

Miocene Central Paratethys stratigraphy – current status and future directions

Werner E. Piller¹, Mathias Harzhauser² and Oleg Mandic²

¹*Institute for Earth Sciences (Geology and Paleontology), University of Graz, Heinrichstrasse 26, A-8010 Graz, Austria*

²*Museum of Natural History Vienna, Geological-Paleontological Department, Burgring 7, A-1014 Vienna, Austria*
email: werner.piller@uni-graz.at

ABSTRACT: The complex geodynamic history of the Paratethys periodically fostered the evolution of a highly endemic biota with only limited exchange between the neighboring Mediterranean and Indo-Pacific provinces. The resulting very peculiar fossil assemblages forced the introduction of a regional chronostratigraphic subdivision for the Western/Central and Eastern Paratethys respectively. For the Central Paratethys we present a summarized and updated database for the individual stages, and we review the current status for correlation with the Mediterranean stratigraphic framework. The Miocene Central Paratethys stages were defined on exclusively paleontological criteria in type sections (holostratotypes and faciostratotypes). They are all bounded by either sedimentary hiatuses or distinct facies changes, inferred to mark lowstands in sea level, and not a single boundary stratotype has been defined. Some correlating tie-points to the Mediterranean succession are based on calcareous nannoplankton and planktonic foraminifers; magnetostratigraphic correlation is very limited. All stages can be assigned to the putatively third-order sea level cycles, with the Eggenburgian, Badenian and Pannonian Stages each spanning three cycles and the Ottnangian, Karpatian, and Sarmatian one each. The Karpatian/Badenian boundary correlates with the Burdigalian/Langhian (Early/Middle Miocene) boundary, and the Sarmatian/Pannonian boundary correlates with the Serravallian/Tortonian (Middle/Late Miocene) boundary. The correlation to third-order cycles and the detection of astronomical signals suggest that not only a regional but also a strong global signal is present in the rock record of the Central Paratethys. Since the current definition of a stage includes its global spread, formally defined regional stages are redundant and therefore also not necessary for the Central Paratethys. However, if stages are essentially regional, then a regional scale as for Central Paratethys would be much more appropriate.

INTRODUCTION

During the Cenozoic Era Africa moved towards Eurasia with a northwards shift and a generally counterclockwise rotation involving several microplates in the Mediterranean area (Kováč et al. 1998; Márton et al. 2003, 2006; Márton 2006; Seghedi et al. 2004). As a consequence, Eurasian paleogeography changed dramatically from vast marine areas interrupted by archipelagos into dry land. This increasing degree of continentalisation was accompanied by the rise of the Alpidic chains which intensively structured topography. Around the Eocene/Oligocene boundary Africa's northward movement and resulting European plate subduction caused the final disintegration of the ancient (Western) Tethys Ocean (Báldi 1980; Harzhauser et al. 2002; Harzhauser and Pillar 2007). The Indo-Pacific Ocean came into existence in the east and various relic marine basins remained in the west. Along with the emerging early Mediterranean Sea, another heritage of the vanishing Tethys was the vast Eurasian Paratethys Sea.

The recognition of the Paratethys as a biogeographic entity which differs from the Neogene Mediterranean goes back to Laskarev (1924). He proposed the existence of this lost sea on the ground of the peculiar character of the mollusc fauna after thoroughly investigating the Vienna, Styrian, Pannonian, Dacian, and Euxinian basins. During its maximum extent the Paratethys spread from the Rhône Basin in France towards Inner Asia. It was segregated into three paleogeographic and geotectonic units (not only two as sometimes reported, e.g., Nevesskaya

1999; Vasiliev et al. 2004, 2005) each recording a different environmental history. The smaller western part consists of the Western and the Central Paratethys being opposed by the larger Eastern Paratethys. The Western Paratethys comprises the Alpine Foreland Basins of France, Switzerland, S Germany and Upper Austria (Senes 1961). The Central Paratethys includes the Eastern Alpine - Carpathian Foreland basins, from Lower Austria to Moldavia, and the Pannonian Basin System. The Eastern Paratethys comprises the Euxinian (Black Sea), Caspian and Aral Sea basins (Nevesskaja et al. 1993). The eastern Carpathian Foreland transforms towards the end of the Middle Miocene, switching from the Central Paratethys into the Eastern Paratethys geo- and hydrodynamic regime. This event coincides with the disintegration of the Central Paratethys triggered by the installation of the Late Miocene Lake Pannon that became restricted to the Pannonian Basin System (Magyar et al. 1999b).

Eurasian ecosystems and landscapes were impacted by a complex pattern of changing seaways and landbridges between the Paratethys, the North Sea and the Mediterranean as well as the western Indo-Pacific (e.g., Rögl and Steininger 1983; Rögl 1998a, 1999; Popov et al. 2004). Senes and Marinescu (1974) and Rusu (1988) perceived four stages in the geodynamic history of the Paratethys. In succession they are Proto-Paratethys, formed in the Late Eocene to Early Oligocene by the initial isolation from the open oceans; Eo-Paratethys (Late Oligocene and Early Miocene); Meso-Paratethys (late Early Miocene to early Middle Miocene); and Neo-Paratethys (later Middle to Late Miocene). (See also Steininger and Wessely 2000.)

This geodynamically controlled paleogeographic and biogeographic differentiation caused major difficulties in the stratigraphic correlation between the Paratethys and the Mediterranean and global stratigraphy respectively. These problems led to the establishment of regional chronostratigraphic and geochronologic scales, which have been extensively documented for the Central Paratethys in the series “Chronostratigraphie und Neostatotypen” which distinctly improved the general knowledge on the Neogene Central Paratethys and its stratigraphy (Cicha et al. 1967; Steininger and Seneš 1971; Báldi and Seneš 1975; Papp et al. 1973, 1974, 1978, 1985; Stevanovic et al. 1990). The different geodynamic, paleogeographic and paleobiogeographic histories within the Paratethys itself, led to definition of chronostratigraphic/geochronologic scales for the Western, and for the Eastern Paratethys, additional to that of the Central Paratethys (e.g., Steininger et al. 1976; Rögl 1996; Popov et al. 2004). In this overview we mainly focus on the Miocene stratigraphy of the Central Paratethys and its correlation to the Mediterranean area (fig. 1).

REGIONAL CHRONOSTRATIGRAPHY AND CORRELATION

The development of prolonged anoxic bottom conditions during the Early Kiscellian (Early Oligocene, cf. Baldi 1986) marks the birth of the Paratethys (e.g., Schulz et al. 2005). As a consequence, black shales (“Fischschiefer”) developed in the Alpine foreland basin, the bituminous, laminated Tard Clay was deposited in the Hungarian basin and menilites in the Carpathian Flysch trough (Báldi 1998). In response to this event, a first endemic mollusc fauna evolved whilst spreading from the Asian Eastern Paratethys towards the west (Popov et al. 1985). This peculiar Solenovian fauna characterizes the Eastern Paratethyan Solenovian Stage. Environmental chemistry – probably brackish water conditions – within the vast inland sea triggered a blooming and rapidly evolving, highly endemic bivalve fauna with genera such as *Janschinella*, *Korobkoviella* and *Ergenica* (Popov et al. 1985; Nevešskaja et al. 1987). The accompanying, monospecific, nannoplankton and diatom blooms also point to reduced salinities and cool-temperate surface waters extending from Bavaria to Transcaspiá (Rögl 1998a).

Late Oligocene – Early Miocene

Egerian stage

The stage was first defined by Báldi (1969) and described in detail by Báldi and Seneš (1975). Its stratotype (holostratotype) was defined at Eger (Wind’s brickyard) in northern Hungary (fig. 2, Báldi 1975; Baldi et al. 1999). At the type-locality the base of the stage is marked by an abrupt lithological change from Kiscell Clay to glauconitic sandstone. This level coincides with the first occurrence (FOD) of *Costellamussiopecten pasini* (Meneghini) (= *Flabellipecten burdigalensis* Baldi, non Lamarck). Generally, the base is defined with the first occurrences of the benthic foraminifer *Miogypsina* (*Miogypsinoidea*) *complanata* Schlumberger and the planktonic foraminifer *Globigerinoides*. Several species of molluscs also occur for the first time (e.g., *Palliolium incomparabile* (Risso), *Costellamussiopecten schreteri* (Noszky), *Laevicardium cyprium* (Brocchi), and *Turritella beyrichi* Hoffman). The stratotype is truncated by an unconformity (Baldi 1975, pp. 100, 110–111). In some of the faciostratotype-sections (Budafok-2, Hungary; Orlek, Slovenia), Egerian beds grade into Eggenburgian sedi-

ments. The boundary coincides with a lithological change which implies distinct shallowing. In other faciostratotype-sections (Máriaalom, Hungary; Kováčov, Slovakia) the top of the Egerian is missing due to an erosional unconformity.

