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Paleoenvironmental changes across the Eocene-Oligocene boundary: insights from the Central-Carpathian Paleogene Basin

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Abstract: The sedimentary sequence of the Central-Carpathian Paleogene Basin provides proxy records of climatic changes related to cooling events at the Eocene/Oligocene boundary (TEE). In this basin, climatic deterioration is inferred from the demise of the carbonate platform and oligotrophic benthic biota in the SBZ19 and from the last species of warm-water planktonic foraminifers in the E14 Zone. Upper Eocene formations already indicate warm-temperate to cool-temperate productivity and nutrient-enriched conditions (Bryozoan Marls, Globigerina Marls). Rapid cooling during the earliest Oligocene (Oi-1 event) led to a temperature drop (~11 °C), humidity, fresh water influx and continental runoff, water mass stratification, bottom water anoxia, eutrofication, estuarine circulation and upwelling, carbonate depletion, sapropelitic and biosiliceous deposition, H_2S intoxication and mass faunal mortality, and also other characteristics of Black Sea-type basins. Tectonoeustatic events with the interference of TA 4.4 sea-level fall and the Pyrenean phase caused basin isolation at the beginning of the Paratethys. The Early Oligocene stage of Paratethyan isolation is indicated by a stagnant regime, low tide influence, endemic fauna development, widespread anoxia and precipitation of manganese deposits. The episodic rise in the sea-level, less humid conditions and renewed circulation is marked by calcareous productivity, nannoplankton blooms and the appearance of planktic pteropods and re-oxygenation. Paleogeographic differentiation of the Carpatho-Pannonian Paleogene basins resulted from plate-tectonic reorganization during the Alpine orogenesis.

Key words: Paratethys, Central Western Carpathians, Terminal Eocene Event, platform drowning, climatic cooling, productivity changes, estuarine circulation, anoxia, eutrophication, semi-isolation.

Introduction

The Late Eocene was a transitional period between the Middle Eocene Climatic Optimum and the Oligocene icehouse (Fig. 1). By the Oligocene, the climate system tended to an "icehouse" world. Gradual cooling of the Earth's climate resulted in the expansion of Antarctic glaciation. Climatic deterioration already began from the Middle/Late Eocene boundary (Oberhänsli 1996), and was followed by the major cooling event in the Early Oligocene (Biolzi 1985; Kennet & Barker 1990; Diester-Haass 1991; Miller et al. 1991; Zachos et al. 1993, 1996; Prothero 1994; Diester-Haass & Zahn 1996, 2001; Salamy & Zachos 1999; Wade & Pälike 2004; Tripati et al. 2005, etc.). This climatic cooling led to significant paleoenvironmental changes in the Carpathian Paleogene basins (Fig. 2). From the Early Oligocene, the Carpathian basins provided the first records of the isolation from the open sea (Early Paratethys — Báldi 1980, 1984, 1986; Protoparatethys — Russu 1988), which enhanced gradually during the Late Oligocene and Miocene (Eoparatethys–Mesoparatethys–Neoparatethys, see Seneš & Marinescu 1974; Nagymarosy 1990; Popov et al. 1993; Rögl 1998, 1999; Kováč 2000; Steininger & Wessely 2000, etc.). Therefore, the Carpathian Paleogene basins mirrored the paleoenvironmental changes of climatic cooling and Paratethyan isolation quite sensitively. Their sedimentary records, like the Globigerina Marls, Menilite Beds, Tylawa Marls, Dynow Marls etc., occurred near the Eocene-Oligocene boundary in practically all the Carpathian Paleogene basins (Fig. 2). Previous studies provided an important insight into paleoenvironmental changes in the Outer Carpathian basins (Van Couvering et al. 1981; Krhovský 1981a,b, 1995; Hanzlíková 1981; Roth & Hanzlíková 1982;

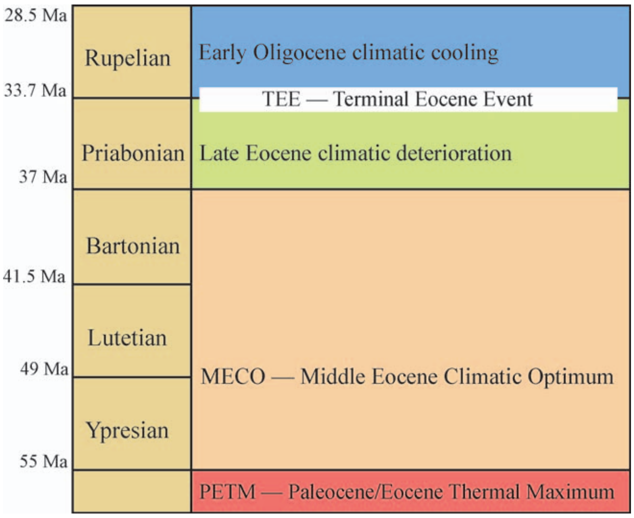

Fig. 1. Climatostratigraphic development of the Paleogene Period from the Paleocene/Eocene Thermal Maximum (PETM), following by the Middle Eocene Climatic Optimum (MECO), climatic deterioration during the Late Eocene and culminating in the climatic cooling of the Terminal Eocene Event (TEE) and during the Early Oligocene.

