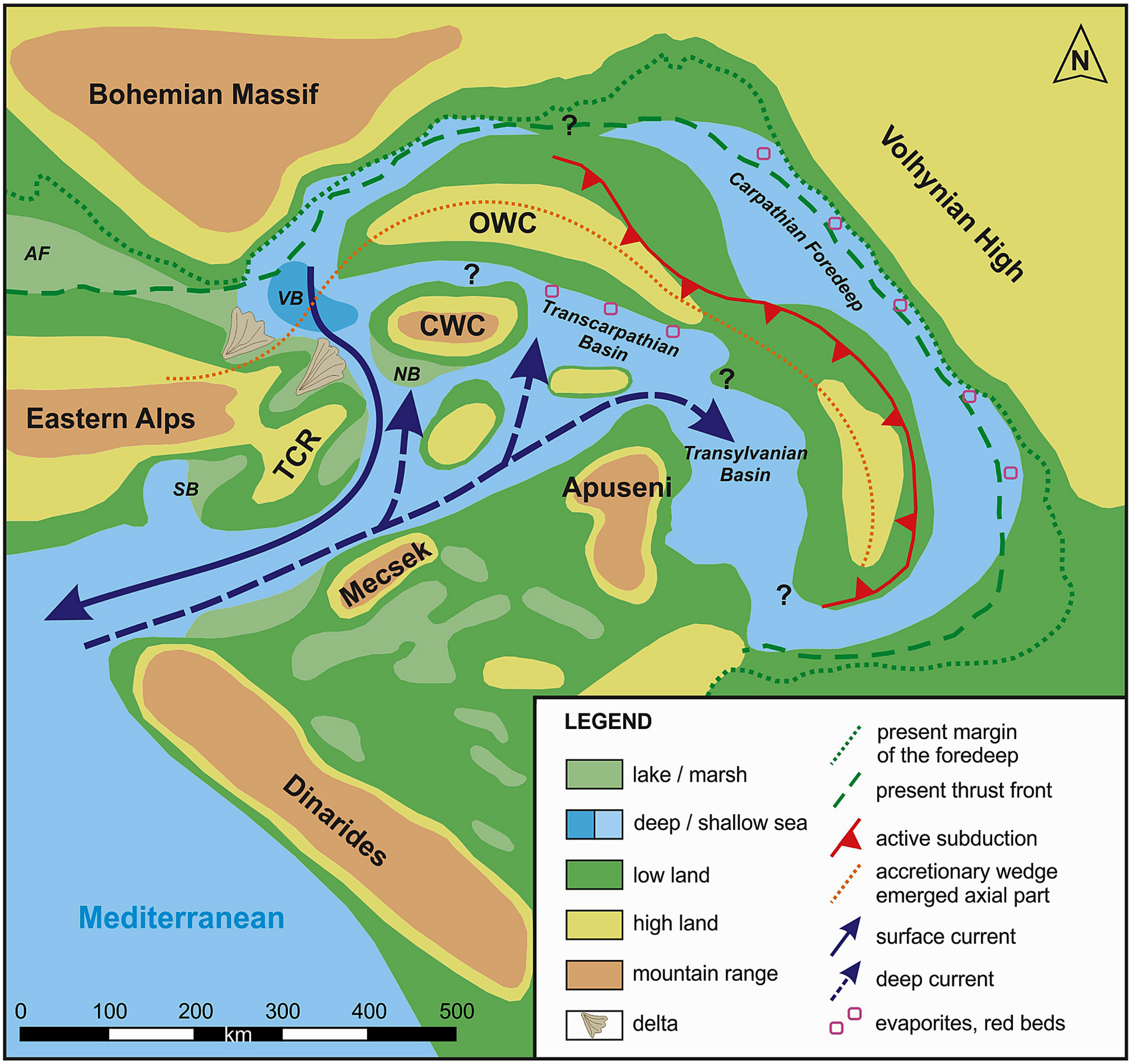


Fig. 7. **Palaeogeography of the Central Paratethys around Early/Middle Miocene boundary: late Burdigalian–early Langhian.** *Explanatory notes:* **(AF)** Alpine Foredeep; **(CWC)** Central Western Carpathians; **(NB)** Novohrad-Nógrád Basin; **(OWC)** Outer Western Carpathians; **(SB)** Styrian Basin; **(TCR)** Transdanubian High; **(VB)** Vienna Basin; **(?)** assumed short living sea connection.

provided Sr-age 16.1–15.3 Ma (Less et al., 2015) (Fig. 1). The results from the Lower Badenian Trenč locality in southern Slovakia with Sr-age 16.02–15.86 Ma (Fordinál et al., 2014) can be included in the same time span as well (Fig. 1). In addition, the "Karpatian" samples from the Hormsdorf borehole (Eibiswald Fm.) provided the youngest age 15.22–15.08 Ma by Ar/Ar method (Handler et al., 2006).

In the latest Ottnangian-early Karpatian, the palaeogeography of the Central Paratethys started to become significantly controlled by tectonic extrusion of the crustal wedge built up from the Central Western Carpathians and Northern Pannonian domain (ALCAPA Mega-unit; e.g., Csontos et al., 1992; Fodor et al., 1998; Kováč, 2000). The oblique collision of the Western Carpathians with the European Platform followed (e.g., Kováč, 2000) and led to deep loading of the platform margin by accretionary wedge complexes. This resulted in accelerated deepening of the Carpathian Foredeep depocenters in the west (Nehyba & Šikula, 2007), while in front of the Alps, continental sedimentation began (Fig. 7).

In the eastern segment of the Carpathian Foredeep internal zone, shallow water to terrestrial deposition of the NN3/NN4 zones is recorded (Sambir Unit; Oszczypko et al., 2006, 2016). The subsiding basin was located between the elevated front of the Eastern Carpathians and the platform forebulge (Fig. 7). The salt and siliciclastic formations (Vorotyshche and Stebnik fms. of the Boryslav-Pokuttya and Sambir units) were overlain by an alluvial fan and deltaic deposits (Sloboda Conglomerates and Dobrotiv Fm.; Oszczypko et al., 2006, 2016). The footprints, rain drops, and red-beds suggest that deposition in a continental environment (delta plain with distributary channels) took place in warm and semi-arid conditions (Stebnik Fm.; Oszczypko et al., 2016). These, often red-coloured sediments, deposited in the eastern part of Carpathian Foredeep, can be correlated with the evaporite deposition (Fig. 3E) in the northern part of Transcarpathian Basin (Eastern Slovak Basin; Kováč et al., 1996; Karoli et al., 1997). The connection towards the Eastern Paratethys across the eastern segment of the foredeep seems to be severed (e.g., Popov et al., 2004; Gozhyk et al., 2015).

The tectonic escape of the Western Carpathians and the Northern Pannonian domain north-eastwards (e.g., Csontos et al., 1992; Fodor et al., 1998; Konečný et al., 2002) caused an abrupt change in the palaeogeography of the Central Paratethys southern realm. In the hinterland of the Eastern Alpine-Western Carpathian orogenic system, the basins located northwards from the Mid-Hungarian Zone (the Styrian and Novohrad-Nógrád basins) accelerated subsidence due to lateral stretching/extension of the moving crustal wedge (Fig. 7). The Ottnangian coal deposition in the Styrian Basin (Sachsenhofer, 1996) was later replaced by the Karpatian sea transgression (Polesny, 2003) due to the opening of a new connection with the Mediterranean, which led via the "Trans-Tethyan-Trench-Corridor" (Rögl, 1998). Similarly, in the Novohrad-Nógrád Basin, in the Western Carpathian hinterland, the Ottnangian terrestrial coal bearing sedimentation changed to brackish and marine in the overlying Karpatian strata (Vass et al., 1979, 2007). The coal seams deposition along the basin margins remained in northern Hungary until the Early Badenian (Ádám, 2006).

Around the Early/Middle Miocene boundary, the marine connection between the northern and southern realms of the Central Paratethys was carried out by a narrow sea corridor running between the Mid-Hungarian Zone in the south and the wedge-top basin (Vienna Basin) located above the Outer Carpathian accretionary wedge and collapsed margin of the Central Western Carpathians in the north-west (Fig. 7). The sea arm was placed in front of the uplifted area, recently forming the pre-Neogene basement of the Danube Basin (e.g., Kováč et al., 1993, 2016, 2017[a,b]). The sea flooded the adjacent subsiding basins in the western part of the Central Western Carpathians as well (e.g., Bánovská Kotlina Depression; Kováč et al., 1993). The basinal equivalent of the "Karpatian" Vienna Basin's deltaic succession represents marine deposits of the NN4 Zone in the northern portion of the Danube Basin. These sediments containing poor foraminiferal associations with abundant bathysiphons can point to the unfavourable conditions of the depositional environment (Murray & Alve, 2011). Later, the displacement of the Western Carpathians northwards activated wrench faults causing subsidence of pull-apart depocenters located along the western (Vienna Basin) and eastern (Eastern Slovak Basin) edge of the Central Western Carpathians in the late Karpatian/earliest Badenian (Kováč et al., 1995; Lankreijer et al., 1995).