Facies: Sedimentologically and lithologically the Egerian is a continuation of the mainly siliciclastic depositional systems of the Oligocene Kiscellian with predominantly silty-clayey sediments. Carbonate formation is subordinate throughout but mixed carbonate-siliciclastic systems occurred. These are dominated by corallinaceans, bryozoans and larger benthic foraminifers such as miogypsinids and lepidocyclinids (Vanova 1975; Báldi 1986; Baldi et al. 1999; Kaiser et al. 2001).

Correlation: This stage straddles the Oligocene/Miocene boundary (Baldi and Seneš 1975) in comprising the upper part of the Chattian and the lower part of the Aquitanian (fig. 1). As pointed out by Baldi et al. (1999) the distribution of larger benthic foraminifers implies a correlation of its lower boundary with the lower boundary of the Shallow Benthic Zone SBZ 22, that is calibrated in the Mediterranean and NE Atlantic with the base of the planktonic foraminiferal zone P22 (Cahuzac and Poignant 1997). Moreover, the recalibration of 3rd order sea level sequences supported by the biostratigraphic results of Mandic and Steininger (2003), implies the position of the upper Egerian boundary in the mid-Aquitainian, and not at its top (fig. 1). Although suggested already by Hungarian stratigraphers (e.g., Baldi et al. 1999), this interpretation contrasts substantially with the current stratigraphic concept (e.g., Rögl et al. 1979; Rögl and Steininger 1983; Steininger et al. 1985; Vakarcs et al. 1998; Rögl 1998b; Mandic and Steininger 2003). The Paleogene/Neogene boundary is difficult to detect in the Central Paratethys since the index fossil for the Aquitanian, *Paragloborotalia kugleri*, is absent. Correlations are usually based on calcareous nannofossils including uppermost NP 24 to NN 1/2 nannozones (Rögl 1998b). In addition, *Miogypsina* species are very useful for biostratigraphic correlation. Whereas the lower Egerian deposits belong to SBZ 23 the upper Egerian limestones with *Miogypsina gunteri* found at Bretka (E Slovakia) (Baldi and Seneš 1975) belong to the lower part of SBZ 24 and thus correspond to the lower Aquitanian (Cahuzac and Poignant 1997). Consequently, in terms of sequence stratigraphy the Egerian/Eggenburgian boundary corresponds with the Aq 2 sea level lowstand of Hardenbol et al. (1998). The following 3rd order transgression-regression cycle already includes Eggenburgian deposits (see below). This interpretation is in accordance with the general regressive trend in the upper Egerian sediments and with erosional unconformities frequently forming their top. In continuous sections the sediments at the boundary were often deposited in very shallow water environments characterized by brackish water faunas. Continuous deep marine sections are only known from the strongly tectonised thrust sheets of the Outer West Carpathians and their equivalents (Krhovský et al. 2001).

Paleogeography: In the Late Oligocene the Paratethys was a huge, west-east oriented sea (fig. 3A). New gateways towards the Western Tethys opened and normal marine conditions were re-established after the anoxia during the Kiscellian (see above). The connection towards the North Sea Basin was open via the Rhine Graben and a connection to the Venetian Basin opened in the southwest (Rögl 1998a; Reichenbacher 2000). This trans-European connection of the Rhine Graben and Maince Basin with the Tethyan Rhône-Bresse Graben and the Paratethyan Alpine Foreland ceased during the late Egerian (Reichenbacher 2000).

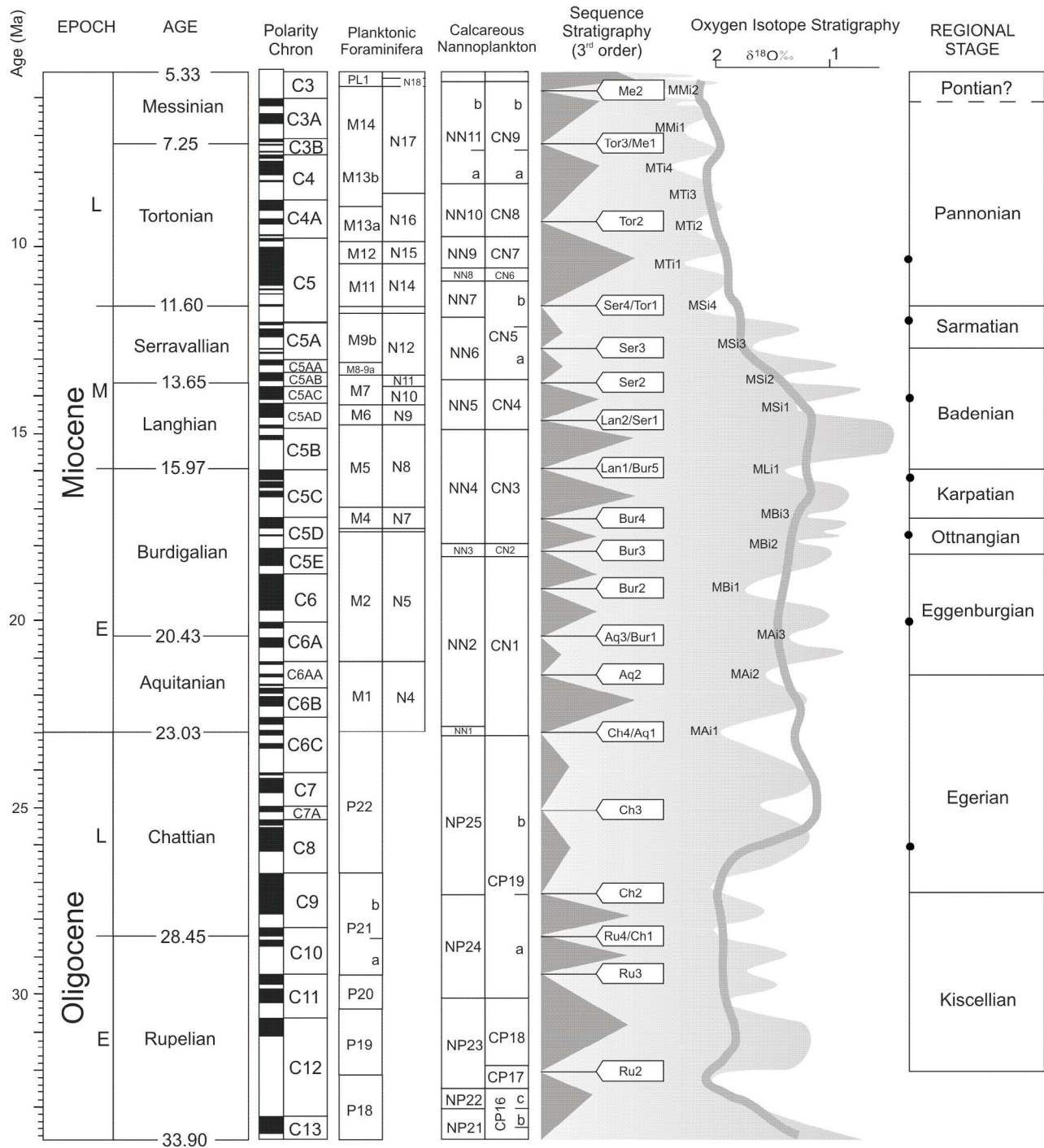


FIGURE 1
 Oligocene – Miocene geochronology, geomagnetic polarity chrons, biozonations of planktonic foraminifers and calcareous nannoplankton (all after Lourens et al. 2004), sequence stratigraphy and sea level curve (after Hardenbol et al. 1998), and oxygen isotope stratigraphy (after Abreu and Haddad 1998) partly recalibrated and correlated to regional chronostratigraphy of the Central Paratethys. The black dots on the right column indicate the stratigraphic position of the holostratotypes of the regional stages.

Paralic coal basins and freshwater environments developed in the westernmost reaches of the Paratethys (Barthelt 1989; Berger 1996), while the seaways on top of the still partly submerged Alpine nappes into the Western Tethys remained open (Wagner 1996; Steininger and Wessely 2000).