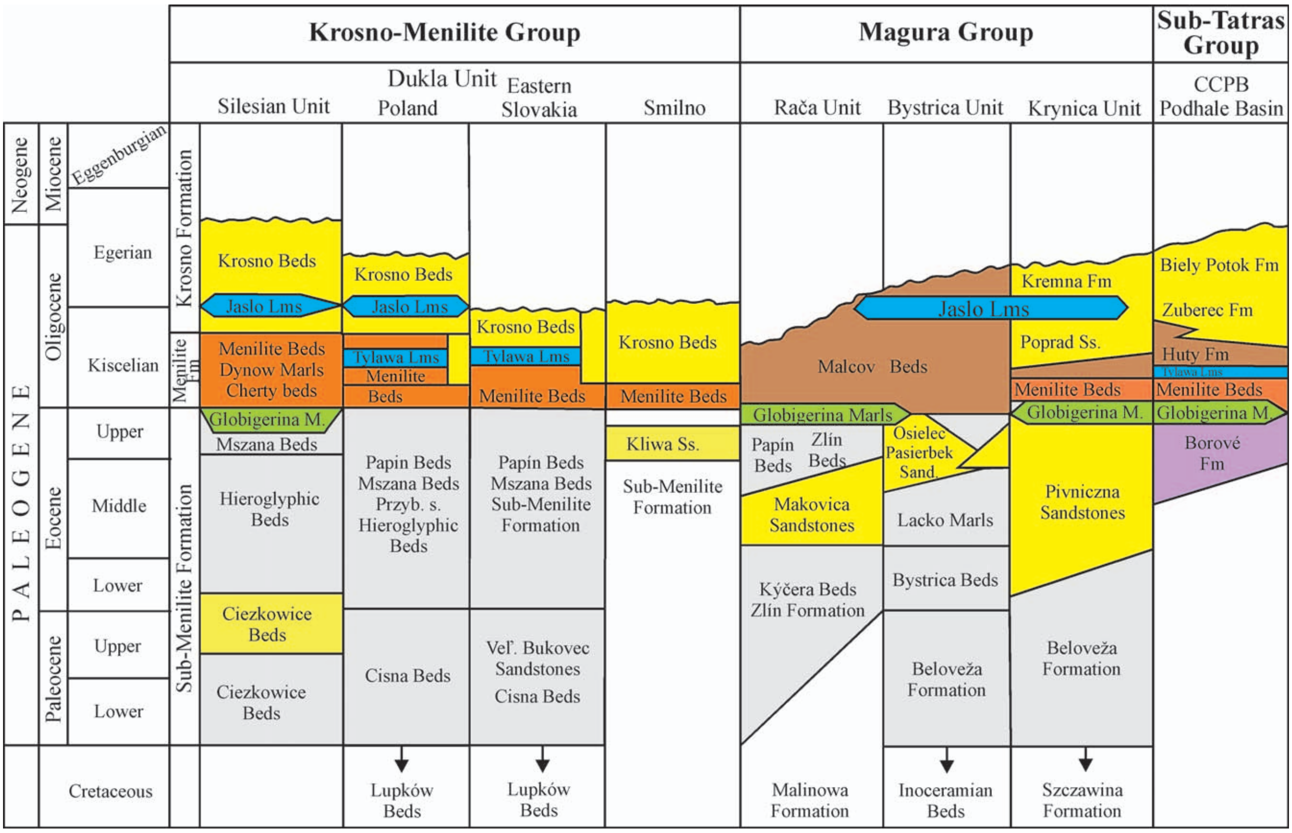

Fig. 2. Lithostratigraphic scheme of the Paleogene formations of the Central and Outer Western Carpathians. The Eocene/Oligocene boundary is marked by the Globigerina Marls and Menilite Beds, which form a marker horizon dividing the lower deep-water turbidite complex (Paleocene–Upper Eocene) from the Supra-Menilite formations (Upper Kiscellian–Lower Miocene). These marker horizons of high-rate productivity and subsequent sapropelitic deposition occur in almost all the Carpathian basins.

Krhovský & Djurasinovič 1992; Krhovský et al. 1993; Leszczyński 1996, 1997; Oszczypko 1996; Oszczypko-Clowes 1998; Gedl 1999; Gedl & Leszczyński 2005; Puglisi et al. 2006; Švábenická et al. 2007; Miclaus et al. 2009, etc.). The purpose of this study is to review the Eocene/Oligocene events, providing their paleoenvironmental and climatic proxies in the Central-Carpathian Paleogene Basin (CCPB — Fig. 3).

Eocene/Oligocene transition in the CCPB

The CCPB accommodates a forearc basinal system of the West Carpathian Mountain chain (Soták et al. 2001; Kázmer et al. 2003). The general litholo-

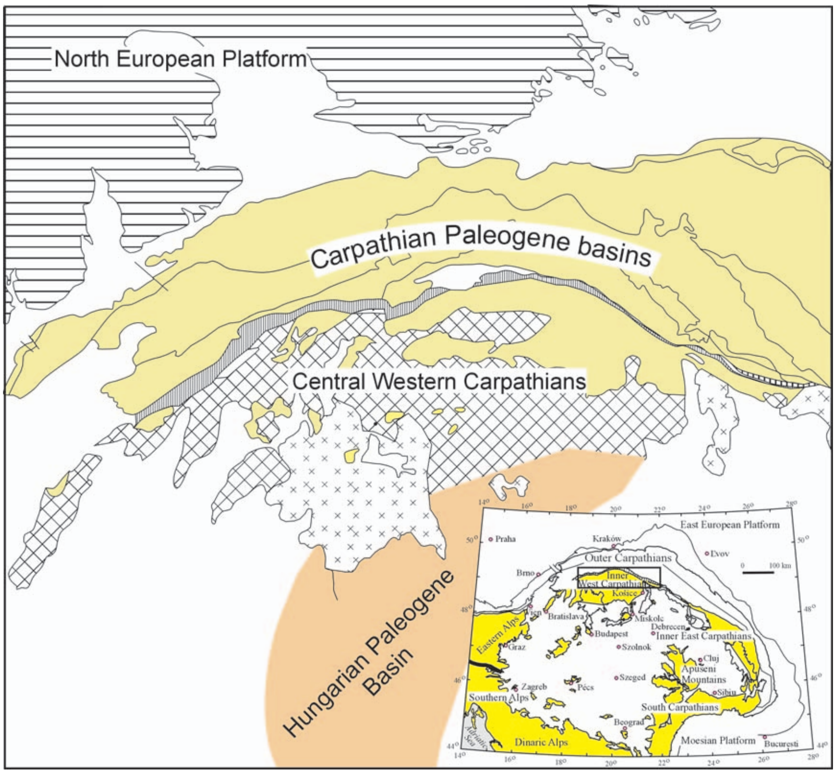

Fig. 3. Paleogene basins of the Carpatho-Pannonian area. The Central-Carpathian Paleogene Basin is accomodated in the Central Western Carpathians.