The climate of the late Burdigalian–early Langhian interval was generally warm with indistinct Mi-2 (15.85 Ma) cooling event (Gradstein et al., 2012). In the Karpatian, the cold deep water of the Central Paratethys welled-up at many locations in the Novohrad-Nógrád Basin in southern Slovakia (Zlinská & Šutovská, 1990; Holcová, 2013), in the Vienna Basin (Schlögl et al., 2011), in the western segment of the Carpathian Foredeep (Holcová et al., 2015[a,b]) and in the Transylvanian Basin (Beldean et al., 2010). The accelerated uplift of the Alps led to the erosion of the Alpine Foredeep Basin margins (Dellmour & Harzhauser, 2012) and to the development of a new network of rivers. The environmental changes were connected with increased precipitation linked to progradation of huge deltas entering the Vienna and Danube basins southern margins around the boundary of NN4/NN5 zones (Rybár et al., 2015). During the late Karpatian, high input of meteoric water led to forming of surface fresh-water pillows on top of the sea-water layer, documented in the eastern margin of Vienna Basin (Abe et al., 2015). Based on the increased fresh-water supply leading to stratification, we can assume a change to an estuarine circulation regime, with surface water flowing out from the western part of the Central Paratethys across the "Trans-Tethyan-Trench-Corridor" (Fig. 3E). Similarly, this estuarine water circulation was outlined by Brzobohatý (1987) for the Vienna Basin.

5.3. The Middle Miocene

The palaeogeography of the Western Carpathians and its surroundings strongly changed before the onset of the Middle Miocene NN5 Zone (Fig. 2). The accretionary wedge of the Outer Western Carpathians was thrust over/towards the platform margin which was followed by the development of a typical slab-loaded foreland basin (Ziegler et al., 2002). The arcuate shape of the Carpathian mountain chain was primarily controlled by the configuration of the European Platform margin (Oszczypko et al., 2006). In front of the advancing nappes, the foredeep external zone developed, and the marine flooding covered broad areas toward the platform (Figs. 8 and 9). The wedge top basin placed above the accretionary wedge in the western part of the orogeny, at the edge of the Central Western Carpathians was destroyed (the Lower Miocene fill of the Vienna Basin). New hinterland basins depocenters opened south of the exhuming, the axial part of the Western Carpathians. During the development of the back-arc basin system, the extension was associated with upheaval of the asthenosphere and coupled with voluminous volcanic activity (e.g., Konečný et al., 2002).

The late Langhian interval, correlated with the Early Badenian regional stage (*sensu* Kováč et al., 2007) or mid-Badenian (*sensu* Hohenegger et al., 2014), is correlated with the NN5 Zone (14.9–13.6 Ma). A common reworking of the *Helicosphaera ampliaperta,* as well as very scarce occurrence of *Praeorbulina glomerosa circularis* in the Central Paratethys, made FO of *Orbulina* the preferred choice for the lower boundary of this interval in practice, though this event is slightly younger than the Burdigalian/Langhian boundary. The NN5 Zone is based on the occurrence of *Sphenolithus heteromorphus, Helicosphaera waltrans, H. carteri, H. mediterranea*, and *H. walbersdorfensis*. The Lower Badenian foraminifera from

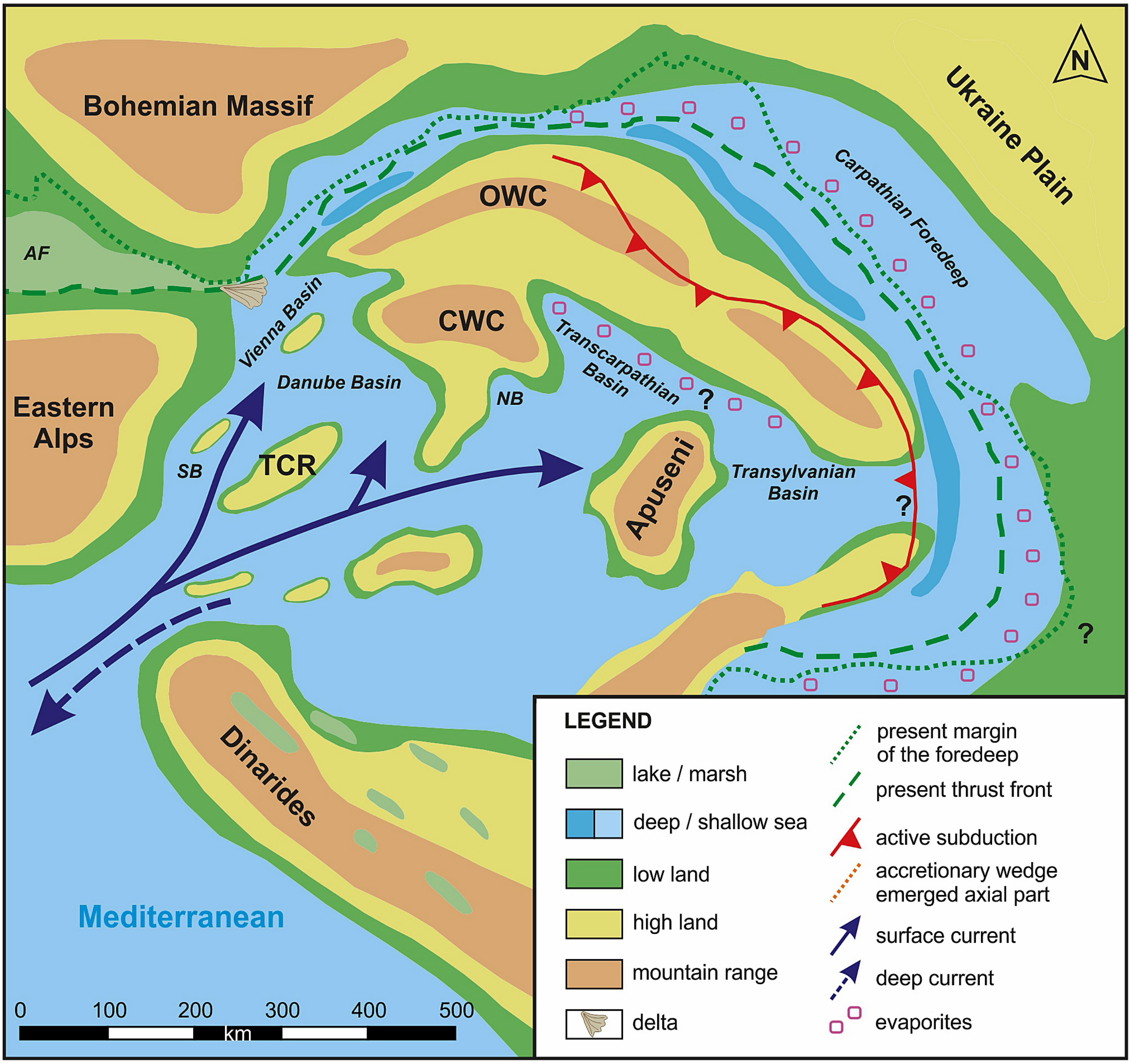


Fig. 8. Palaeogeography of the Central Paratethys during the early Middle Miocene: late Langhian – base of the Serravallian. *Explanatory notes*: **(AF) Alpine Foredeep; (CWC) Central Western Carpathians; (OWC) Outer Western Carpathians; (NB) Novohrad-Nógrád Basin; (SB) Styrian Basin; (TCR) Transdanubian High; (?) assumed short living sea connection.**

the Vienna Basin (Moravský Ján-3 borehole) yielded Sr-age 14.52–14.21 Ma (Hudáčková et al., 2003). The borehole is proposed as a type locality for this interval (Fig. 1). Similar age of 14.39–14.21 Ma was yielded by Ar/Ar dating in the Retznei quarry (Handler et al., 2006).

During the late Langhian (14.9–13.6 Ma), a pronounced transgression is evidenced in the whole Western Carpathian area. The sea connection with the Mediterranean led through the "Trans-Tethyan-Trench-Corridor" (Rögl, 1998), providing a strait deep enough for unrestricted water communication. The northern realm of the Central Paratethys (the Carpathian Foredeep) with shallow water carbonate deposition reached its maximum extent (Fig. 8), and the transgression flooded both the foreland and the front of the Carpathians. Locally, the subsidence exceeded the sedimentation rates, resulting in the establishment of deep marine environment. The water depths in the Carpathian Foredeep Basin varied from shallow to deep neritic in its northern and southern marginal parts and upper bathyal in its axial portion. The Early Badenian sedimentation rates reached 250 to 500 m/Ma in the west (Vass & Čech, 1983; Meulenkamp et al., 1996) and 200 m/Ma in the north (Oszczypko, 1997, 1998), while the north-eastern stable shelf subsided very slowly with sedimentation rates ranking between a few dozens to 50 m/Ma (Oszczypko, 1998). Subsiding grabens were successively filled in with near-shore conglomerates and slope deposits (Ney et al., 1974; Oszczypko, 1998; Oszczypko et al., 2006).