Eggenburgian Stage

The Eggenburgian Stage was defined by Steininger and Seneš (1971, p. 45-46). The stratotype Eggenburgian is located in NE Austria at Loibersdorf some 60 km NW from Vienna (fig. 2; Steininger 1971). Nowadays it outcrops poorly and a continuous section is not available. The base of the Eggenburgian in the area of the stratotype section is transgressive on Palaeogene terrestrial-fluvial-limnic sediments or on crystalline basement rocks. Due to a complex paleotopography in the Eggenburgian type region this transgressive development causes a heterochronous onset of Eggenburgian sediments. This complex evolution was more recently unravelled by a statistically based mollusc stratigraphy by Mandic and Steininger (2003). In the entire Central Paratethys an erosional gap is frequently developed at the base of the stage and in all examples the basal sediments reflect a transgressive pattern (Rögl and Steininger 1983).

The biostratigraphic frame of the Eggenburgian is based largely in its characteristic mollusc fauna with large-sized taxa, in particular pectinids and cardiids (Steininger and Seneš 1971). The base is marked by the first occurrence of *Oopecten gigas* (Schlotheim), the top (base Ottngian Stage) by the first occurrence of *Pecten hermansenni* (Dunker). A subdivision into lower, middle and upper Eggenburgian is based on mollusc biostratigraphy (Mandic and Steininger 2003). The lower Eggenburgian is defined biostratigraphically by the total range of *Rudicardium grande* (Hoelzl), the middle Eggenburgian by the total range of *Laevicardium kuebecki* (Hörnes). The total range of *Oopecten gigas* spans both biostratigraphic units. The upper Eggenburgian is defined by the FOD of *Gigantopecten holgeri* (Geinitz) and *Flexopecten palmatus* (Lamarck).

Benthic foraminifers are of lesser biostratigraphic importance, e.g., the first occurrences of *Elphidium ortenburgerse* Egger, *E. felsense* Papp, and *Uvigerina posthantkeni* Papp. *Miogypsina intermedia* Droger is reported from the Austrian Molasse Basin (Papp 1960). The ostracod genus *Falunia* Gerckoff and Moyes and calcareous nannoplankton taxon *Helicosphaera ampliaperta* occur for the first time.

Facies: The majority of well studied Eggenburgian sediments come from shallow water depositional environments. The Eggenburgian is dominated by sandy and pelitic near-shore sedimentation. Carbonates are scarce, patchy and usually of hybrid character. A typical example are the shallow marine corallinean rhodolite carpets associated with fine to medium sand, inhabited by the scutellid echinoid *Parmulechinus hoebarthi* (Kühn) in the Horn Basin of northern Austria (“Scutellensande”) (Steininger 1971; Kroh 2005). In deeper bathyal settings typical grey calcareous clays with intercalations of sands – the so-called “Schlier” – developed. Only in the Outer Carpathians did a relic Flysch trough remain with prevailing turbiditic sequences (Báldi 1998; Popov et al. 2004).

Correlation: Supraregional correlation is possible by a few tie points only: calcareous nannoplankton clearly indicates the presence of zones NN2 and NN3 (Steininger et al. 1976;

Roetzel et al. 1999) and the mammal fauna places it into the European land mammal zone MN3 (Mein 1989; Steininger et al. 1996; Steininger 1999). The latter zone is detected in the upper Eggenburgian sediments of the type region bearing *Gigantopecten holgeri*. The lower part of the Eggenburgian is correlated with the upper MN2 zone. In terms of sequence stratigraphy and sea level changes the general stratigraphic development of the Eggenburgian coincides to three 3rd order sea level changes. These can be correlated with the Aq 2 lowstand, marking the base of the Eggenburgian, Aq 3/Bur 1 and Bur 2, and end with the Bur 3 lowstand (top Eggenburgian/base Ottngian) of Hardenbol et al. (1998). The mollusc fauna of the basal Eggenburgian (which is correlated herein with the upper Aquitanian), bears Oligocene relics of northern origin (e.g., *Drepanocheilus speciosus*) (Steininger 1963), whereas the tropical fauna of the middle Eggenburgian is correlated herein with the transgression of the Bur 1 sequence (fig. 1; Mandic et al. 2004). The Bur 2 lowstand (Hardenbol et al. 1998) at the base of the upper Eggenburgian is marked by a prominent erosional surface and reworking of basement rocks. The fossil assemblages in upper Eggenburgian deposits reflect a substantial faunal turnover marked by numerous first occurrences of species with proto-Mediterranean origin (Mandic and Steininger 2003). This is interpreted as a consequence of a prominent flooding event allowing the faunal migration from the latter region. Finally, the next prominent erosional surface – topping the upper Eggenburgian siliciclastics – is correlated herein with the Bur 3 lowstand and the base of the Ottngian. The lower Ottngian sediments differ distinctly due to the onset of a warm-temperate carbonate factory, indicated by bryozoan-corallinean limestones (Zogelsdorf Formation) (Nebelsick 1989). Larger benthic foraminifers (*Amphistegina*) and hermatypic corals are subordinate, the latter form only very small patches. Contrary to Vakarcz et al. (1998) we consider the major sea level lowstand at the base of the Eggenburgian to be equivalent not to Aq 3/Bur 1 but to Aq 2, which accords better with calcareous nannoplankton, mammal and mollusc data (Steininger et al. 1976; Mandic and Steininger 2003).

Paleogeography: Broad connections into the Eastern Paratethys allowed the spreading of the middle Eggenburgian mollusc faunas with *Laevicardium kuebecki* as far east as the Crimean Peninsula and Georgia (Rögl 1998a). In addition, the western seaway via the Alpine Foreland, which was sealed during the late Egerian and maybe also during the earlier Eggenburgian, started to open. The sea invaded the foreland successively from the west and entered the Central Paratethys (Berger 1996) at last with the late Eggenburgian. This newly established marine pathway of the Paratethys via the Alpine Foredeep into the Rhône Basin probably coincided with a hypothetical second flow from the Eastern Mediterranean (Martel et al. 1994). These connections gave rise to a new hydrodynamic regime reflected in the meso- and macrotidally controlled deposits throughout the Alpine Foreland basins lasting from the late Eggenburgian to the middle Ottngian (Allen et al. 1985; Faupl and Roetzel 1990). A second area of tidal deposits is described by Sztanó (1995) from the Eggenburgian of the North Hungarian Bay.

Faunistically, the changes in paleogeography are reflected by the immigration of western Mediterranean taxa such as the echinoid *Arbacina catenata* (Desor) (Kroh and Harzhauser 1999) and several bryozoans (Vavra 1979). Among molluscs, the prominent faunistic overturn is marked by the introduction of Burdigalian Mediterranean pectinids such as *Flexopecten*



FIGURE 2
Geographic distribution of all Miocene stratotype localities of Central Paratethys stages.

palmatus and *Gigantopecten holgeri* going along with the extinction of the Eggenburgian endemics such as *Oopecten gigas* and *Laevicardium kuebecki* (Mandic and Steininger 2003). Another important immigration is represented by the fossil sea cow *Metaxytherium krahuletzki* being conspicuously common in the upper Eggenburgian but absent in older horizons. All that happens distinctly prior to the onset of the warm-temperate carbonate production on top of upper Eggenburgian siliciclastics (Roetzel et al. 1999). The onset of the warm-temperate carbonate factory is probably coeval with the slight cooling indicated by Zachos et al. (2001) and by the MBi-2 isotope event of Abreu and Haddad (1998) (Fig. 1). The loss of tropical mollusc taxa between the middle and the upper Eggenburgian could coincide with the MBi-1 isotope event.

Ottnangian Stage

The stratotype is defined in a clay pit near the village of Ottnang in Upper Austria (Rögl et al. 1973, fig. 2). The type section is characterized by the onset of well-bedded, blue-grey, fine sandy, micaceous claymarls (locally called “Ottnang Schlier”) which are underlain by fine to medium grained quartz sands (“Atzbach Sands”). The base of the Ottnangian is not defined in this section, the top is cut by erosion. In basinal sections of the Alpine foredeep, sedimentation is considered to be continuous from the Eggenburgian into the Ottnangian, whereas in more eastern locations (e.g., Hungary) the base is marked by a disconformity. In Slovakia and northern Hungary, seemingly continuous sedimentation from the Ottnangian into the Karpatian is reported.