gy of the CCPB around the Eocene/Oligocene transition is compiled in type-section (Fig. 4). The sedimentary sequence is developed from the Lutetian formations, continued across the Eocene/Oligocene boundary and terminated to the Late Oligocene/Miocene? formations. The initial phase of deposition is represented by conglomerates, boulder breccias and poorly sorted sandstones, representing the sediments of rock-fall avalanches, alluvial fans and delta-fed fans. These sediments are overlain by the Lutetian transgressive formations, composed mostly of sandstones and nummulitic limestones. Their stratigraphic age ranges within the SBZ16–19 (Köhler 1998; Bartholdy et al. 1999). Apart from the carbonate platform facies, Middle Eocene pelagic sediments are also preserved. These contain late Ypresian fauna with a dominance of subbotinid species, early Lutetian fauna dominated by acarininid species, mid-Lutetian fauna with a dominance of morozovellid species and late Lutetian to Bartonian fauna with morozovelloid and truncorotaloid species. Biostratigraphic determination of the Middle Eocene formations is based on the foraminiferal index species, such as *Turborotalia frontosa* (late Ypresian — Biozone E7/P9), *Acarinina cuneicamerata*, *A. praetopilensis*, *Morozovella aragonensis*, *M. gorondatxensis* (early to middle Lutetian — Biozone E8–10/P10–P12), *Acarinina* (*T.*) *topilensis*, *Morozovella spinulosa* (late Lutetian — Biozone E10–11/P12) and *Morozoveloides crassata*, *Truncorotaloides rohri* (Bartonian — Biozone E12–13/P13–P14, biostratigraphic classification *sensu* Soták 2007).

The sedimentary sequence of the CCPB graded up to the Priabonian–Rupelian formations (Fig. 5), belonging to the E14–O3 Biozones (Soták et al. 2007). The sequence started with the Globigerina Marls, the basal part of which belongs to the E14/P15 Biozone, marked by its nominate taxon *Porticulasphaera semminvoluta* and associated species, like *Globigerinatheka index*, *Globigerinatheka* aff. *subconglobata*, *Subbotina linaperta* and *S. corpulenta*. The zonal boundary P15/P16 is shown by the LAD of *Porticulasphaera semminvoluta*, followed by the significant increase in the abundance of globigerinathekids. The mid-Priabonian microfauna is dominated by large-sized tests of *Globigerinatheka index*, and their *acme* clearly denotes the E15/P16 Biozone (after Gonzalvo & Molina 1992). Associations in this zone are very rich in large subbotinids, comprising species of *S. corpulenta*, *S. cryptomphala*, *S. gortani*, *S. pera* and *S. praeturrilina*. Stratigraphically important species of *Turborotalia cerroazulensis* lineage are represented by *T. pomeroli* and *T. cerroazulensis*. The distribution of these species indicates the subdivision between the upper part of the E15 Biozone (= the lower part of the P16) and the lower part of the E16 Biozone (= the upper part of the P16). The succesive formation contains more planoconvex species of turborotaliids, resembling *T. cocoaensis*. The presence of this species pro-

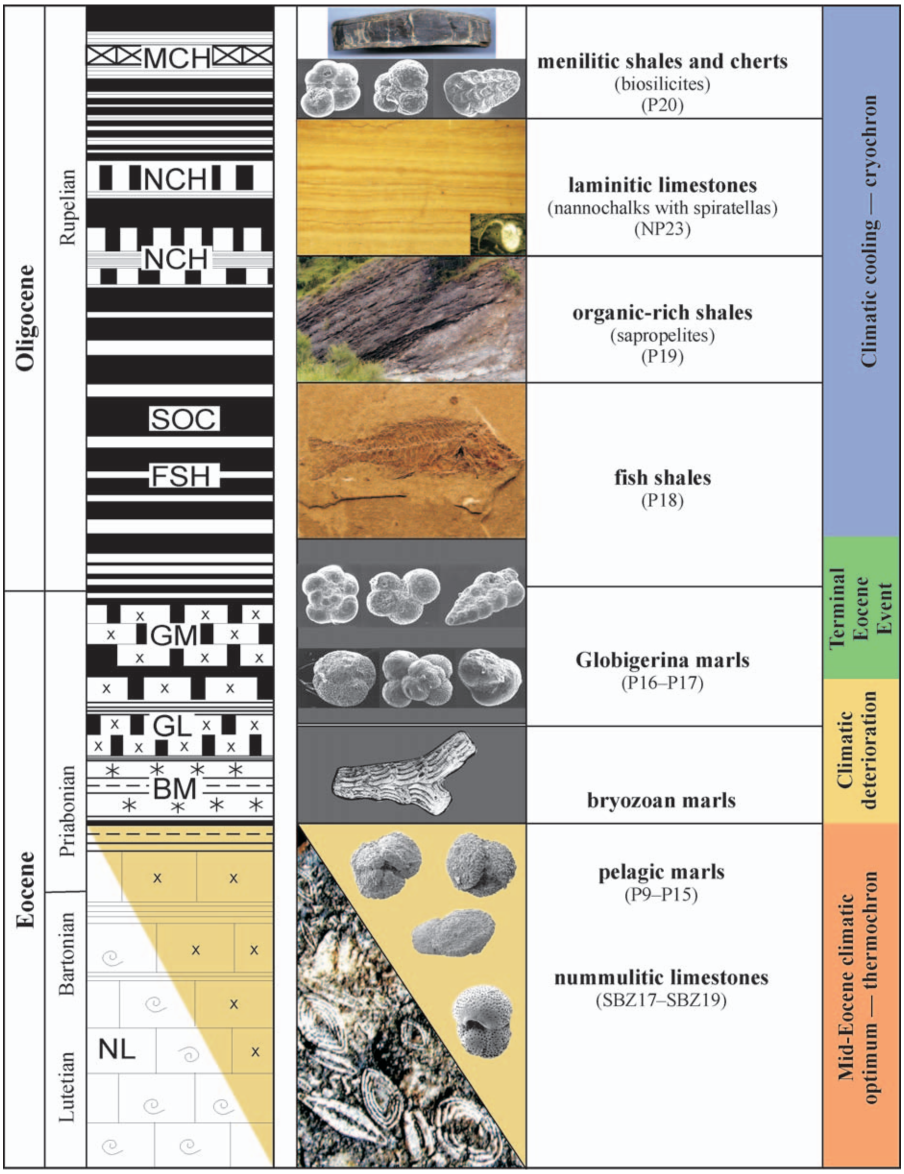

Fig. 4. Composite section of the transitional formations through the Eocene/Oligocene boundary in the Central-Carpathian Paleogene Basin. Successive changes in lithology and life environments reflect the changes in climatic conditions during the deposition. ***Abbreviations*:** **NL** — nummulitic limestones, **BM** — bryozoan marls, **GL** — Globigerina limestones, **GM** — Globigerina marls, **FSH** — fish shales, **SOC** — sapropelitic organic-rich claystones, **NCH** — nannochalks, **MCH** — menilitic cherts.