In the southern realm of the Central Paratethys, an epicontinental sea with a number of subsiding basins and archipelagos

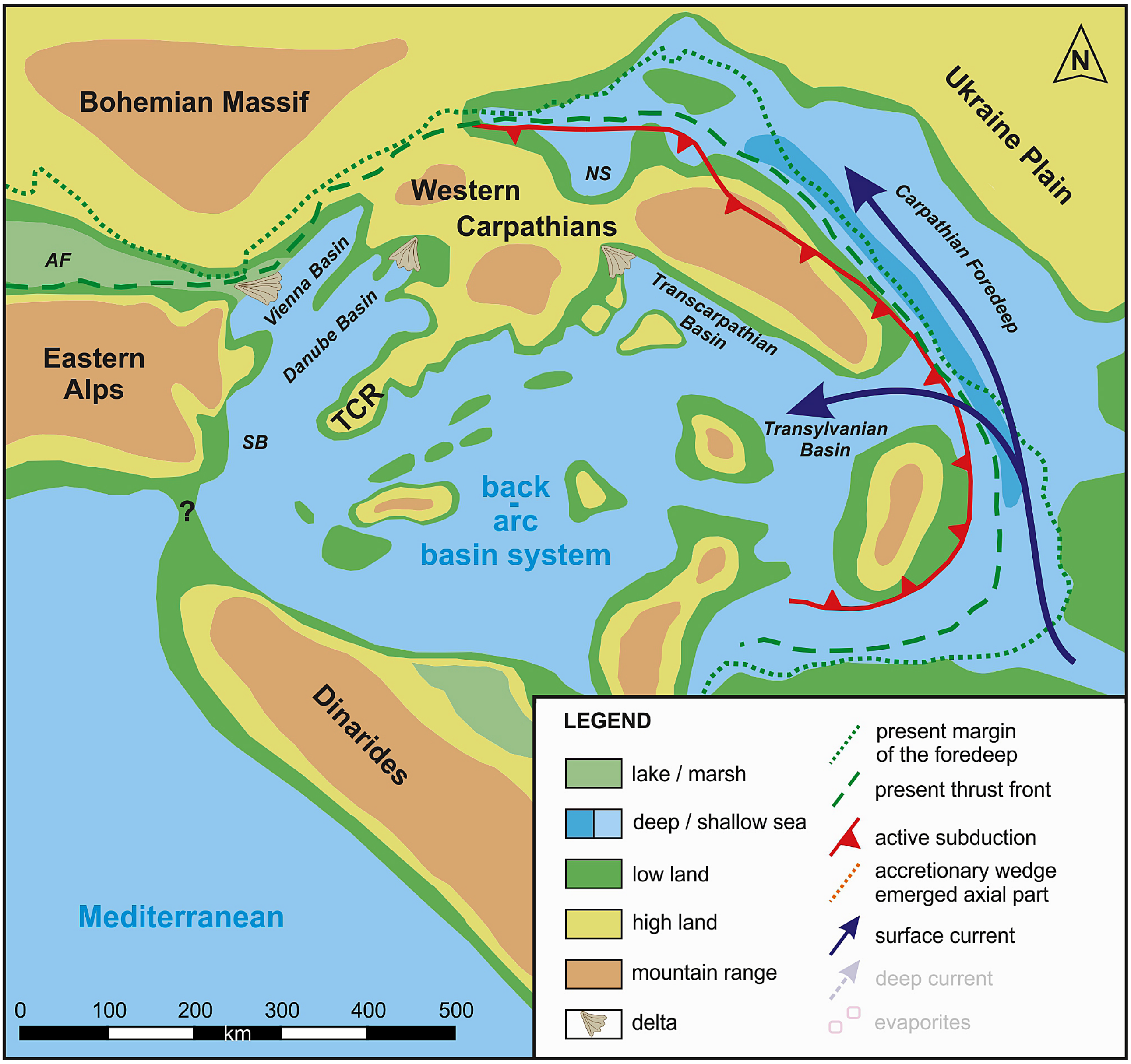


Fig. 9. Palaeogeography of the Central Paratethys during the late Middle Miocene: Serravallian. *Explanatory notes*: **(AF)** Alpine Foredeep; **(NS)** Nowy Sącz piggy-back Basin; **(SB)** Styrian Basin; **(TCR)** Transdanubian High; **(?)** assumed short living sea connection.

developed (Fig. 8). Depositional environment started to be significantly diversified (Hudáčková et al., 2009). The Lower Badenian sea, covering partly the Central Western Carpathians and Northern Pannonian domain, reached up to the shallow bathyal zone (Kováč et al., 1997, 1998, 2007), but the depth gradually decreased to neritic during the NN5 Zone.

The Badenian is characterised by the culmination of the Middle Miocene Climatic Maximum (Gradstein et al., 2012). In the Early Badenian, the beginning of the carbonate production along the back-arc basin system margins was recorded in the Novohrad-Nógrád Basin (Vass et al., 1979, 2007), later at the Eastern Alps and Western Carpathians junction (Mandic, 2004; Holcová et al., 2015[a]). This type of sedimentation probably corresponds with the aridification recorded in the Mediterranean (15.07 Ma; Hüsing et al., 2010). The climatic instability with aridification events is also well-demonstrated later (14.5 Ma) in the western segment of the Carpathian Foredeep (Doláková et al., 2014; Holcová, 2015[a,b]; Nehyba et al., 2016). A model of the simple, two-layer, anti-estuarine circulation (Fig. 3F), based on the benthic foraminifera assemblage and stable isotopes analysis across the NN5 Zone, was proposed for the Central Paratethys by Báldi (2006) and on fish fauna by Brzobohatý (1987) and Brzobohatý et al. (2007). During this time, most of the foraminiferal assemblages of the Western Carpathian basins were rich (both in diversity and abundance) in biserial agglutinated forms, making some authors regard them as biostratigraphically significant on a large regional scale (Grill, 1941, 1943; Cicha et al., 1975). The dominance of agglutinated forms can indicate conditions with high organic flux and higher temperature on the sea floor, often coupled with strong tidal currents (Murray & Alve, 2011). Such conditions were confirmed by micropalaeontological and sedimentological studies from this region (Vass, 1977; Vass et al., 1979; Kováč et al., 1999, 2004, 2007; Bartakovics & Hudáčková, 2004).

At the Early/Late Badenian boundary, or the Langhian/Serravallian boundary (*sensu* Kováč et al., 2007), a gradual weakening in the communication with the Mediterranean Sea can be recognised, leading to total cessation of the water exchange between the western and eastern parts of the Central Paratethys. It is documented by the evaporite deposition (Fig. 8) in the Carpathian Foredeep, Transcarpathian, and Transylvanian basins in the east (e.g., Kováč et al., 1995; Filipescu & Girbacea, 1997; Kováč & Hudáčková, 1997; Chira & Draghici, 2002; Andreyeva-Grigorovich et al., 2003[b]; Bukowski et al., 2003; Popov et al., 2004; Báldi, 2006; Krézsek & Bally, 2006; Bąbel, 2007, Śliwiński et al., 2012; Peryt, 2013[a,b]; Filipescu & Filipescu 2015; Gozhyk et al., 2015). This sedimentary event is referred to as the Badenian Salinity Crisis (BSC) which started at 13.81 ± 0.08 Ma (de Leeuw et al., 2010).

A partial isolation of the Carpathian Foredeep with possible occasional marine ingressions from the south-east is documented by the prominent deposition of evaporites (Gonera, 2013; Peryt, 2013[a,b]). The beginning of BSC can be approximately correlated with the Ser1 3rd order sequence boundary (13.82 Ma; Hardenbol et al., 1998) and the end is below the Ser2 boundary of the sea-level change (13.5 Ma; Hardenbol et al., 1998). Moreover, some authors (e.g., de Leeuw et al., 2010) suggested a causal relationship between the cooling during the glacial event Mi-3b (13.8 Ma; Gradstein et al., 2012) and evaporite deposition by assuming a sea level drop to restricted communication of the evaporitic basins with the World Ocean. Taking into consideration the structural evolution of the Carpathians and the influence of the local climate, we are inclined to believe that these factors are, besides global cooling, the prominent cause of BSC. On the other hand, the cooling trend probably caused the extinction of large *Helicosphaera waltrans* in the Mediterranean area (14.357 Ma; Abdul Aziz et al., 2008; Hüsing et al., 2010).

Taking into account new radiometric data, this episode of the evaporite formation can be of relatively short duration – around 600 ky (e.g., Bukowski & Szaran, 1997; Bukowski et al., 2010; de Leeuw et al., 2010; Studencka & Jasionowski, 2011). The shallow (stable shelf) parts of the basin were dominated by a sulphate facies, whereas its deeper parts, located along the Carpathian front were occupied by a chloride-sulphate facies (Garlicki, 1979; Kasprzyk, 1999, 2005; Bąbel, 2004; Peryt, 2006). During that time, the Carpathian orogenic front was still mobile, which is recorded by the occurrence of flysch blocks in the salts of the Wieliczka and Bochnia salt mines (Kolasa & Ślączka, 1985). At the end of the evaporite deposition when the basin was the shallowest (Gonera, 1994; Czepiec, 1996), a tectonic uplift of the foreland resulted in the development of regional unconformity. Erosion of up to 50–100 m of evaporitic and sub-evaporitic deposits is assumed (Rzeszów Palaeo-Ridge).

The Serravallian time interval (~13.8–11.6 Ma) we correlated with the Late Badenian–Sarmatian regional stages with a base below the beginning of the NN6 Zone, based on LO of *Sphenolithus heteromorphus* (on NN5/NN6 boundary), and the end above the FO of *Discoaster kugleri* (on NN6/NN7; *sensu* Martini 1971). The presence of *Reticulofenestra pseudoumbilicus* > 7 μm, discoasters, *Sphenolithus abies*, *Helicosphaera walbersdorfensis*, *Calcidiscus premacintyrei* and later FO of *Calcidiscus macintyrei* > 11 μm, locally acme of *Calcidiscus* spp., *Braarudosphaera bigelowi parvula*, ascidian spicules (*Perforocalcinella fusiformis*) are often documented in the Western Carpathian sedimentary record (e.g., Andreeva-Grigorovich, 2002; Andreeva-Grigorovich et al., 2003; Kováč et al., 2005, 2006, 2008; Schütz et al., 2007; Chira & Malacu, 2008; Bartol, 2009; Lelek et al., 2016). The NN7 Zone was designated in the upper part of this interval in the Pannonian Basin area (Sarmatian s.s., Nagymarosy, 1982; Galović, 2014), in the Carpathian Foredeep and in sediments of the Eastern Paratethys (Gaździcka, 1994; Oszczypko-Clowes et al., 2012; Gozhyk et al., 2015).