Generally, the Ottnangian is a strictly twofold stage with a normal marine development in its lower part and a predominance of restricted marine to fresh water environments in its upper part. The most characteristic and important biota are marine molluscs, partly of boreal affinity, but mainly of Paratethyan Eggenburgian origin. Some of the Ottnangian mollusc faunal elements are of biostratigraphic importance: the FOD of *Pecten hermansenni* (Dunker) mark the base of the stage. The foraminiferal fauna is very similar to that of the Eggenburgian (Harzhauser and Piller 2007). Among planktonic taxa *Cassigerinella spinata* Rögl occurs and *Globigerina ottnangiensis* Rögl is abundant. Among benthic foraminifers *Sigmoilopsis ottnangensis*, *Bolivina matejkai*, *B. scitula*, and *Amphicoryna ottnangensis* are characteristic, as also *Pappina primiformis* and *Pappina breviformis* (Steininger et al. 1976; Cicha et al. 1998). The late Ottnangian is characterized by the occurrence of a highly distinct endemic bivalve fauna, the so called “*Rzehakia* fauna” (=synonymous to “*Oncophora* fauna”; cf. Senes 1973). This consists of endemic genera such as *Rzehakia* and *Limnopageta*, which offer an excellent correlation tool within Paratethyan deposits (Ctyroky 1972; Mandic and Coric 2007).

Facies: In the lower Ottnangian, sedimentation is dominated by siliciclastics with widespread tidally influenced deposits and the characteristic sandy/silty “Schlier” sediments (Faupl and Roetzel 1987, 1990). The warm-temperate bryozoan-coral-linacean limestones are known only from the Eggenburgian region (Zogelsdorf Fm.), which have been considered up till now to be of late Eggenburgian age (see discussion above). The car-

bonates previously correlated with the upper Ottnangian from the Hungarian Bántapuszta section (Kóky 1973) recently have been correlated with the Karpatian (Mandic 2003). The shift was due to reinterpreting biostratigraphic data from pectinid bivalves, foraminifers and calcareous nannoplankton, and to the reconsideration of the regional geologic history. On the evidence of silicoflagellate assemblages and the frequent occurrence of diatomites, Bachmann (1973) favoured temperate water conditions for the lower Ottnangian deposits.

The upper part of the Ottnangian is represented by fluvial-lacustrine environments of the Upper Freshwater Molasse in the Western Alpine Foreland Basin (Berger 1996). With the exception of the Northern Alpine Foreland Basin and its continuation into the Polish foredeep, no fully marine environments are known from the Carpathian-Pannonian-Dinaride domain where brackish to freshwater sedimentary environments prevailed (Kováč et al. 2004; Kotlarczyk et al. 2006). Consequently, during the late Ottnangian and the synchronous Kotsakhurian in the Eastern Paratethys biogeographic relations between the Paratethys and the Mediterranean Sea ceased. This Ottnangian crisis is reflected in nearshore settings by brackish water conditions and a sudden evolutionary peak in bivalves, resulting in a large number of endemic genera of the so-called “*Rzehakia* fauna” (see above). This fauna expanded from the Eastern Paratethys into the Central and Western Paratethys Sea during the late Ottnangian (Steininger 1973).

Correlation: The Ottnangian was differentiated because of a regressional phase at the end of the Eggenburgian (Senes 1973), inferred to be due to tectonic movements particularly effective in the Carpathian area. These tectonic activities, however, enhanced a global sea level fall at the beginning of the Ottnangian which can be correlated with the TB 2.1. cycle of Haq et al. (1988) and represents Bur 3 of Hardenbol et al. (1998). The Ottnangian therefore corresponds to only one 3rd order sea level change (Kováč et al. 2004). Biostratigraphic correlation outside Paratethys is very limited. Within the Ottnangian the foraminiferal genus *Catapsydrax* occurs for the last time. Rögl (1998b) linked this event with the LAD of *C. unicavus/C. dissimilis*, which defines the boundary between M3 and M4 of Berggren et al. (1995). In terms of nannoplankton stratigraphy zones NN3 and NN4 are represented (Steininger et al. 1976; Rögl et al. 2003a). Magnetostratigraphic correlation points to a rough correspondence of the entire Ottnangian to Chron C5D.

Paleogeography: During the early Ottnangian the paleogeographic configuration remains similar to that during the Eggenburgian, but in the late Ottnangian the uplift of the Alpine Foreland Basin terminated the western connection to the Mediterranean (Rögl 1998a). In addition, the sea level fall during the Early Miocene global sea level cycle TB 2.1. (Haq et al. 1988) accentuates the beginning isolation of the Paratethys from the Mediterranean Sea during the late Ottnangian. Geographic differences within the endemic “*Rzehakia* fauna” between Bavaria, Austria and Moravia might indicate a further disintegration of the Paratethys into several isolated brackish lakes (Mandic and Coric 2007).

Karpatian Stage

This stage was erected by Cicha and Tejkal (1959) and defined by Cicha et al. (1967) in the first volume of the series *Chronostratigraphie und Neostatotypen* based on the stratotype section Slup (fig. 2) in Moravia (Czech Republic). Brzobohatý et al. (2003) updated this volume with a wealth of new data. The

stratotype section is characterized by bedded, fine-grained sands and lenses of coarse sands with a rich molluscan fauna. Its base is marked by an unconformity, forcing Cicha and Rögl (2003) to issue a “Provisional definition of the Karpatian”, in the updated volume. This discontinuity occurs in all shallow marine settings (Rögl et al. 2003a). Continuous sedimentation between the Ottnangian and the Karpatian is expected only in deeper parts of Central Paratethys basins particularly in the Pannonian realm (Cicha and Rögl 2003).

Originally, the stage was established to document the surge of new faunal elements from the Mediterranean at its base. Some of the molluscs occur for the first time in the Karpatian (e.g., *Conus steindachneri*, *Thais exilis*, *Gyrineum depressum*, *Acanthocardia paucicostata*, *Cerastoderma arcella*, *Ervilia pusilla*, *Paradonax intermedia*), most, however, continue into the Badenian (Harzhauser 2002; Harzhauser et al. 2003). Due to this continuation, differentiating Karpatian and Badenian gastropod assemblages is sometimes difficult (Harzhauser et al. 2003). Restricted to the Karpatian are *Modiolus excellens* Csepregy-Meznerics and *Maetra (Barymaetra) nogradensis* Csepregy-Meznerics (Mandic 2003).

The stage is defined biostratigraphically with the FAD of *Uvigerina graciliformis* Papp and Turnowsky (Papp et al. 1971). Several other uvigerinids co-occur, such as *Pappina primiformis*, *P. breviformis* and *Uvigerina acuminata*. In general, foraminifers exhibit a relatively great number of FODs (Harzhauser and Piller 2007), with planktonic taxa less diverse. The most important planktonic foraminiferal event is the FOD of *Globigerinoides bisphericus* Todd in the upper Karpatian. Calcareous nannoplankton floras are characterized by *Helicosphaera ampliaperta*, *H. carteri*, *H. mediterranea*, *Reticulofenestra pseudoumbilica*, *Sphenolithus heteromorphus*, and *Pontosphaera multipora* (Steininger et al. 1976; Švábenická et al. 2003). Soliman and Piller (2007) described a low-diversity dinoflagellate association with dominant *Operculodinium centrocarpum*, *Lingulodinium machaerophorum*, *Reticulatosphaera actinocoronata* and *Spiniferites* spp.

Facies: The base of the Karpatian sequences is represented mainly by terrestrial, alluvial, fluvial, and deltaic deposits which upsection pass rapidly into marine, neritic to shallow bathyal sediments. Sedimentation is dominated by green-blue and grey pelites and silty calcareous shales in offshore environments and clayey sand in marginal areas. The lower Karpatian has still similarities with the Ottnangian, pointing to cool-temperate water masses with high numbers of siliceous fossils (Rögl et al. 2003b). Suboxic bottom conditions in the basins, upwelling and temperate water are also suggested based on planktonic foraminifers (Cicha et al. 2003). Carbonates as known from the Hungarian Bántapuszta section (Kóky 1973) are scarce and correspond in composition to the lower Ottnangian corallinacean-bryozoan type. Warmer water indicators, such as *Globigerinoides* or *Globorotalia*, appear in the late Karpatian together with a thermophilic mollusc fauna (Harzhauser et al. 2003).

Correlation: After a long history of misinterpretations and miscorrelations (for a more recent compilation see Harzhauser et al. 2003) the Karpatian Stage is nowadays consistently considered to be time-equivalent to the latest Burdigalian. Although the base of the stage cannot be biostratigraphically tightened, the calcareous nannoplankton flora with the co-occurrence of *Helicosphaera ampliaperta* and *Sphenolithus heteromorphus*

places the entire Karpatian record in calcareous nannoplankton zone NN4. The occurrence of *Globigerinoides bisphericus* in the upper part of the Karpatian allows a correlation with foraminiferal zone M4b of Berggren et al. (1995) and also places it in the latest Burdigalian.