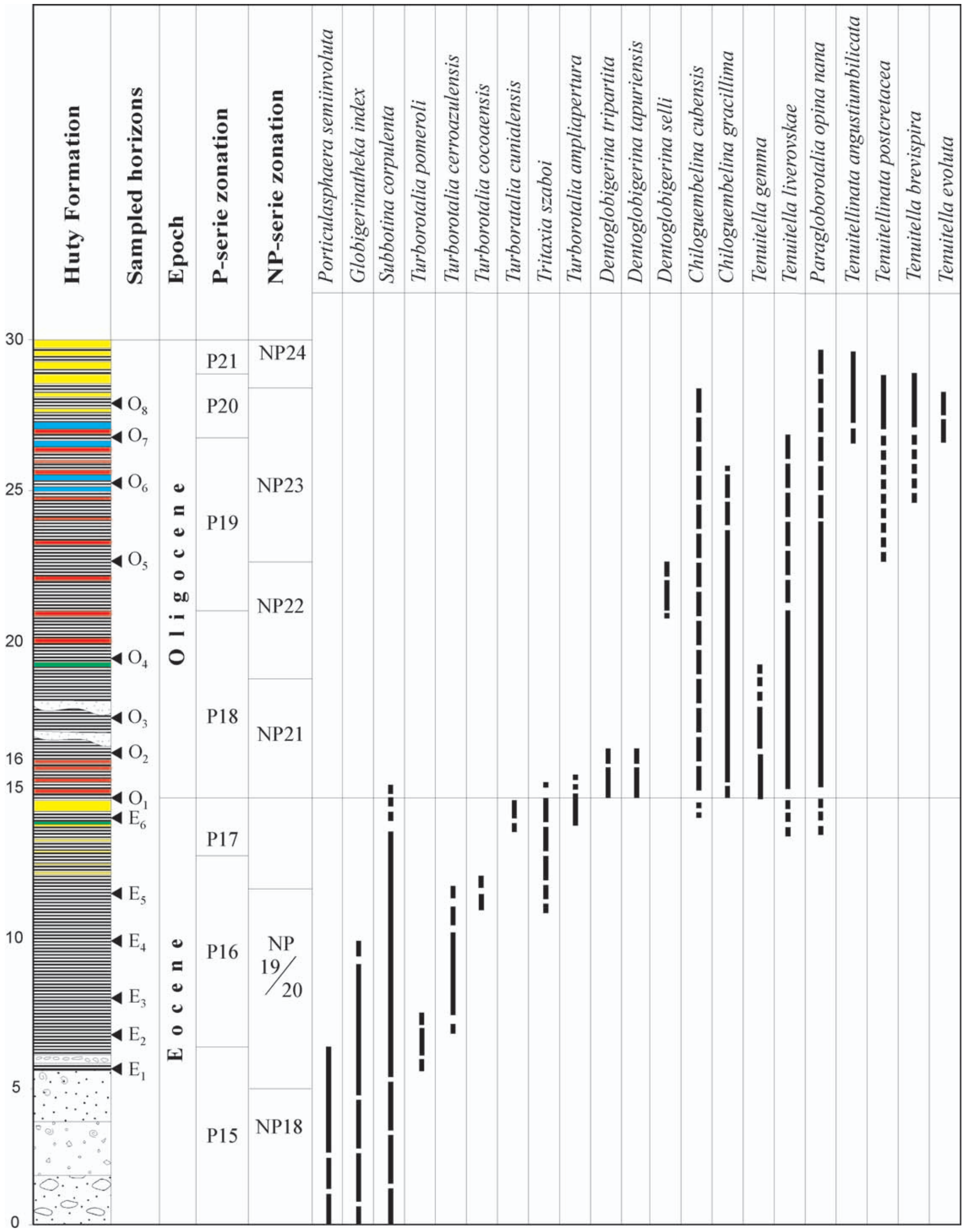

Fig. 5. Biostratigraphic determination of the Eocene/Oligocene boundary in the Central-Carpathian Paleogene Basin showing the vertical distribution of the foraminiferal index species (data adapted from Soták et al. 2007).

vides evidence of the uppermost Upper Eocene biozone (E16/P17). These intervals are considerably richer in benthic foraminifers similar to those recognized in the Buda Marls (microfauna with *Tritaxia szaboi* or *Cylindroclavulina rudilosta* — e.g. Hantken 1875; Sztrákos 1987). Close to the E/O boundary, the *Turborotalia cerroazulensis* lineage disappeared and its last species *T. cunialensis* occurred together with the first appearance of the new species *T. ampliapertura* (FAD 33.8 Ma). Based on the dinocysts, the Late Eocene age of the lower part of the marly sequence is indicated by the FAD of *Reticulatosphaera actinocoronata* and by the lack of the Early Oligocene taxa (e.g. *Chropteridium* spp.). Boundary intervals are very rich in volcanic biotite, similar to the biotite-bearing layers at the Massignano section (e.g. Coccioni et al. 2000), Eocene/Oligocene sections in Hungary (Báldi 1984), etc.