The Sr-calibration curve in the interval of the 15 to 11 Ma is too flat which causes problems in interpretation. In the Carpathian Foredeep, the $^{87}Sr/^{86}Sr$ ratio of the analysed otoliths points to a wide time span of 14.78–13.10 Ma at the top of the Lower Badenian sediments, falling into an interval of poor time resolution of the $^{87}Sr/^{86}Sr$ chemostratigraphy (Brzobohatý et al., 2016). The Devínska Nová Ves clay pit (Vienna Basin), with Sr-age 13.7–13.39 Ma (Hudáčková et al., 2003), was chosen as the Upper Badenian type locality (Fig. 1). The overlying strata (Sandberg locality at the foot of Devínska Kobyla hill) provided

Sr-age 13.17–12.88 Ma (Fordinál et al., 2014) and the uppermost part of the profile was dated by Fordinál et al. (2014) to Sr-age 12.9–12.63 Ma, and to Sr-age 12.54 Ma by Less (this work, Fig. 1). The sediments at the very top of Devínska Kobyla hill, the so-called "Lingula Bed", were dated to Sr-age 13.5–10.3 Ma by Less (this work) and ranked to the Sarmatian.

After the salinity crisis, in the Late Badenian and Sarmatian, the deepest depocentres of the Central Paratethys developed in the Carpathian Foredeep with sedimentation of deep water turbidites at the junction of the Western and Eastern Carpathians (Oszczypko, 1998, 1999; Oszczypko-Clowes et al., 2012; Lelek et al., 2016). The Carpathian nappes were shifted northwards for the last time. It was followed by an episode of relaxation of the strain and extensional fault tectonics leading to the opening of the Nowy Sącz piggy-back Basin in the northern domain of the accretionary wedge (Fig. 9). The sea invaded into the previously eroded Outer Carpathians and fresh-water and shallow-marine sediments up to 600 m thick were deposited unconformably upon folded and partly eroded flysch deposits of the Magura Flysch Belt (Oszczypko, 1973; Oszczypko-Clowes at al., 2009).

In the southern realm of the Central Paratethys, the Upper Badenian sedimentary record documents an increase of deltaic facies towards the overlying strata. The landscape started to resemble the present state of the main mountain ranges. In the western part of the back-arc basin system, shallow water carbonate platforms were often present, occasionally with traces of salinity fluctuation (Kókay, 1985). Later, during the Sarmatian, a shallow water deltaic to a coastal plain environment dominated along the south-western slopes of the Western Carpathians (Fig. 9). In contrast to the shallow water environment in the west, in the Eastern Slovak Basin (northern part of the Transcarpathian Basin) and in the Transylvanian Basin, a deep water environment dominated during the whole time interval (Kováč et al., 1995, 1996; Suciu et al., 2005; Filipescu et al., 2014). The deep water environment of the Eastern Slovak Basin is documented by the Upper Badenian microfauna with the agglutinated species *Bogdanowiczia pokutica* (Hudáčková unpublished data). This type of deep, open marine environment, resembling the Carpathian Foredeep, is characterised by a common occurrence of stenohaline pteropods at the base of this interval. This horizon (*Spirialis* or *Limacina* beds of the Spirialis Clay Member) is generally considered to be isochronous in the Central Paratethys and is interpreted (Peryt, 2006) as a result of mass extinction caused by mixing of the upper water layer of normal salinity and moderate temperature with the lower warmer water layer of high salinity. The *Spirialis* – *Limacina* preservation proves the anoxia at the basin's bottom, where the lack of bioturbation makes the fossilization of fragile tests possible (Báldi, 2006). The foraminiferal assemblages of the interval's upper part indicate an outer shelf setting, where the water salinity was close to normal (Czepiec & Kotarba, 1998).

The similarity of the deep water sedimentary facies and the environment in basins located in the eastern portion of the Central Paratethys led us to assume that the Transcarpathian and Transylvanian basins could still communicate with the Carpathian Foredeep (Fig. 9), probably through seaways across the emerging, but not yet fully uplifted parts of the accretionary wedge of the Outer Carpathians (Alexandrowicz & Pawlikowski, 1980; Bohn-Havas & Zorn, 1993; Andreyeva-Grigorovich & Savytskaya, 2000; Śliwiński et al., 2012; Palcu et al., 2015). Considering the following facts: *(i)* the presence of the Upper Badenian associations of planktic organisms such as pteropods, radiolarians, and planktic foraminifers (Alexandrowicz & Pawlikowski, 1980; Bohn-Havas & Zorn, 1993; Andreyeva-Grigorovich & Savytskaya, 2000; Śliwiński et al., 2012); and *(ii)* the common presence of the Sarmatian piggy-back basins on top of the Outer Western Carpathian accretionary wedge (Oszczypko-Clowes at al., 2009), we can imagine a scenario where the accretionary wedge of the Outer Carpathians could be partly flooded from the east for a short time (Fig. 3G,H). Moreover, *(iii)* this hypothetical opening of connections with the Eastern Paratethys, as evidenced by palaeogeographic maps (Popov et al., 2004), can help us explain the rapid change of the brine chemistry of the Central Paratethys in the Sarmatian (e.g., Pisera, 1996; Harzhauser & Piller, 2004; Palcu et al., 2015).

During the Late Badenian, the connection of the Central Paratethys with the Mediterranean was probably closed during longer time intervals, leading to major changes in the circulation of water masses, which led to the development of low oxic conditions at the sea bottom (Fig. 3G) in the whole area (Báldi, 2006; Kováčová & Hudáčková, 2009; Kováčová et al., 2009; Gonera, 2013; Filipescu & Filipescu, 2015; Kováč et al., 2017[a]). The time span can be correlated with the Mi-4 glacial event (Gradstein et al., 2012; Ogg et al., 2016). The stratified water column and low oxic environment is typical by substantial domination of the stress genera such as *Bulimina, Bolivina*, and *Cassidulina*, alternating with layers without benthos, but with mass occurrence of the planktic species *Globigerina bulloides* and *Turborotalita quinqueloba* (Kováčová & Hudáčková, 2009; Kováčová et al., 2009; Peryt, 2013[a]). Nevertheless, during the Late Badenian, in addition to isolation and low oxic conditions at the Central Paratethys sea bottom, occasional open circulation regime with the Mediterranean and Eastern Paratethys can be suggested (Figs. 3G and 9) on the basis of mosaic distribution of two distinct types of nannofossil assemblages reported from the Central Paratethys (Bartol, 2009; Bartol et al., 2014; Galović, 2014). In the same time, this fact supports the idea of water currents flowing into the Central Paratethys from the Mediterranean and Eastern Paratethys which might have mixed together (Stradner & Fuchs, 1979; Pavšič & Mihajlović, 1981; Andrejeva Grigorovič & Turčinova, 1983; Mihajlović & Knežević, 1989; Andreeva-Grigorovich, 2002; Brânzilă & Chira, 2005; Bartol et al., 2014; Bitner et al., 2014). Báldi (2006) suggests that during the NN6 Zone (Late Badenian) the circulation between the western part of the Central Paratethys and Mediterranean switched from the anti-estuarine to estuarine one.

Marine anti-estuarine communication (Fig. 3G,H) between the Central and Eastern Paratethys during the Late Badenian can be considered due to the occurrence of new planktic foraminifers, including *Globigerina druryi, G. decoraperta*, and *G.*

nepenthes and the presence of the Tenuitellinata Zone in the Transylvanian Basin (Filipescu & Silye, 2008; Hudáčková et al., 2013). Based on the occurrence of the foraminiferal species of *Streptochilus*, this connection persisted till the Sarmatian (Filipescu & Silye, 2008). It is assumed that the water masses flowing out of the Eastern Paratethys entered the semi-closed basin due to anti-estuarine circulation for several short time spans (Fig. 3H). Kókay's concept (Kókay, 1985), based on the analysis of molluscs fauna (Konkian event), is supported by the continuous sea level rise in the Eastern Paratethys due to freshwater input by rivers (Popov et al., 2010) leading also to decreasing salinities from the Late Badenian–Sarmatian onward. The negative surface water oxygen isotope values at the end of the NN6 Zone are interpreted not just as a result of global cooling (Bicchi et al., 2003; Kováčová et al., 2009) but also as slightly reduced salinity in the Central Paratethys (Hudáčková & Kováč, 1993; Báldi, 2006; Kováčová & Hudáčková, 2009). On the other hand, the benthic foraminiferal assemblages of the Sarmatian sea marginal environments in the western and southern parts of the Central Paratethys reflect the variability and prove the isolation which led to local development of shallow marine hypo- and hypersaline environments (Harzhauser & Piller, 2004, 2007; Kováč et al., 2005; Koubová & Hudáčková, 2010; Tóth et al., 2010; Peryt & Jasionowski, 2012; Zlinská et al., 2010; Filipescu et al., 2014; Gozhyk et al., 2015).