Within this biostratigraphic frame, the unconformity at the base of the Karpatian and the following transgression can be correlated with the sea level rise at the beginning of the global 3rd order sea level cycle TB 2.2. of Haq et al. (1988) and Bur 4 of Hardenbol et al. (1998). The Karpatian/Badenian (Burdigalian/Langhian) boundary is characterised by a significant sea level drop (Haq et al. 1988; Hardenbol et al. 1998), expressed as a hiatus traceable throughout the Central Paratethys (Rögl et al. 2002). Continuous sedimentation from Karpatian to Badenian has never been observed. The top of the Lower Miocene in the Paratethyan basins is marked by erosional surfaces or by an angular discordance between the Lower and Middle Miocene strata, frequently called the “Styrian unconformity” (Stille 1924; Latal and Piller 2003). As a consequence, the Karpatian matches only one 3rd order sea level cycle (TB 2.2., Bur 4 as base).

Paleogeography: The Karpatian starts with a transgression and a reorganisation of the paleogeographic pattern (Rögl et al. 2003b). The northward migration of a variety of biota was favoured by a general warming trend and by a new broad connection with the Mediterranean that established via the Slovenian “Trans-Tethyan Trench Corridor” (Bisticic and Jenko 1985). This seaway enabled a free faunal exchange between the Central Paratethys and the Mediterranean area. This change in environment is adjoined by a dramatic tectonic turnover in the Central Paratethys area leading to a change from W-E trending basins towards intra-mountain basins (Rögl and Steininger 1983; Rögl 1998a; Kovác et al. 2003). A typical example for the geodynamic impact is the abrupt, discordant progradation of upper Karpatian estuarine to shallow marine deposits over lower Karpatian offshore clays in the Alpine Foreland Basin and in the Carpathian Foredeep (Adámek et al. 2003). The widespread formation of evaporites in the Rumanian part of the Carpathian Foredeep and in the Transylvanian Basin points to a poor or even absent connection with the Eastern Paratethys during the latest Early Miocene.

FIGURE 3 →
Paleogeographic sketch-maps of the Mediterranean – Central Paratethys region (grey: land; white: water):

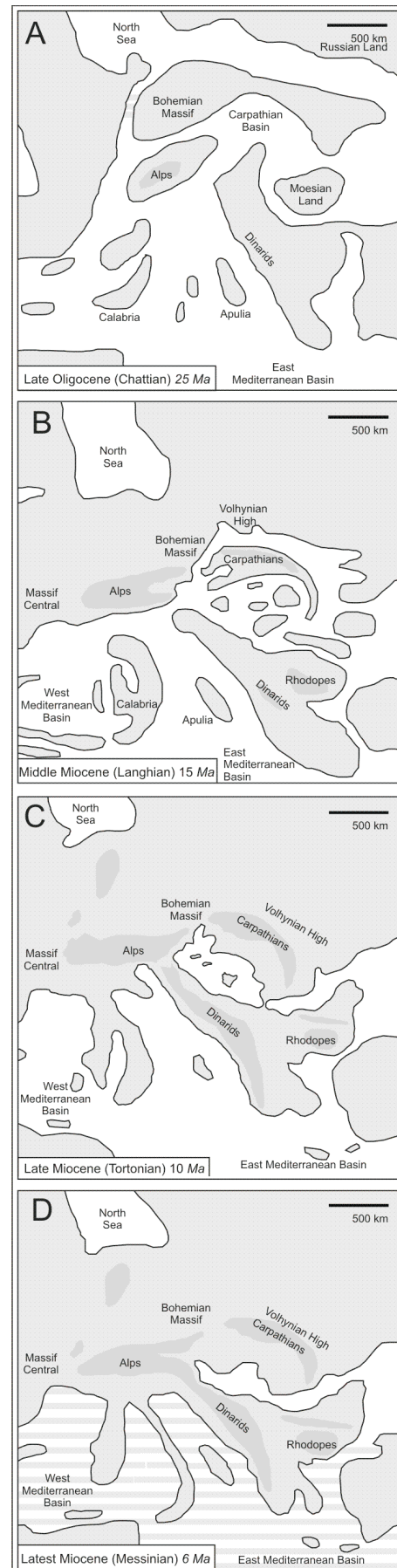
A: During the Late Oligocene good marine connections with broad seaways between the Central Paratethys, Mediterranean and North Sea were present. The Central Paratethys was a predominantly west-east oriented sea.

B: By the early Middle Miocene the connection to the North Sea had vanished, a connection to the Mediterranean was open only through the “Trans-Tethyan-Trench-Corridor” and the connection into the Eastern Paratethys was reduced to a few narrow gateways.

C: During the early Late Miocene the Central Paratethys has changed into Lake Pannon with no marine connections, neither into the Mediterranean nor into the Eastern Paratethys.

D: In the latest Miocene the Messinian Salinity Crisis heavily affected the Mediterranean basins by desiccation and the deposition of thick evaporites (grey hatching). The relict fresh water systems of the Central Paratethys contributed to the hydraulic regime of the Eastern Paratethys.

Maps modified from Rögl (1998) and Popov et al. (2004).



Middle Miocene

Badenian Stage

Papp and Steininger (1978) defined the Badenian Stage based on the stratotype locality Baden-Sooss, south of Vienna (fig. 2). The type-locality is a clay pit in which the characteristic grey-blue basinal clay (local name: “Baden Tegel”) is exposed. The Baden Tegel is well known for its excellent fauna consisting of a highly diverse benthic foraminiferal assemblage and more than 400 of molluscan species besides other invertebrates and vertebrates. The base at the stratotype section was never exposed, the top is unconformably overlain by Sarmatian and Pannonian deposits respectively. The clay pit is not actively worked and the outcrop is now poor.

The base of the Badenian was defined with the first occurrence of *Praeorbulina* (Papp and Cicha 1978) following a transgression above the unconformity due to the Styrian tectonic phase (Stille 1924) and the sea level lowstand at the Bur 4/Lan 1 sequence boundary (Latal and Piller 2003; Strauss et al. 2006). Biostratigraphic subdivision is based on planktonic foraminifers (*Orbulina suturalis*, *Velapertina indigena*, *Globigerinoides quadrilobatus*, *Globorotalita druryi*, *Globorotalia peripheroronda*, *Globoquadrina altispira*) as well as on smaller (*Uvigerina grilli*, *U. macrocarinata*, *U. venusta*, *U. brunensis*, *Pappina parkeri*, *P. neudorfensis*) and larger benthics (*Borelis haueri*, *B. melo melo*, *Planostegina group costata*, *P. giganteiformis*, *Amphistegina mammila*) (cf. Cicha et al. 1998). Several fossil groups increase dramatically in diversity at the onset of the Badenian. This event, the “Early Badenian Build-up Event (EBBE)”, has been explicitly worked out for gastropods, with 505 taxa having their FOs, and for foraminifers, with FOs of 82 taxa (Harzhauser and Piller 2007). These authors denominated this event as “Early Badenian Build-up Event (EBBE)”.

A threefold subdivision of the Badenian is generally carried out based on significant paleoecologic and paleogeographic changes reflected in the composition of the biota (Papp et al. 1978; Kováč et al. 2004). The lower Badenian is represented by the “Lagenidae Zone”, the middle Badenian by the “*Spiroplectammia* Zone”, and the upper Badenian by the “*Bulimina/Bolivina* Zone” (Grill 1943). This subdivision is particularly conspicuous in the eastern Central Paratethys and the Carpathian Foredeep and resulted in the establishment of three substages – Moravian for the lower, Wielician for the middle, and Kosovian for the upper Badenian. Most characteristic are the widespread evaporites of the Wielician Substage (Papp et al. 1978), which occur in the Carpathian Foredeep (Peryt 2001) and in the Transylvanian Basin (Krížsek and Filipescu 2005).

Facies: Besides the highly fossiliferous offshore clays, the Badenian is the climax of the Paratethyan carbonate production. Corallinacean limestones are ubiquitous, but the only noteworthy coral reef phase of the Central Paratethyan succession occurs during the Badenian. Early Badenian reefs in southern parts of the Paratethys are fairly diverse. Especially in the Styrian Basin several small coral reefs composed of *Montastrea*, *Tarbellastraea*, *Leptoseria*, *Acropora*, and *Porites* developed, which, however, had to keep pace with high terrigenous and volcanoclastic input (Friebe 1993; Riegl and Piller 2002; Erhart and Piller 2004). By Late Badenian times a variety of photozoan and heterozoan carbonate facies was still present (Dullo 1983), but a distinct change in coral constructions had occurred. Even in the southern Central Paratethyan

basins (e.g., Vienna Basin, Styrian Basin) complex reefs are observed no more. They were replaced by coral carpets, developed along detached islands and dominated by *Porites*, *Tarbellastraea*, *Caulastrea*, *Acanthastrea*, and *Stylocora* (Piller and Kleemann 1991; Riegl and Piller 2000, 2002). This shift in reef structure and diversity seems to be linked to the climatic deterioration triggered by the global Mid-Miocene Climate Transition (Shevenell et al. 2004). In northern parts of the Paratethys this change is more severe, leading to a loss of algal-bryozoan-coral bioconstructions in favour of algal-serpulid-vermetid “reefs” (Pisera 1996; Studencki 1999).