The Eocene/Oligocene boundary is indicated in the uppermost part of the Globigerina Marls. The Oligocene marls and superimposed Menilite-type formations are strongly reduced in the quantity and diversity of foraminiferal microfauna. They are dominated by small-size globigerinids of the *G. praebulloides–G. officinalis* plexus. The small globigerinids contain paragloborotaloids, tenuitellids and chiloguembellinids, which are considered to be the index fossils of the early Rupelian. Most of these species, like *Dentoglobigerina tapuriensis*, *D. tripartita*, *Turborotalia ampliapertura*, *Paragloborotalia opima nana*, *Tenuitella gemma*, *T. liverovskae*, *Chiloguembelina cubensis*, *Ch. gracillima*, etc., exhibit the FAD or ACME within the O1/P18 Biozone. Some of the associated species already indicated the O2/P19 Zone (e.g. *Dentoglobigerina selli*, *Paragloborotalia opima pseudocontinuosa*, *P. semivera*, *Parasubbotina carpathica*, etc.). The mid-Rupelian formation is barren in content of foraminiferal fauna, but it contains the endemic nannofossil species of *Reticulofenestra ornata* (NP23 Zone). In this part of the sedimentary sequence, the Menilitic shales are associated with tuffaceous horizons, Tylawa-type limestones and biosilicite horizons.

The Upper Rupelian sequence became flysch-like in character, since it contains poor microfauna. Planktonic foraminifers suggest their possible attribution to the O2–O3/P19–P20 Biozones and this is based on the presence of younger species such as *Tenuitellinata angustiumbilicata*, *Tenuitellinata postcretacea*, *Tenuitella brevispira*, *Tenuitella evoluta*, etc., and abundance of chiloguembelinids. Superimposed units of the Sub-Tatras Group are stratigraphically shifted to the late Rupelian (LAD of chiloguembelinids in the basal part of the Huty Formation related to the NP23/NP24 boundary — Van Simaeys et al. 2004), early Chattian (FAD of "*Globigerina*" *ciperoensis angulisuturalis* in the Zuberec Formation — O4/P21a Biozone) and late Chattian to early Aquitanian (FAD of *Discoaster drugii* and *Helicosphaera scissura* in the topmost part of the Biely Potok Formation = Ostrysz Beds). The biostratigraphy of the CCPB was updated by Molnár et al. (1992), Vass et al. (1993), Olszewska & Wieczorek (1998), Soták (1998a,b), Starek et al. (2000), Gedl (2000a,b), Soták et al. (2001, 2007), Garecka (2005), etc.

Proxies and methods

Paleoenvironmental changes in the CCPB are indicated by various proxies. Their identification is based mostly on the foraminiferal microfauna, using climatic index taxa, species abundance and diversity, life-mode strategies, habitat-groups, morphotypes, trophic preferences, dissolved-oxygen index, coiling directions and paleobathymetry (e.g. Premoli Silva & Boersma 1988; Spezzaferri & Premoli Silva 1991; Keller et al. 1992; Spezzaferri 1995; Van Eijden 1995; Spezzaferri et al. 2002; Bicchi et al. 2003; Molina et al. 2006; Wade et al. 2007; Wade & Pearson 2008, etc.). Besides foraminifers, the paleoenvironmental conditions have also been inferred from calcareous nannofossils, organic-walled dinoflagellates, diatoms, moluscs, pteropods and fish fauna (e.g. Rusu 1995; Diester-Haas & Zahn 1996; Monechi et. al. 2000; Pross & Schmiedl 2002; Van Simayes et al. 2004). Geochemical proxies have been used to approximate the isotopic paleotemperatures (after Craig 1965 — $T = 16.9 - 4.2\,[\delta^{18}O_c - \delta^{18}O_w] + 0.13\,[\delta^{18}O_c - \delta^{18}O_w]^2$), atmospheric CO_2 concentration, high-rate productivity (trace metals), organic-rich deposition, trophic conditions and low-oxygen environments (e.g. Zachos et al. 1996; Murphy et al. 2000; Van Breugel 2006; Vetö et al. 2007).

Eocene/Oligocene events in the CCBP

The multiproxy study allows us to interpret the paleoenvironmental changes of platform drowning, climate cooling, biomass productivity, eustatic events, CCD deepening, continental run-off and riverine input, eutrophication, widespread anoxia, water-mass stratification, sapropelitic deposition, volcanogenic activity and semi-isolation in the CCPB.

Demise of the carbonate platform

The carbonate platform — or ramp-type basin — (cf. Wright & Burchette 1998) was established by the sea-level rise during the Middle Eocene Climatic Optimum. Climatic control of nummulid-rich production implies the tropical-type carbonate ramps in the CCPB (cf. Schlager 2003). Their development in the CCPB culminated during the SBZ17 Zone. Sedimentary units of the carbonate ramp are formed by keep-up cycles of nummulitic bank, back-bank, lagunar, rubble-flat and fore-bank facies.

Nummulitic ecosystems were adapted to clear-water and oligotrophic conditions, requiring temperatures above 20 °C (Sarangi et al. 2001). Such temperatures dominated during the late Lutetian–early Priabonian time, which is regarded as a thermochron in global climate (Rusu 1995). The oxygen isotopic composition of the nummulitic limestones in the CCPB ($\delta^{18}O = -2$ ‰) indicates a seawater temperature of around 22 °C. This isotopic temperature corresponds to the temperature derived from nummulitic tests in Hungary using the Ca/Mg method (21.4–25 °C, see Berlin et al. 1976). As well as temperature, the nummulitids were highly sensitive to even a small increase in nutrient availability and primary productivity (Hottinger 1983). Their life strategy was strongly dependant on a trophic regime, which had to be oligotrophic, due to their algal-bearing symbionts. Therefore, the nummulitids were not competitive in more nutrient-rich waters (Hallock 1987; Hallock et al. 1991). Open-marine microfossils from the nummulitic formations in the CCPB, such as porticulosphaerids and discoasterids, belong to the warm-water and low-nutrient taxa (Figs. 6, 7). Nummulitic facies are barren in dinoflagellates, which were not favoured by oligotrophic waters (Gedl 2000a). The mollusc fauna of the Lutetian formations in the CCPB consists of thermophile forms, such as spondylids, pectinids, ostreids, etc. (Papšová in Gross & Köhler et al. 1980).