5.4. The Late Miocene

The early Tortonian interval (~early Pannonian; 11.6–10 Ma) in the Western Carpathian basins is due to lack of NN7 and

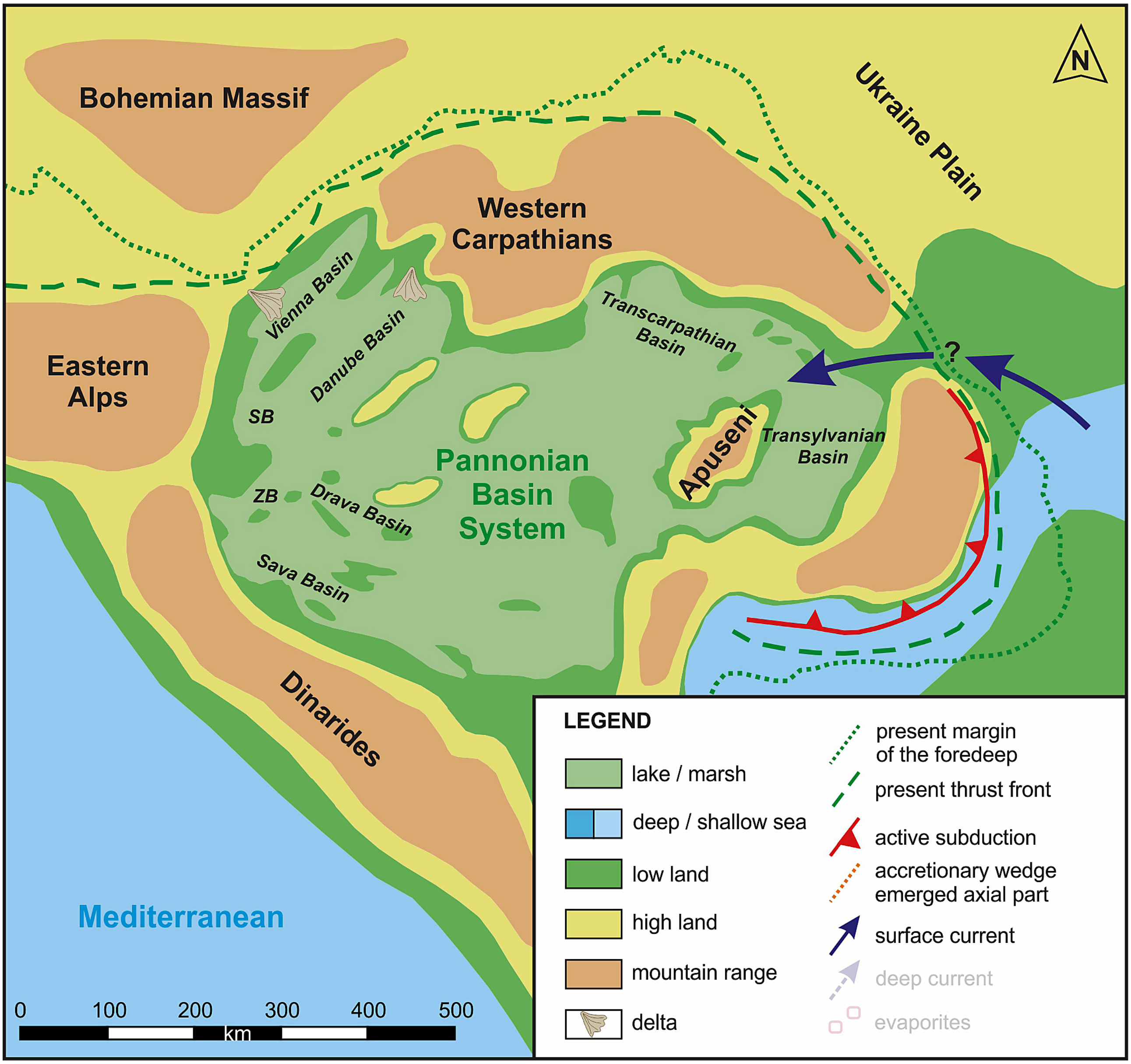


Fig. 10. **Palaeogeography of the Pannonian Basin System during the early Late Miocene: early Tortonian.** *Explanatory notes:* **(SB) Styrian Basin; (ZB) Zala Basin; (?) assumed short living sea connection.**

NN8 zones in this region correlated only with the endemic NN9 Zone (Kováč et al., 2006). Moreover, this time span is confirmed by Be-isotope dating (Šujan et al., 2016), and linked to the upper part of Tor1 cycle of sea-level change (Hardenbol et al., 1998). The basal Pannonian sedimentary sequences were deposited in an environment of decreased salinities, where foraminifera and ostracods of reduced test size persisted (*Trochammina*, *Nonion* sp.), which were able to withstand the dysoxic or even episodic anoxic conditions (Korecz, 1985; Kováč et al., 2008). Occurrences of the calcareous nannoplankton in the Central and the Eastern Paratethys (Luljeva, 1989; Mărunteanu & Papaianopol, 1996; Reischenbacher et al., 2007; Chira & Malacu, 2008; Cziczer et al., 2009; Radionova et al., 2012; Kováč et al., 2017[a,b]) imply short living marine connections with the Eastern Paratethys. The early Tortonian interval is documented by the presence of the endemic nannoplankton species of the genera *Praenoelaerhabdus, Noelaerhabdus, Isolithus, Bekelithela*, what define also local biozones (Nagy et al., 1995; Mărunteanu, 1997; Andreeva-Grigorovich et al., 2003; Andreyeva-Grigorovich et al., 2003[a]; Ćorić, 2004, 2005, 2006; Andreeva-Grigorovich, 2006; Chira, 2006; Kováč et al., 2006, 2008, 2017[a,b]), which can be correlated with the standard NN9 Zone and the MN9 and MN10 mammal biozones (Kováč et al., 2006, 2011).

During the Late Miocene, the Central Paratethys sedimentary environment gradually changed to brackish and fresh-water, due to the isolation of Lake Pannon (e.g., Kováč, 2000; Uhrin & Sztanó, 2012; Neubauer et al., 2015). The accommodation space was filled with a large amount of deltaic sediments entering the basin from its margins (Fig. 10), so the deep water lake environment gradually changed to a shallow water one (e.g., Magyar et al., 1999; Šujan et al., 2016). At the end of this period, the influence of the Messinian salinity crisis and coeval sea level fall in the Mediterranean was overprinted by tectonics in the closed residual lake system (Uhrin et al., 2009; Leever et al., 2011), and the alluvial plains started to develop on the majority of the back-arc basin territory. The mountain ranges gained features similar to their present form. Despite the fact that some authors do not accept the possibility of connections to the Eastern Paratethys (e.g., Magyar et al., 1999, 2013) during the early Pannonian, the immigration of marine fauna (foraminifera) and nannoplankton correlated with the NN9 Zone could supports this idea (Kováč et al., 2006, 2008).