Correlation: Based on the FOD of *Praeorbulina* in the Styrian Basin, the Vienna Basin and the Alpine Foreland Basin the early Badenian can be correlated with the early Langhian of the Mediterranean (Rögl et al. 2002). In all known shallow water sites of the Central Paratethys the very base of the Middle Miocene, however, is missing. This is clearly related to the widespread and pronounced unconformity and reflected by the missing first evolutionary stages from *Globigerinoides bisphericus* to *Praeorbulina* in nearly all basins and sections (aside from one section in the Styrian Basin; Rögl, pers. comm. 2007). The latter is usually represented by co-occurring *Po. glomerosa curva* and *Po. glomerosa glomerosa* only (Rögl et al. 2002). In terms of nannoplankton stratigraphy the lowermost Badenian still correlates to NN4 due to the occurrence of *Helicosphaera ampli-aperta* and *Sphenolithus heteromorphus* (Rögl et al. 2002; Spezzaferri et al. 2002, 2004). Higher up NN5 is clearly reflected by the presence of *Helicosphaera waltrans* together with *S. heteromorphus* (Rögl et al. 2002). With these biostratigraphic tie points the transgression at the base of the Badenian can clearly be correlated to the global sea level cycle TB 2.3. of Haq et al. (1988) and Bur 5/Lan 1 of Hardenbol et al. (1998) (Kováč et al. 2004; Strauss et al. 2006). The top of this lower Badenian cycle is marked by an unconformity in seismic surveys in the Vienna Basin, pointing to a sea level drop of more than 120 m (Kreutzer 1986; Weissenbäck 1996; Harzhauser and Piller 2007). Furthermore, in many marginal settings, e.g. the Alpine Foreland Basin and the Eisenstadt-Sopron Basin, the end of the marine sedimentation of the first Badenian cycle can be correlated to the same event (Mandic et al. 2002; Mandic 2004; Kroh et al. 2003). Based on the co-occurrence of *Orbulina* and *Praeorbulina* in the underlying deposits (e.g., Rögl et al. 2002), the basin-wide occurrence and the remarkable magnitude of the sea level drop a link with the global sea level drop at about 14.2 Ma is reasonable. This event was triggered by the expansion of the East Antarctic ice sheet (Flower and Kennett 1993; Shevenell et al. 2004) and corresponds to the Lan 2/Ser 1 sequence boundary of Hardenbol et al. (1998).

The second Badenian cycle is interpreted to be an expression of the global sea level cycle TB 2.4. of Haq et al. (1988). A distinct lowstand wedge and a well-developed transgressive wedge are observed in seismic studies in the Vienna Basin (Kreutzer 1986; Strauss et al. 2006). In the Carpathian Foreland basins and in the Transylvanian Basin a pronounced evaporitic phase starts, known as the Wielician crisis (Steininger et al. 1978; Kasprzyk 1999; Chira 2000), which correlates to the Lan 2/Ser 1 lowstand of Hardenbol et al. (1998). While evaporite formation continued in the east throughout the middle Badenian, in the western parts of the Central Paratethys this cycle is characterised by corallinacean platforms with frequent caliche formation and vadose leaching (Dullo 1983; Schmid et al. 2001). The occurrence of *Sphenolithus heteromorphus* places this middle Badenian sediments in nannoplankton zone NN 5.

The renewed flooding of the third Badenian cycle is biostratigraphically dated by the onset of nannoplankton zone NN6 (Hudáckova et al. 2000; Kovác et al. 2004). The base of this biozone is defined by the last occurrence of *Sphenolithus heteromorphus* and corresponds to the Langhian/Serravallian boundary in the Mediterranean, which was calibrated by Foresi et al. (2002a) at 13.59 Ma (see also Gradstein and Ogg 2004; Gradstein et al. 2004; Lourens et al. 2004). Therefore, the Langhian/Serravallian boundary is located within the Badenian and correlates roughly to the middle/upper Badenian boundary. Considering the dating and the magnitude of this cycle a correlation with the global cycle TB 2.5. of Haq et al. (1988) can be expected. This late Badenian is characterised by a stratified water body indicated by the deposition of dysoxic pelites in basinal settings in the entire Central Paratethys area (e.g., Hudáckova et al. 2000). Even the platforms became affected by repeated hypoxic events as documented by Schmid et al. (2001).

Paleogeography: The paleogeographic situation changed strongly during the Badenian. During the early Badenian the “Trans-Tethyan Trench Corridor” via Slovenia was still open and connected the Mediterranean Sea with the Pannonian basin system (fig. 3B). The connections into eastern directions, however, are still controversial. While Rögl (1998a) and Steininger and Wessely (2000) postulate an open connection into the Eastern Paratethys (fig. 3), Studencka et al. (1998) and Popov et al. (2004) indicate a land barrier between both seas. Rögl (1998a) discussed an already subducted marine pathway between the southern margin of the Black Sea plate and the Pontids, connecting the Eastern Mediterranean with the Central Paratethys. A repeated re-opening of the Tethyan gateway between the Mediterranean and the Indo-Pacific during the Langhian (early Badenian) (Rögl 1998a; Popov et al. 2004) and even into the Serravallian (Jones 1999) is highly probable.

During the middle Badenian the eastern seaways were sealed. Water supply for the Central Paratethys was only warranted via the “Trans-Tethyan Trench Corridor”. This gateway was finally closed in the late Badenian. The entire Central Paratethys was therefore depending on a connection with the Eastern Paratethys via today’s western Black Sea area. Faunistic differences between the diverse Central Paratethys and the impoverished Eastern Paratethys at that time, however, exclude the Eastern Paratethys as passage into the west (Studencka et al. 1998). Again, the enigmatic seaway between the Black Sea plate and the Pontids might have acted as gateway (Rögl 1998a). Such a connection would also be obligatory to explain the immigration of new radiolarian assemblages into the Central Paratethys as discussed by Dumitrica et al. (1975).

The next dramatic change in marine biota occurs with the dawn of the Sarmatian. Of the Badenian fauna, 588 LOs of gastropods and 121 of foraminifers are recorded and designate this event as the strongest turnover event of the Paratethyan history. Harzhauser and Piller (2007) christen this event the “Badenian-Sarmatian-Extinction-Event” (BSEE). The faunal re-orientation was triggered by a strong restriction of the open ocean connections of the Central Paratethys (Rögl 1998a), corresponding to the Ser 3 sequence boundary of Hardenbol et al. (1998) and the begin of cycle TB 2.6. of Haq et al. (1988) (Kovác et al. 1999, 2004; Harzhauser and Piller 2004b; Strauss et al. 2006). The Badenian/Sarmatian boundary would thus be related with the glacio-eustatic isotope event MSi-3 at 12.7 Ma (Abreu and Haddad 1998). Correspondingly, a considerable hi-

atus at the Badenian/Sarmatian boundary is indicated by a strongly erosive discordance in seismic lines in Paratethyan basins (Harzhauser and Piller 2004a, b).

Sarmatian Stage

The Sarmatian as a regional stage was already defined in the Vienna Basin by Suess (1866). Its stratotype was designated in the northern Vienna Basin at the Nexing section (fig. 2) (Papp and Steininger 1974) which is characterized by biogenic sediments of molluscan shells. The type section is part of the upper Sarmatian and does not represent a boundary stratotype. Also at its top no boundary to the Pannonian is preserved. Lithology and depositional environment at the type-locality are very specific and not representative for the stage (Harzhauser and Piller accepted). At large, the Sarmatian is a strongly twofold stage. The lower Sarmatian, above a pronounced and widespread unconformity, is dominated by fine siliciclastic sediments. The highly variable carbonate facies, characteristic for the Badenian, vanished completely within the entire Paratethys Sea at the Badenian/Sarmatian boundary. The upper Sarmatian sediments reflect a mixed carbonate-siliciclastic regime all over the Central Paratethys (Harzhauser and Piller 2004a, b).