Nummulitic fauna of the CCPB began to disappear in stressful conditions, inferred in climatic cooling and nutrient excess. Cooler and mesotrophic conditions are indicated from the Bartonian (EBi 1 shift to lower temperatures sensu Abreu & Anderson 1998), when the sequence increases in the abundance of orthophragminids and heterosteginids (Fig. 6, after Bartholdy et al. 1999). The early Priabonian transgression renewed growth production on carbonate ramps. Their re-establishment occurred during the last warm-temperate conditions

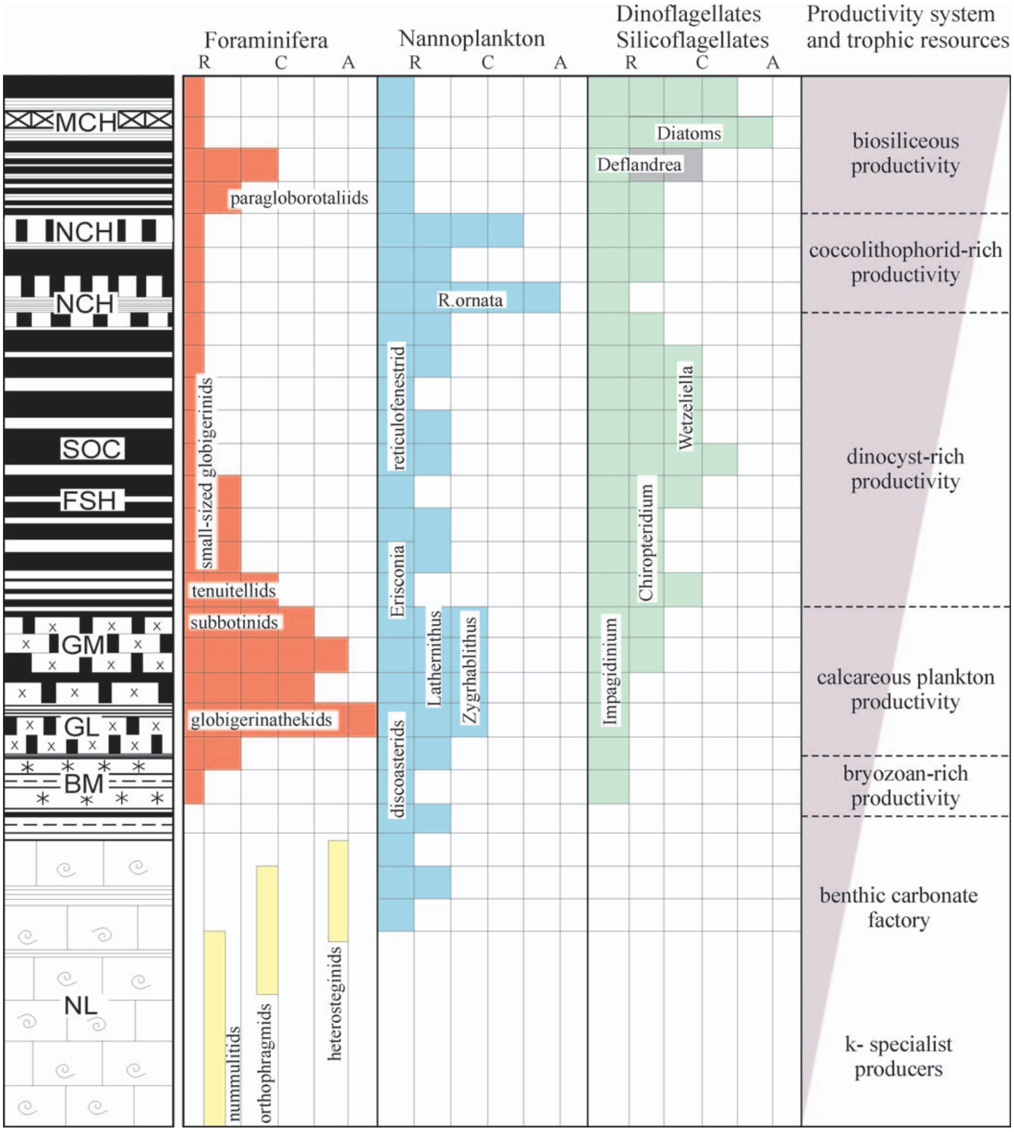

Fig. 6. Productivity changes in the Central-Carpathian Paleogene Basin. The Late Eocene–Oligocene sequence exhibits the upward decrease in carbonate productivity, increase in organic-rich productivity with episodical blooms of calcareous plankton and after it the predominance of biosiliceous productivity (results derived from foraminiferal, coccolithophorid, dinoflagellate and diatom-based data). Abbreviations see Fig. 4.

of the SBZ19 Zone. The eustatic rise during the early Priabonian is well documented by aggradation of carbonate ramp over structural highs and this is recognized by the incipient flooding surfaces on the Mesozoic strata (e.g. in Western High Tatra sections). This transgressive phase reached maximum flooding in formations with *Nummulites millecaput* (Köhler 1998), which as giant nummulitids benefitted from the low energy environment with lower light intensity and fewer nutrients (Hallock & Glenn 1986). A major turnover to climatic cooling occurred around the demise of the nummulitids (cf. Geel 2000). The nummulites-bearing formation of the CCPB reveals the deepening-upward tendency in basin paleobathymetry (e.g. in the Jobová Rázstoka section, Važec-Priepady section). Platform drowning is demonstrated by the distribution of benthic foraminifers in the Borové Formation, showing the gradual decrease in shallow-water taxa (*Elphidium*, *Parrellina*, *Calcarina*, etc.) and the increase in deep neritic to shallow bathyal taxa (e.g. *Uvigerina*, *Bolivina*, etc.). As a consequence of cooling, nutrification and drowning, the warm-water carbonate deposition in the CCPB ceased with the latest nummulites in the P16 Zone = SBZ20 Zone (Köhler 1998), or in the SBZ21 Zone (Buček & Filo 2004).