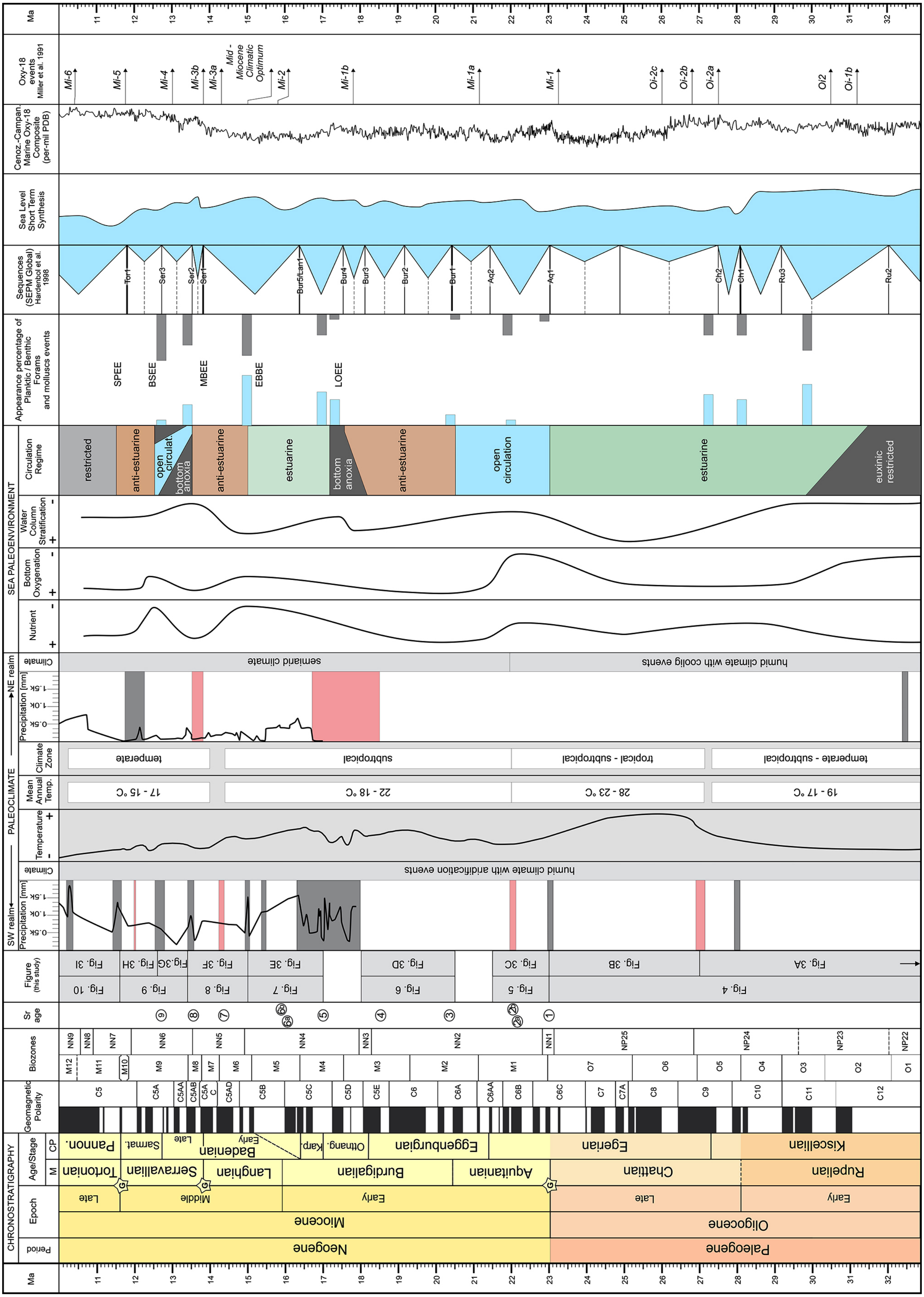
6. DISCUSSION

While compiling the publication's results, (re)interpreted facts, and additional new data for the palaeogeography, climate changes, depositional environment, and water circulation regimes of the Central Paratethys Sea, several aspects have been taken into account: *(i)* the individual maps (Figs. 4–10) of the studied time intervals should represent significant changes in palaeogeography and the overall changes in the extent of water-covered areas controlled by the main phases of geodynamic evolution of the Western Carpathians; *(ii)* an exact definition of the chosen time interval for the palaeogeographic maps had to be based on standard biostratigraphic markers, or at least had to be supported by Sr-isotope dating results; (*iii*) opening and closing of straits/gateways to other marine realms had to be regarded; and (*iv*) the importance and influence of the regional climate was taken into account.

(i) The time intervals of the presented palaeogeographical maps were chosen to reflect the gradual shortening and growth of the Outer Western Carpathians accretionary wedge, the exhumation of the axial part of the Central Western Carpathians, and the opening and closing of individual basins in the hinterland of the orogenic system (e.g., Kováč et al., 1993, 1999, 2016, 2017[a]; Kováč, 2000). Maps reflect the maximal possible extent of the Central Paratethys marine flooding and ignore the local hiatuses, or local changes of the coastline, which are generalised. The extent of the water-covered areas was estimated from an extrapolation sedimentary record into the connecting areas with no record, and/or from the similarity of sedimentary record between two or more neighbouring basins with the same facial pattern. The size of the erosion was interpreted by data on exhumation or burial in the individual parts of the orogen (e.g., Hurai et al., 2004, 2006; Kováč et al., 1994, 2006; Kotulová, 2010; Danišík et al., 2004; Králiková et al., 2014[a,b], 2016). This allowed us to conclude that the thickness of the Oligocene and Miocene strata along the northern margin of Central Western Carpathians was at least about 1–2 km more than at present (e.g., Kováč et al., 1994, 2016). On the other hand, the pre-Cenozoic basement of the Neogene basins, which are presently deeply buried, used to be on the surface (e.g., Danube Basin; Kováč et al., 2016, 2017[a]).

(ii) The problem of the exact determination of the used time ranges stems from the inaccurate definition of the regional stage boundaries (Chapter 3; Figs. 2 and 11) and their correlation with the Mediterranean or Eastern Paratethys chronostratigraphy (e.g., Piller et al., 2007; Gradstein et al., 2012; Reichenbacher et al., 2013; Gozhyk et al., 2015).

One of the most important and long outstanding issues of the Central Paratethys regional stratigraphy is the boundary between the Karpatian and Badenian regional stages, which is equated with the Burdigalian/Langhian boundary in the Mediterranean, based on FO of *Praeorbulina*. Berggren et al. (1995) discussed FAD of *P. glomerosa sensu stricto* at 16.1 Ma as a possible Langhian boundary marker, due to the diffused onset of *P. sicana* at the type locality (Fornaciari et al., 1997[b]; Iaccarino et al., 2011). In terms of the Central Paratethys nannofossil zonation, the base of the Badenian was originally correlated with the base of the NN5 Zone (Papp et al., 1978), but based on FO of *Praeorbulina*, it is now placed in the uppermost part of NN4 Zone (e.g., Kováč et al., 2004; Piller et al., 2007; Rögl et al., 2007). Unfortunately, the genus *Praeorbulina* is rare within the NN4 Zone in the sediments of the Western Carpathian basins and usually, it makes its FO only later in the NN5 Zone along with *Orbulina suturalis*. This event can be correlated with global LO of *Helicosphaera ampliaperta* at 14.9 Ma (Gradstein et al., 2012). However, *Helicosphaera ampliaperta* disappears around 16 Ma and then appears again at 15.8 and 15.45 Ma in the Mediterranean (Turco et al., 2011).



Ma
Oxy-18 events Miller et al. 1991
Mi-6
Mi-5
Mi-4
Mi-3b
Mi-3a
Mid - Miocene Climatic Optimum
Mi-2
Mi-1b
Mi-1a
Mi-1
Oi-2c
Oi-2b
Oi-2a
Oi2
Oi-1b
Cenoz.-Campan. Marine Oxy-18 Composite (per-mil PDB)
Sea Level Short Term Synthesis
Sequences (SEPM Global) Hardenbol et al. 1998
Tor1
Ser3
Ser2
Ser1
Bur5/Lan1
Bur4
Bur3
Bur2
Bur1
Aq2
Aq1
Ch2
Ch1
Ru3
Ru2
Appearance percentage of Planktic / Benthic Forams and molluscs events
SPEE
BSEE
MBEE
EBBE
LOEE
SEA PALEOENVIRONMENT
Circulation Regime
restricted
anti-estuarine
open circulat.
bottom anoxia
estuarine
open circulation
euxinic restricted
Water Column Stratification
Bottom Oxygenation
Nutrient
PALEOCLIMATE
NE realm
Climate
semiarid climate
humid climate with cooling events
Precipitation [mm]
Climate Zone
temperate
subtropical
tropical - subtropical
temperate - subtropical
Mean Annual Temp.
17 - 15 °C
22 - 18 °C
28 - 23 °C
19 - 17 °C
Temperature
humid climate with aridification events
SW realm
Figure (this study)
Fig. 3I
Fig. 3H
Fig. 3G
Fig. 3F
Fig. 3E
Fig. 3D
Fig. 3C
Fig. 3B
Fig. 3A
Fig. 10
Fig. 9
Fig. 8
Fig. 7
Fig. 6
Fig. 5
Fig. 4
Sr age
Biozones
Geomagnetic Polarity
CHRONOSTRATIGRAPHY
Age/Stage
CP
M
Pannon.
Sarmat.
Badenian
Karp.
Ottnang.
Eggenburgian
Egerian
Kiscellian
Tortonian
Serravallian
Langhian
Burdigalian
Aquitanian
Chattian
Rupelian
Epoch
Late
Middle
Early
Miocene
Oligocene
Period
Neogene
Paleogene

Fig. 11. Correlation chart: Chronostratigraphy, Palaeoclimate, Palaeoenvironment and Water Circulation Regimes.
Chronostratigraphy: Modified from the Time Scale Creator; Harzhauser & Piller, 2007; Ogg et al., 2016; Sr-dating - this study) with marked time intervals figured at palaeogeographical maps (Figs. 4–10 of this study). *Explanatory notes:* (**asterisk**) Global Boundary Stratotype Section and Point; (**CP**) Central Paratethys; (**M**) Mediterranean. **Sr – data** (after Hudáčková et al., 2003; Fordinál et al., 2014; Less - this study): **1.** Oligocene/Miocene boundary, base of the Aquitanian (Kováčov site; 23.11 Ma). **2.** early Aquitanian, late Egerian (**2a** - Dolné Plachtince site; 22.15 Ma, **2b** - Bretka site; 22.15 Ma). **3.** base of the Burdigalian, early Eggenburgian (Sverepec site; 20.35 Ma). **4.** middle Burdigalian, late Eggenburgian (Rapovce site; 18.52 Ma). **5.** late Burdigalian, top of the Ottnangian (Cunín-21 borehole; 16.95 Ma). **6.** Early/Middle Miocene boundary, Karpatian/earliest Badenian (**6a** - Gbely-100 borehole; 16.05 Ma; **6b** - Trenč site; 15.94 Ma and Cerová–Lieskové site; 15.87 Ma). **7.** Langhian, Early Badenian (Moravský Ján 3 site; 14.36 Ma). **8.** base of the Serravallian, early Late Badenian (Devínska Nová Ves clay pit; 13.54 Ma). **9.** middle Serravallian, base of the Sarmatian (Sandberg site; 12.76 Ma).
Palaeoclimate: Precipitation curve modified from Böhme et al. (2011), Bruch et al. (2011), Vass et al. (1987), and this study. *Explanatory notes:* (**pink coloured squares**) arid climate, carbonate or evaporites deposition; (**grey coloured squares**) humid climate and/or coal deposition. **Temperature, Mean Annual Temperature, and Climate zones** modified from Planderová (1990), Böhme (2003), Harzhauser (2005), Kováčová et al. (2011), Jiménez-Moreno et al. (2008), and this study.
Palaeoenvironment of the sea, circulation models based on changes of depositional systems of the sea and reflected by microfossil proxies. **Nutrient, Bottom oxygenation, Water column stratification** – analysis modified from (Lehotayová, 1978, 1982, 1984; Nagymarosy, 1982; Szczechura, 1982; Zlinská & Šutovská, 1991; Oszczypko-Clowes, 1999, 2001, 2008, 2012; Peryt, 1999, 2013[a,b]; Holcová, 2001[a,b]; Spezzaferri & Ćorić, 2001; Mandic et al., 2002; Oszczypko & Oszczypko-Clowes, 2002, 2003, 2006, 2009, 2014; Rögl et al., 2002, 2007; Spezzaferri et al., 2002, 2004, 2009; Andreeva-Grigorovich et al., 2003; Rögl & Spezzaferri, 2003; Bartakovics & Hudáčková, 2004; Ćorić et al., 2004; Gedl, 2004; Oszczypko-Clowes & Oszczypko, 2004, 2011; Spezzaferri, 2004; Kováčová & Hudáčková, 2005, 2009; Báldi, 2006; Tomanová Petrová & Švábenická, 2007; Báldi & Hohenegger, 2008; Kováč et al., 2008; Ozdínová, 2008[a,b]; Tóth & Görög, 2008; Zágoršek et al., 2008; Kováčová et al., 2009; Grunert et al., 2010[b], 2013; Hudáčková & Zlinská, 2010; Jamrich & Halásová, 2010; Koubová & Hudáčková, 2010; Peryt & Gedl, 2010; Gedl & Peryt, 2011; Kopecká, 2012; Zlinská et al., 2010; Gonera, 2013; Bitner et al., 2014; Doláková et al., 2014; Ozdínová & Soták, 2014; Peryt et al., 2014; Holcová et al., 2015[a,b]; Kaczmarek et al., 2016; Nehyba et al., 2016, and VIENNDAT (Hudáčková & Hudáček, 2004)).
Circulation regimes analysis modified from Krhovský (1981[a,b]), Kókay (1985), Brzobohatý (1987), Brukner-Wein et al. (1990), Krhovský et al. (1992), Oszczypko-Clowes (2001), Bicchi et al. (2003), Schulz et al. (2005), Báldi (2006), Kováčová et al. (2009), and this study.
Appearance percentage of the planktic and benthic foraminiferal taxa analysis based on Cicha et al. (1998), Holcová (2008), VIENNDAT database (Hudáčková & Hudáček, 2004). **Paratethys mollusc extinction and build-up events** after Harzhauser (2005). *Explanatory notes:* **(LOEE)** late Ottnangian extinction event; **(EBBE)** early Badenian build-up event; **(MBEE)** middle Badenian extinction event; **(BSEE)** Badenian–Sarmatian extinction event; **(SPEE)** Sarmatian–Pannonian extinction event.