The base of the Sarmatian was defined by the occurrence of a highly endemic fauna, particularly molluscs and to a lesser extent foraminifers. Both groups allow the establishment of an ecostratigraphic subdivision, which comprises for the lower Sarmatian the *Mohrensternia* Zone and lower *Ervilia* Zone among molluscs and the *Anomalinoidea dividens* Zone, *Elphidium reginum* Zone and *Elphidium hauerinum* Zone among benthic foraminifers. The upper Sarmatian contains the *Porosonion granosum* Zone and is subdivided into the upper *Ervilia* Zone and *Sarmatimactra vitaliana* Zone by molluscs. Contemporaneous with the abrupt increase in endemics, a total loss in stenohaline biota occurs at the Badenian/Sarmatian boundary. Since radiolarians, planktonic foraminifers, corals and echinoderms are completely absent these sediments were also named “brackish stage” (Suess 1866). Although this denomination was rejected later (see Papp 1974a) the idea of an environment with reduced salinity for the Sarmatian in general was favoured until recently (e.g., Kovác et al. 1999). After a well-based opposition to this interpretation by Pisera (1996), Piller and Harzhauser (2005) presented a range of data pointing clearly to normal marine conditions for most of the Sarmatian environments. Geophysical correlation, based on many oil-exploration boreholes, works well and consistently with ecostratigraphy in the Central Paratethys basins (Harzhauser and Piller 2004b).

Facies: The lower Sarmatian is characterized by siliciclastic sediments, often with conglomerates at the base overlain by fine-clastics, the latter frequently deposited on tidal flats or in estuaries and rich in low-diversity molluscan faunas (Harzhauser and Piller 2004b). Diatomites with marine diatoms and silicoflagellates are a more open-water facies (Rögl and Müller 1976; Harzhauser and Piller 2004a, b; Schütz et al. 2007). Off-shore deposits are represented by marls and silty clays with an impoverished bivalve fauna (Kojumdgieva et al. 1989). Carbonate rocks are represented only rarely by autochthonous build-ups formed by the polychaete *Hydroidea* and by bryozoans (Harzhauser and Piller 2004a, b; Piller and Harzhauser 2005). These bioconstructions are best developed in the Carpathian Foredeep, extending as a chain of patches from Poland via Moldavia and Rumania to Bulgaria (Pisera 1996). The lower Sarmatian is terminated by the basin-wide occurrence of con-

glomerates (e.g., Vienna and Styrian basins; Harzhauser and Piller 2004b)

At the onset of the late Sarmatian, sedimentation switched from a siliciclastic to a carbonate dominated system throughout the Central Paratethys. Oolites and coquina-dominated sands start to spread in nearshore settings and on shallow shoals giving rise to small carbonate platforms (Harzhauser and Piller 2004a, b, accepted). Mass occurrences of the larger foraminiferal species *Spirolina austriaca* d'Orbigny characterize the latest Sarmatian. Coevally a drastic increase occurred in bivalve shell thickness (e.g., *Venerupis*, *Sarmatimacra*; Papp et al. 1974; Piller and Harzhauser 2005). The early Sarmatian polychaete-bryozoan communities collapsed and were replaced by unique foraminiferan build-ups characterized by the sessile nubeculariid genus *Sinzowella* in association with calcareous algae and microbial carbonate. The Sarmatian oolites are the only Miocene oolites in the entire Central Paratethys area.

Correlation: Correlating outside the Central Paratethys is problematical, due to the restricted connection of Paratethys to the Mediterranean and the lack of most stenohaline faunas. Planktonic foraminifers are almost entirely absent (Cicha et al. 1998; Harzhauser and Piller 2007). The only saviour is the calcareous nannoplankton, low in diversity and with endemic taxa (Steininger et al. 1976; Stradner and Fuchs 1979). The absence of *Sphenolithus heteromorphus* indicates a correlation with zone NN6 (e.g., Schütz et al. 2007), and the occurrence of *Discoaster kugleri* in the uppermost part indicates NN7.

The very pronounced sea level lowstand at the Badenian/Sarmatian boundary can be correlated with the Ser 3 sequence boundary of Hardenbol et al. (1998). The lowstand at the end of the Sarmatian, representing the Sarmatian/Pannonian boundary, accordingly can be correlated with the Ser 4/Tor 1 sequence boundary which coincides with the Serravallian/Tortonian boundary (Lourens et al. 2004). The entire Sarmatian corresponds to only one 3rd order sea level cycle – TB 2.6. of Haq et al. (1988) (Harzhauser and Piller 2004b; Kováč et al. 2004). Biostratigraphic data combined with astronomically derived ages place the Serravallian/Tortonian boundary at 11.54 Ma (Lirer et al. 2002; Foresi et al. 2002b) what is in accordance with the age (11.5 Ma) proposed by Rögl et al. (1993) and Kováč et al. (1998a, b) for the Sarmatian/Pannonian boundary. The sea level lowstand between the lower and upper Sarmatian can be interpreted as lowstand between two 4th order cycles (Kosi et al. 2003; Harzhauser and Piller 2004b; Strauss et al. 2006). In the Eastern Paratethys the Sarmatian has an analogue in the regional stages Volhynian and (lower) Bessarabian (Rögl 1998a, b; Harzhauser and Piller 2004b, 2007). Although without any tie point, the sediments of the Sarmatian show a clear astronomical signal with a 400 ka eccentricity component which may have triggered the 4th order cycles and, in addition, 100 ky and 2.35 Ma components (Harzhauser and Piller 2004b).

Paleogeography: During the Sarmatian the Paratethys became almost completely sealed off from the Mediterranean. The Central Paratethys was, however, well connected to the Eastern Paratethys (Rögl 1998a). From there, a narrow marine connection into the Mediterranean Sea formed far in the east due to tectonic movements along the S-Anatolian fault system (Chepalyga 1995; Steininger and Wessely 2000). The fair connection between the two Paratethyan seas is reflected by a striking similar faunistic inventory characterised by a highly

endemic and considerably impoverished fauna lacking most stenohaline taxa (Kolesnikov 1935; Papp et al. 1974). This peculiar character of the marine fauna was recognised already by Suess (1866) who then introduced the term Sarmatian.

Late Miocene

Pannonian Stage

The stratotype of the Pannonian Stage is located in a clay pit in Vösendorf (Lower Austria) close to the southern border of the city of Vienna (Papp 1985). The type section contains highly fossiliferous clays with sandy interlayers; its base is not exposed. All surface outcrops show a discontinuous sedimentation between the Sarmatian and Pannonian, although several authors refer to “transitional beds” (e.g., Janoschek 1942; Papp 1951). This interpretation was evoked by reworked Sarmatian fossils at the base of the Pannonian deposits (Harzhauser et al. 2004).

The turn from the Sarmatian to the Pannonian is marked by a major incision in faunal content with an extinction rate over 90% for gastropods and foraminifers. This is the “Sarmatian-Pannonian-Extinction-Event” (SPEE) (Harzhauser and Piller 2007). The Pannonian Stage was established on its very peculiar mollusc fauna with a high degree of endemism and rapid evolutionary radiations (Müller et al. 1999) reflecting the evolution of a long living lake system, called Lake Pannon (fig. 3C). Among bivalves the genera *Mytilopsis*, *Congerina*, and *Lymnocardium* and among gastropods the genus *Melanopsis* are the most important representatives mirroring this evolutionary history. The development of the fauna was controlled by the gradual freshening of the water body as well as by geodynamic processes, resulting in profound changes in lake geometry (Magyar et al. 1999b).

The evolutionary lineages of molluscs allow a clear biostratigraphic subdivision within the lake sediments, as already realized by Fuchs (1875) and elaborated in a great detail by Papp (1951) who applied a letter zonation (Pannonian A-H) instead of eco-biozones. The type section is stratigraphically located in Zone E. After a first attempt by Rögl and Daxner-Höck (1996), the letter zonation has been traced back to biozones by Magyar et al. (1999a, b) and Harzhauser et al. (2004). This molluscan biozonation can be differentiated for littoral and sublittoral depositional environments. A biozonation based on dinoflagellates has been established (Magyar et al. 1999a).