Paleoenvironmental changes in the CCPB are apparent from the development of the temperate-type carbonate ramp with foramol or foralgal facies in the Priabonian (sensu Carannante et al. 1988). Unlike the tropical-type carbonates, the foramol facies of the CCPB (e.g. in Hybica section) are poor in nummulitids and rich in skeletal grains, benthic foraminifers, crustose algae and bryozoans. Widespread foramol-type carbonates in the Priabonian also resulted from progressive eutrophication of the CCPB, since cooler waters are usually richer in nutrients. The large foraminifers are represented

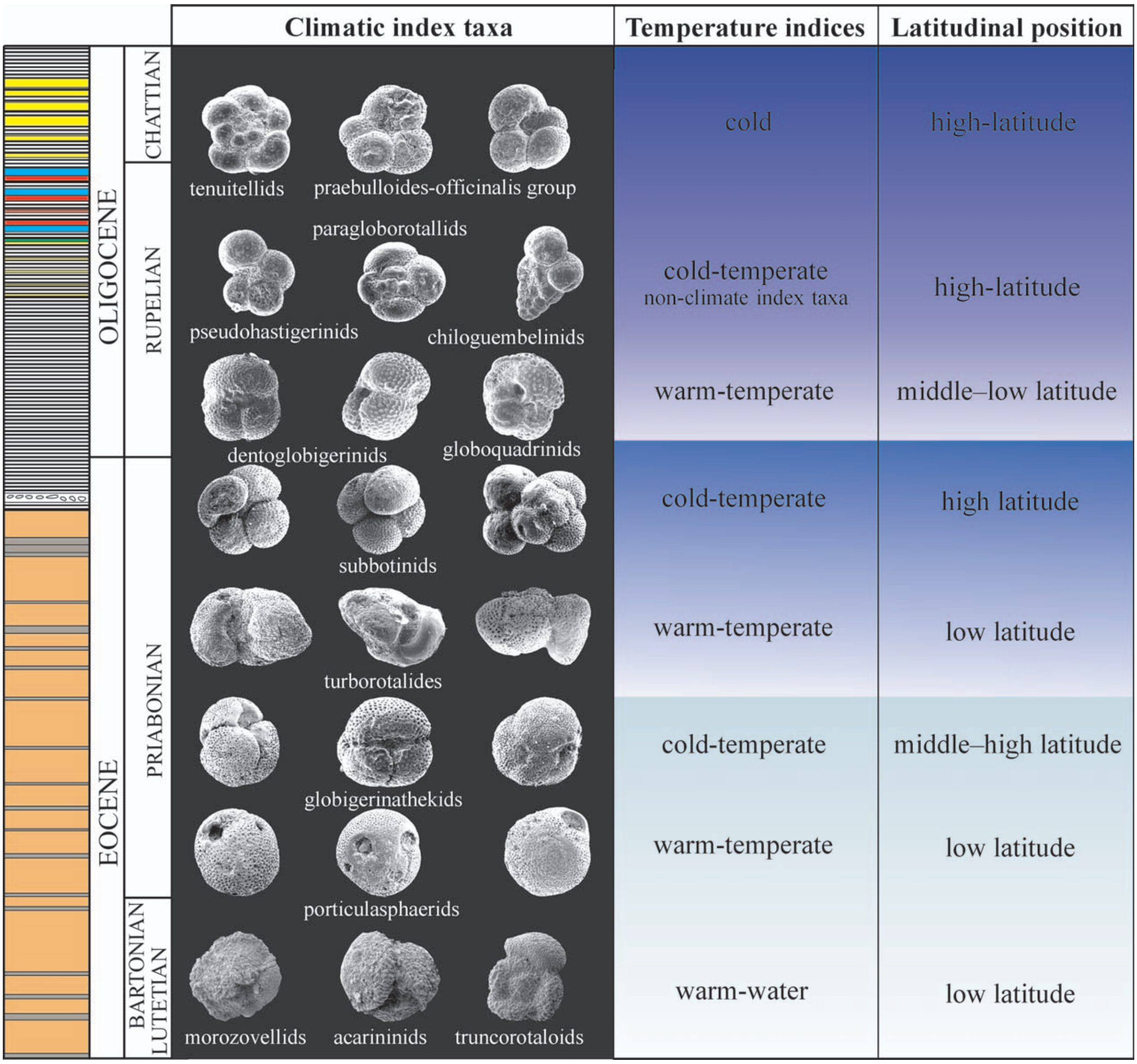

Fig. 7. Climatic index taxa of the planktonic foraminiferal microfauna from the Central-Carpathian Paleogene Basin. Temperature indices of planktonic foraminifers provide evidence of three-fold cooling, marked by increased colour intensity, which culminated in cold and high-latitude climatic conditions in the Middle Oligocene.

mainly by orthophragminids (e.g. *Discocyclina*), which tolerated lower temperature and higher dispersion, and preferred a lower energy environment, greater depth and a soft substrate (Rasser et al. 1999). In addition to orthophragmids, the Priabonian carbonates are also rich in bryozoans, which are recognized as the most abundant organisms on the cool-water shelves (Surlyk 1997; Light & Wilson 1998). The CCPB shelves were bypassed by the Bryozoan Marls, consisting of numerous species and various growth forms (Zágoršek 1992, 2000). The mass abundance of bryozoans indicates temperate cool-water and mesotrophic growth conditions (Fig. 6), and these also facilitated the bryozoan reef expansion in glacial periods during the Quaternary (cf. James et al. 2000; James & Bone 2000). Prosperity of bryozoan fauna usually indicates elevated resources of land-derived nutrients, or the availability of nutrients via upwelling (e.g. Light & Wilson 1998). The Late Eocene was a time of widespread appearance of bryozoan-rich sediments, as in the Buda Basin (Kázmer et al. 1993), North Apennine Basin (Braga & Barbin 1988), Adriatic Basin (Marjanac et al. 1988), Alpine Foreland Basin (Rasser et al. 1999), Australian basins (James & Bone 2000), etc. Therefore, the Bryozoan Marls imply similar paleoenvironmental conditions of deposition (climatic cooling, nutrification, upwelling, etc.).

High productivity rate of planktonic foraminifers

Progressive cooling in the CCPB induced changes in the productivity rate and temperature preferences of planktonic foraminifers. During the Middle to Late Eocene, the planktonic foraminiferal productivity decreased in warm-water species and increased in bloom-forming globigerinids (Fig. 6). Lutetian pelagic sedimentation resulted mostly from the warm-water productivity, which is indicated by muricate species, that is foraminifers heavily calcified due to a high content of dissolved carbonates. Muricate morphotypes, such as morozovellids, acarininids and truncorotaloids, are the most common foraminifers in the Lutetian microfauna of the CCPB. Warm-water productivity of calcareous nannoplankton is documented by *Discoaster*-dominated associations. Muricate foraminifers and discoasterids are oligotrophic organisms (Monechi et al. 2000; Poletti et al. 2004), which benefited under the nutrient-poor conditions during the Lutetian–Bartonian.