The deposits assigned to the Karpatian and lowermost Badenian in the Western Carpathian realm yield similar Sr-ages around the 16 Ma (Kováč et al., 2017[a], Sant et al., 2017, and the new data). In spite of the short time span shown by the Sr-isotope method, these sediments of the same facies and age have been assigned to different regional stages. Concerning the validity of our Sr-isotope data, the geological record of the Novohrad-Nógrád Basin in southern Slovakia is continuous, or perhaps only with a negligible hiatus between the marine clays of the NN4 Zone (Čebovce site; Karpatian) and the overlying Lower Badenian carbonates with *Praeorbulina* in the oyster bed at the Trenč locality (Sr-age 16.02–15.86 Ma; Fordinál et al., 2014). These are followed by shallow water sediments belonging to the NN5 Zone at the locality Príbelce. The next example of this discrepancy concerns the clays of the NN4 Zone deposited in the Karpatian bathyal sedimentary environment at the north-eastern margin of the Vienna Basin (Cerová–Lieskové locality with Sr-age 16.1–15.3 Ma; Less et al., 2015). These are very similar in age and environment to the shallow bathyal sediments of the Vienna Basin western part (Sedlec HJ-2 borehole; Brzobohatý & Stráník, 2012), but these deposits are ranked to the Early Badenian. The deposits studied in the lower part of the borehole (Lower Lagenidae Zone; Brzobohatý & Stráník, 2012), can be correlated with sediments containing *Praeorbulina* of the Grund Formation in front of the Eastern Alps (Švábenická, 2002; Ćorić et al., 2004). As in the previous case, we can exclude, or possibly consider a negligible hiatus between the two bathyal depositional systems. In addition, in the Eastern Slovak Basin the evaporites positioned in the lower part of the Karpatian sedimentary record (Kováč et al., 1995) indicate isolation of the basin from the east and influence of an arid climate, where red beds sedimentation is documented in the eastern segment of the Carpathian Foredeep at the NN4/NN5 zones boundary (Kováč et al., 1996; Gozhyk et al., 2015; Oszczypko et al., 2016). Moreover, the upper part of the Karpatian sedimentary record in the western part of the Eastern Slovak Basin represents a continuous transition from the Karpatian to the Lower Badenian strata (Mirkovce Fm.; Kováč & Zlinská, 1998). So far, these facts have led us to conclude, that all of the aforementioned localities were deposited between 16–15 Ma, thus, near the boundary of the NN4/NN5 zones. In this time interval, we assume that in the Central Paratethys a change of circulation occurs from estuarine to anti-estuarine (Fig. 7). The general absence of *Praeorbulina* in the sediments in most basins of the Western Carpathians can be attributed to the alternation of water circulation regimes during this time. The exact dating of the Karpatian/Badenian boundary and pinning it in the stratigraphic chart is out of the scope of the present paper and remains a problem to solve it in the future.

The next problem of the regional stages stems from the biostratigraphy approach which often uses benthic organisms linked to a particular environment, shifting diachronously across the basin in time. The only true isochronous levels across the basins are based on planktic organisms (Fig. 2). The consequence of the "ecostratigraphical approach" in regional studies is the so-called "layer-cake stratigraphy", in which the isochronous bio-stratigraphical boundaries occur only in ideal cases. This type of approach can lead to the assignment of an incorrect age of sediments with similar facies development but deposited during various time spans. However, diachronous occurrence of some index fossils in the same sedimentary facies has often been documented (e.g., Molčíková, 1974). In addition, the isochronous coexistence of different associations of benthic organisms in shallow and deep water sedimentary environments is common (Hudáčková et al., 2013), due to basin floor topography and to different stages of development characterizing the depositional system.

The Late Miocene record of Lake Pannon can serve as a good example of the "layer-cake stratigraphy zonation" usually applied at the time (Papp, 1951, 1953), considering only an ideal case of basin fill, where the classic "law of superposition" rules. This "model" does not take into account the time-shift of shelf – shelf slope – basin floor depositional systems. Nowadays, we are aware of sequences of non-horizontal isochronous surfaces, where different associations of various ecological demands (e.g., depths) might live at the same time, but represent traditional "ecozones". These are not necessarily time-dependent, but follow the heterochronous shift of the environment (e.g., Magyar et al., 1999, 2013; Kováč et al., 2011; Šujan et al., 2016).

A similar phenomenon was also observed in the late Middle Miocene sediments of the Vienna Basin. The Sarmatian sedimentary record (based on 3D seismic data) was used for correlation of several boreholes, where foraminifera were studied. The result is that the traditionally used subdivision into the three successive zones: the large elphidia*, E. hauerinum,* and small miliolids zones (*sensu* Grill, 1941, 1943; Papp, 1951) is valid only in some specific cases. The heterochronity of assemblages, depending on the former basin topography, was clearly documented in the well cores and was also confirmed by the interpretation of seismic profiles. It was demonstrated, that the associations of each of these zones can represent a lateral shift of their habitat, and not a time succession of these biozones (Hudáčková et al., 2013).

The Karpatian stage was defined "ecostratigraphically" by common occurrence of *Uvigerina graciliformis* and the presence of the NN4 Zone (Brzobohatý et al., 2003) and the Early Badenian is often characterised by the presence of the NN5 Zone and the common occurrence of *Orbulina suturalis* (Hohenegger et al., 2014) in applied regional stratigraphy of the Western Carpathian basins. Taking into consideration the short time interval around the boundary of the NN4/NN5 zones and the close palaeogeographical connections among the basins of the Central Paratethys documented in Fig. 7, it can be assumed that the upper Karpatian–lowermost Badenian sediments in the Western Carpathians can be regarded as equivalents of the lower Langhian sedimentary record in the Mediterranean (~16–15 Ma). At present, these sediments are often placed into the late Karpatian or early Badenian, despite the common presence of very similar planktic and benthic fauna, as exemplified by gastropods (e.g., *Stellaria testigera*, *Galeodea echinophora*, *Mitrella hilberi*, *Amalda glandiformis*, *Ringicula minor*), bivalves (e.g., *Leionucula ehrlichi*, *Yoldia nitida*, *Parvamussium felsineum*, *Laternula fuchsia*), scaphopods (e.g., *Fissidentalium badense*), and decapods (*Retropluma slovenica, Styrioplax exiguus, Balsscallichirus florianus, Munidopsis lieskovensis, Munidopsis palmuelleri, Jaxea kuemeli, Necronectes schafferi*) collected in various localities of the Karpatian (NN4) or the Lower Badenian (NN5) sediments of the Pannonian Basin System (Harzhauser et al., 2011; Hyžný, 2011, 2016; Hyžný & Schlögl, 2011; Hyžný et al., 2014, 2015; Gašparič & Hyžný, 2015). These deposits can probably be both from a shelf and deep water with a pronounced diachronity of the same sedimentary facies.