Facies: Deltaic gravels, sands, whitish marls and lignites accumulated along the coasts of Lake Pannon. Typical deposits in basal settings are grey-blue clays and marls as exposed in the stratotype section (Papp 1985). During phases of high water tables the deep lake areas have been exposed to hostile dysoxic conditions resulting from a well developed hypolimnion (Harzhauser and Mandić 2004). Carbonate sediments are completely lacking in the Central Paratethys whilst oolites and bryozoan bioconstructions are still frequent in upper Bessarabian deposits of the Eastern Paratethys (Pisera 1996). During the late Pannonian, the northwestern part of the lake – e.g. Vienna Basin – turned into floodplain-environments as the coastline retreated (Magyar et al. 1999b; Harzhauser and Tempfer 2004). The central and southern part remained as a subbasin complex filled by prodelta turbidites and prograding deltaic deposits, several hundred meters deep (Popov et al. 2004). Despite its shrinking size, the southern coastline along the northern Dinarids was quite stable throughout the Pannonian (Magyar et al. 1999b; Popov et al. 2004) (fig. 3D).

Correlation: For Harzhauser et al. (2004) the lower to middle Pannonian lake deposits represent the single cycle TB 3.1. of Haq et al. (1988), starting at the Middle Miocene/Upper Miocene (=Serravallian/Tortonian) boundary due to the influence of the glacio-eustatic sea level lowstand Ser 4/Tor 1 of Hardenbol et al. (1998) (Strauss et al. 2006). The upper Pannonian sediments belong to the two 3rd order cycles TB 3.2. and 3.3. (Haq et al.1988) starting with the Tor 2 lowstand of Hardenbol et al. (1998). As in the Sarmatian, a clear astronomical signal with a 100 ka, a 400 ka, and a 2.35 Ma component has been detected (Harzhauser et al. 2004). Although currently floating, this signal may enable a more precise correlation of the Pannonian with the global chronostratigraphic scale.

Stevanovic et al. (1990), misled by similarities of the endemic mollusc faunas, erroneously correlated upper Pannonian deposits of Lake Pannon with deposits of the Eastern Paratethys Pontian Stage. This stratigraphic concept became formalised with the publication in an independent volume of the series *Chronostratigraphie und Neostatotypen*. The result is an erroneous usage of the “Pontian Stage” for Lake Pannon deposits, being now deeply rooted in the literature until recently (e.g., Saftic et al. 2003). A very detailed integrative stratigraphic evaluation of magnetostratigraphic, geochronologic and biostratigraphic data clearly demonstrated that the upper Pannonian deposits, starting with the regional *Congeria praerhomboides* Zone, precede the Pontian of the Eastern Paratethys by at least 2 Ma (Magyar et al. 1999a). Even this very clear data did not result in the definite suppression of the name “Pontian” from Lake Pannon deposits. The introduction of a new regional stage (“Transdanubian”) representing the interval between the base of the upper Pannonian and the Eastern Paratethys stage Pontian (Sacchi and Horvath 2002) creates more problems than it solves. Thus, the base of the Pontian Stage, usually correlated with the base of the Messinian and with a 3rd order sequence-stratigraphic surface in the topmost part of the Lake Pannon infill (= TB 3.3. of Haq et al. 1988) was currently shown to be at least 1 Ma younger than the base of the Messinian (Popov et al. 2004; Vasiliev et al. 2005). At that time, however, Lake Pannon probably has become already completely desiccated (fig. 3D). The Pontian is here dismissed from the regional chronostratigraphic scheme of the Pannonian Basin System.

Paleogeography: Lake Pannon was an enclosed basin of highly variable extent (Magyar et al. 1999b) covering the Pannonian Basin system which was framed by the Alps, the Carpathians and the Dinarids (fig. 3C). The development of the lake illustrates the ongoing continentalisation in central and south-eastern Europe and progressive restriction of the aquatic realm in the Central Paratethys area. Lake Pannon formed at about 11.6 Ma in place of the relic Central Paratethys Sea. At that time, the Eastern Paratethys reached westward into the Dacian Basin. Its associated Bessarabian fauna is a direct descendent of the late Middle Miocene Sarmatian/Volhynian faunas (Kolesnikov 1935) and differs fundamentally from the Lake Pannon assemblages.

DISCUSSION AND CONCLUSIONS

The high degree of endemism existing in the Paratethys from time to time, caused by strong isolation from other oceanic realms (e.g., Mediterranean, Indo-Pacific, Atlantic), together with inadequate definitions of the Mediterranean stages until the second half of the 20th century, induced the establishment of a regional chronostratigraphic/geochronologic classification for both the Western/Central and the Eastern Paratethys. The defini-

tion of these regional stages, however, is based solely on fossil contents. These biota – and even some stages – roughly represent Assemblage and Abundance zones (Acme zones) in terms of biozone definitions, supported by a few good marker taxa in some of the stages. For none of the Miocene Central Paratethyan stages is a boundary stratotype defined.

Correlation with Mediterranean/global chronostratigraphy is based on scattered biostratigraphic tie points, particularly those of calcareous nannoplankton and planktonic foraminifers. The occurrence of nannoplankton taxa correlates much better with the zonation of Martini (1971) than with other zonal schemes. Among calcareous nannoplankton these are the FO of *Helicosphaera ampliaperta* in the Eggenburgian, the LO of *Sphenolithus belemnos* in the Ottnangian, the LO of *H. ampliaperta* at the end of the lower Badenian, the LO of *Sphenolithus heteromorphus* at the end of the middle Badenian, the total range of *H. waltrans* in the early Badenian, the FO of *Discoaster kugleri* in the Sarmatian, and the FO of *D. hamatus* in the Pannonian. Planktonic foraminiferal markers are represented by *Catapsydrax* in the Ottnangian, by *Globigerinoides bisphericus* in the upper Karpatian, and by the *Praeorbulina* lineage in the Badenian. With increasing isolation in the course of the Miocene such tie points become scarcer.

Due to the poor outcrop situation and lack of long sections magnetostratigraphic correlation is only very limited. Some surface data for the Early, Middle and Late Miocene have been summarized by Daxner-Höck et al. (1998), Magyar et al. (1999a), Scholger and Stingl (2004), and Harzhauser et al. (2004), which concentrate mostly on mammal-bearing sequences. All these data are very punctiform, comprising usually only one or two chrons. Therefore, their interpretation is largely dependant from the *a priori* age model.

All stages are bounded by sea level lowstands which coincide with 3rd order sea level cycles and can be correlated with the sea level curve of Haq et al. (1988) and sequence stratigraphic cycles of Hardenbol et al. (1998). The Eggenburgian, Badenian and Pannonian Stages span three 3rd order cycles, the Ottnangian, Karpatian, and Sarmatian correlate to only one cycle each.

Taking all available data into account, the Karpatian/Badenian boundary is clearly correlated with the Burdigalian/Langhian (Early/Middle Miocene) boundary and the Sarmatian/Pannonian boundary with the Serravallian/Tortonian (Middle/Late Miocene) boundary. The base of the Neogene (Chattian/Aquitainian boundary = Oligocene/Miocene boundary) falls within the Egerian and the Aquitanian/Burdigalian boundary within the Eggenburgian. Both boundaries can not be identified with more precision. The Ottnangian and Karpatian Stages correlate to the upper Burdigalian and within there to 3rd order cycles TB 2.1. and TB 2.2. of Haq et al. (1988). The Langhian/Serravallian boundary can be correlated with the middle/upper Badenian boundary based on the LO of *Sphenolithus heteromorphus*, the Sarmatian can be correlated with cycle TB 2.6., bounded by Ser 3 lowstand at its base and Ser 4/Tor 1 (Hardenbol et al. 1998) at its top. The lower Pannonian coincides with cycle TB 3.1., the upper Pannonian with cycles TB 3.2. and TB 3.3. The Pontian Stage belongs to a different geodynamic terrain and has to be excluded from the Pannonian Basin System.

This correlation clearly shows that, regional geodynamic processes notwithstanding, the global sea level signal is still visible in these isolated basins. In concert with regional parameters this

global signal is responsible for the general sedimentary and also biotic development of the Central Paratethys. For the middle and upper part of the Miocene Central Paratethyan successions also a distinct astronomical signal is evident. At the moment, this signal cannot be pin-pointed into the ATNTS 2004 (Lourens et al. 2004).

Since the Central Paratethyan regional stages follow a clear global signal (sea level changes, astronomical forcing), and their definition in terms of chronostratigraphic rules is very poor or even missing, and their usage is merely biotically (or biostratigraphically) founded, the necessity of this regional chronostratigraphic subdivision has to be seriously questioned. The answer to this question is, however, linked to the general definition of stages. The current definition of a stage includes its global spread (see discussion in Aubry et al. 1999). In this case formally defined regional stages are redundant and not necessary for the Central Paratethys. However, if stages are essentially regional, then a regional scale as for Central Paratethys would be much more appropriate.

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