(iii) The existence of the straits/gateways across the mountain chains was estimated due to indirect evidence about the faunal and floral immigrations from surrounding basins (e.g., Rögl, 1998; Steininger & Wessely, 2000; Palcu et al., 2015). The possibility of the origin of some stratigraphically important species in the Central Paratethys and their subsequent immigration to surrounding basins was also not excluded. For example, Švábenická (2002) suggested that *Helicosphaera waltrans* originated in the Central Paratethys and probably also *Helicosphaera walbersdorfensis*. In this regard, one of the most problematic issues was/is the connection between the Carpathian Foredeep and the hinterland basin system around the Lower/Middle Miocene boundary, which remains a matter of discussion and future research. What we know is that the earliest Badenian transgression (NN4 Zone; ~16 Ma), which can be correlated with the early Langhian sea level rise (Hardenbol et al., 1998), is known from the Styrian Basin in the south, but also from the foredeep at the junction of the Eastern Alps and Western Carpathians (Rögl, 1999; Ćorić et al., 2004; Hohenegger et al., 2014). According to the aforementioned facts, the assumed sea connections reach from the edge of the Central Western Carpathians via the Eastern Slovak Basin through the former fore-arc basin up to the foredeep, but also along the Carpathian Foredeep internal zone, which at present, is located below the front of the Outer Western Carpathian accretionary wedge (Fig. 7). These sediments are buried beneath the Western Carpathian thrust front in the north, or they became a part of the Outer Carpathian nappes in the east. This fact is supported by the continuous presence of the NN4 and NN5 zones in the Carpathian Foredeep units and in the platform sedimentary cover in the east (Popov et al., 2004; Gozhyk et al., 2015). Later, in the Badenian (NN5 Zone; ~15 Ma), a normal marine condition is documented from the whole Western Carpathian domain.

(iv) Most of the authors dealing with climate changes use values which have been set in the World Ocean and/or in other places of Eurasia (e.g., global cooling events Mi-1 to Mi-4, Fig. 11; Gradstein et al., 2012). As we can see in any current weather forecast for the Slovak Republic, the local climate of the Western Carpathians is noticeably affected by microclimatic factors. A distinguishable impact of the Eastern European Platform continental weather is an important element in the region. However, the area lying northwards from the Western

Carpathian mountain chain is often controlled by the Atlantic Ocean as opposed to the lowlands in the south-west occasionally influenced by the Mediterranean. Therefore, the Central Paratethys climate characterised with the aforementioned "global data" may be only informative and an evaluation of local pollen analyses and other climate indicators is necessary. Coal seams in the studied area often point to pronounced precipitation and warm climate, and they are a reaction to distinct regional causes such as mountain uplift. On the other hand, the evaporites in the Central Paratethys eastern part (foredeep) were deposited during the Early Miocene and not only at the Langhian/Serravallian boundary. Therefore, the recurrent evaporite deposition cannot be attributed exclusively to the global cooling events as proposed by de Leeuw et al. (2010) and other authors. In this case, evaporite sedimentation reflects the tectonic events which closed the marine connections between the basins. Moreover, these basins, rimming the grooving accretionary wedge of the Carpathians, were probably influenced by more arid climate of the Eastern European Platform (Chapter 4; Fig. 11). Altitudinal and latitudinal zonation of the palaeoflora indicates mountain uplift, but sediment provenance and transport direction of pollen (wind, rivers) have to be considered. A good example is the Early Miocene uplift of the Eastern Alps, which is not manifested in the Western Carpathian pollen spectra, while the Badenian uplift of the Western Carpathian mountain chain is perfectly visible (e.g., Kvaček et al., 2006).

Another frequent mistake is to apply the global sequence boundaries without considering any local factors (Fig. 11). On the regional scale, the considerable influence of factors such as tectonics and sediment input can strengthen or diminish the signals of the global sea level change. This is documented by studies from the Western Carpathian region (Vienna Basin, Danube Basin, and Eastern Slovak Basin) where each basin shows an individual sequence-stratigraphic record, often differing in the northern and southern parts of the same basin (e.g., Hudáčková et al., 1996; Kováč et al., 1998, 1999, 2004; Kováč, 2000).

7. Conclusions

The compiled palaeogeographical maps of the Central Paratethys Sea reveals the significant impact of geodynamic processes on the ratio of water and continental areas during the Cenozoic evolution of the Western Carpathians in selected time intervals (Figs. 4–10).

The interpreted water circulation regimes (Fig. 3), supplemented by Sr-isotope dating, helped us to understand the problem of correlation between the regional Central Paratethys and the standard Mediterranean stages, based on planktic organisms (Fig. 2). Encouraging results were obtained in periods with an anti-estuarine circulation regime, while estuarine circulation in the Central Paratethys led to inaccuracy, mainly while defining the Early/Middle Miocene boundary in this area.

The observed global trends of the climate changes were found to be obscured by tectonics and by local precipitation patterns, both resulting in the deposition of coal seams and evaporites in various parts of the Central Paratethys realm within several time intervals (Fig. 11). Nevertheless, dating of low oxic or evaporite events of the Central Paratethys can be used as time correlation levels in the Late Oligocene, Early Miocene, Langhian/Serravallian boundary, and Early Serravallian time intervals.

The presented palinspastic palaeogeographic maps of the selected time intervals, supplemented by the Central Paratethys Sea water circulation pattern changing in time, can be characterised and interpreted as follows:

(i) In the Kiscellian (~35–27 Ma; ~Rupelian–early Chattian) collision of the Alpine orogenic system with the European Platform led to the formation of a semi-closed Paratethys Sea. Its isolation is indicated by the euxinic regime (Fig. 3A). The area covered by the northern realm of the Central Paratethys Sea included the platform shelf and its slopes (passive margin), the remnant oceanic basins (flysch troughs), and a fore-arc basin located along the actively deformed margin of the overriding plate. The southern realm of this sea was undoubtedly a retro-arc basin (Fig. 4). During the early Egerian (Chattian; ~27–23 Ma) the opening of the western marine connection across the retro-arc basin is suggested with an estuarine circulation regime (Fig. 3B). Finally, in the late Egerian (~Aquitanian; ~23–21.5 Ma), for a short while, two seaways are suggested, via the remnant oceanic realm and via the retro-arc basin (Fig. 5) with restricted immigration of planktic organisms. An open circulation regime can be assumed (Fig. 3C).

(ii) During the Eggenburgian–Ottnangian (early Burdigalian; ~20.5–18 Ma), the final collision of the Eastern Alps with the platform led to the (re)opening of the western gateway in front of the Alps (Fig. 6). The tectonic north-eastward extrusion of the Central Western Carpathian segment was associated with growing of the accretionary wedge and initial rifting of the hinterland basin system, north of the Mid-Hungarian fault zone. A change to an anti-estuarine circulation regime can be assumed due to warming and less precipitation in the Central Paratethys (Fig. 3D). Isolation of the eastern segment of the Carpathian Foredeep led to evaporite deposition. Isochronous deposition of evaporites and coal seams in the southern realm of Central Paratethys (Fig. 6) indicate that precipitation was irregularly distributed. Uplift of the Alpine mountain range complexes can be the reason of fluctuation in humidity at the end of this period.

(iii) During the Karpatian–early Badenian time span (late Burdigalian–early Langhian; ~17–15 Ma), the oblique collision of the Western Carpathians with the platform led to growing of the Outer Carpathian accretionary wedge, which generated subsidence in the Carpathian Foredeep, and led to the opening of pull-apart depocenters of the Vienna and Eastern Slovak basins in the marginal zones of the hinterland area (Fig. 7). Gateway towards the east was closed, while marine connection through the "Trans-Tethyan-Trench-Corridor" with the Mediterranean opened in the south-west. The upwelling regimes documented in several basins of the Central Paratethys southern realm were the precursor of the change from an anti-estuarine to estuarine circulation regime, reconstructed for this time interval (Fig. 3E). Uplift of the Central Western Carpathian mountain chains can be the reason of fluctuation in humidity in their hinterland (coal seams; Fig. 7).

(iv) During the remainder of the Early Badenian (~Langhian; ~15–13.8 Ma), areas flooded by the Central Paratethys Sea again reached their maximal extent (Fig. 8). In front of the uplifted Western Carpathian mountain chain, the Carpathian Foredeep depocenters show a shift of maximal subsidence eastward, while in the back-arc basin area, an accelerated synrift subsidence started. Communication of the semi-closed sea with the Mediterranean across the "Trans-Tethyan-Trench-Corridor" changed to an anti-estuarine circulation (Fig. 3F), likely due to aridification and salinity increase in surface water. At the Early/Late Badenian boundary (Langhian/Serravallian), the evaporite deposition took place in the northern and eastern parts of Central Paratethys, referred as Badenian Salinity Crisis (BSC) in the Carpathian Foredeep, Transcarpathian, and Transylvanian basins.

(v) After BSC, at the beginning of the Late Badenian–Sarmatian time interval, (~Serravallian; ~13.5–11.6 Ma) a partial isolation of the Central Paratethys (Fig. 9) led to low oxic conditions (Fig. 3G) in the whole area. However, an occasional open circulation regime with the Mediterranean and Eastern Paratethys realm could be expected. Even during the Late Badenian (13.8–12.6 Ma), an anti-estuarine circulation (Fig. 3G) supported the water exchange with the eastern domain. The water masses from the Eastern Paratethys entered the semi-closed basin system again in the Sarmatian (12.6–11.6 Ma). This regime is assumed only for a short period of time (Fig. 3H). During this time span, a gradual fill up of the Western Carpathian Foredeep is documented (Fig. 9). Uplift of the Western Carpathians can be the reason of humidity fluctuations in this area (coal seams).

(vi) In the early Pannonian (early Tortonian; ~11.6–10 Ma), sedimentation in the Western Carpathian Foredeep ceased and in the hinterland, a new phase of subsidence led to the formation of the Pannonian Basin System – called Lake Pannon (Fig. 10). A very short opening of the eastern gateway can be assumed at the beginning of this interval proved by the presence of the endemic nannoplankton (Fig. 2), while total isolation is presumed later (Fig. 3I).

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