Foramimiferal microfauna indicate a gradual impoverishment of paleoenvironmental conditions following the Bartonian/Priabonian boundary (Fig. 7). In the CCPB, the last muricates disappeared between the P14/P15 Zones. Priabonian associations of foraminifers differ markedly, with an increasing content of porticulasphaerids and later prevalence of globigerinathekids. The early Priabonian species of *Porticulasphaera semiinvoluta* is still considered to represent warm-temperate habitats, such as the species of the “mexicana” group (Bolli 1972). On the contrary, the globigerinathekids were interpreted as subtropical and cooler water forms, which inhabited middle-latitude sites (Blow 1969; Premoli Silva & Boersma 1988; Boersma & Premoli Silva 1991). The most recent studies of globigerinathekids (Spezzaferri et al. 2002; Galeotti et al. 2002; Payros et al. 2006) point to their temperate to cold-temperate preferences and mesotrophic character (Fig. 7). The abundant increase in globigerinathekids in the middle Priabonian marls provided the first indication of progressive climatic cooling in the CCPB. Consequently, warm-water species, such as those of the *Turborotalia cerroazulensis* group, became less common, and truly tropical forms are missing here (e.g. tubulospinose hantkeninids). Globigerinathekids disappeared abruptly in the middle of the P16 Zone, and they were replaced by cooler late Priabonian foraminiferal microfauna in the Globigerina Marls.

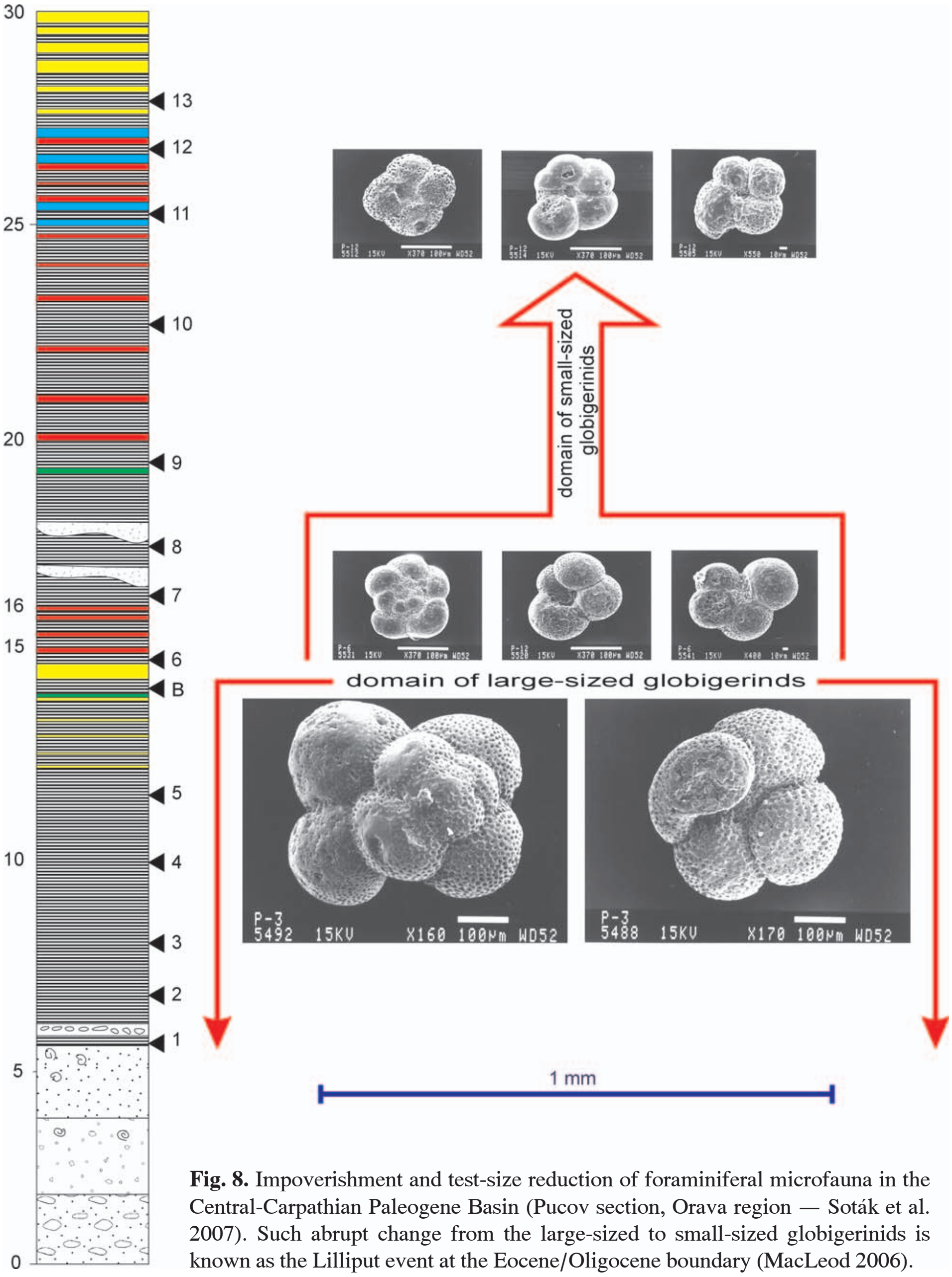

Fig. 8. Impoverishment and test-size reduction of foraminiferal microfauna in the Central-Carpathian Paleogene Basin (Pucov section, Orava region — Soták et al. 2007). Such abrupt change from the large-sized to small-sized globigerinids is known as the Lilliput event at the Eocene/Oligocene boundary (MacLeod 2006).

During the Late Eocene, the productivity of the CCPB increased considerably in response to bloom-forming globigerinids. So-called “Globigerina Marls” are considered as *prima facie* (Van Couvering et al. 1981), which denotes a prominent increase in bioproductivity around the Eocene/Oligocene boundary. Their microfauna consists of “globigerinids” with cool-water preferences (cf. Olszewska 1983). In the CCPB, the Globigerina Marls are dominated by subbotinids, which attained large growth size, high population density and lower diversity (Fig. 8). Subbotinids represented the cold-temperate habitats of the Eocene microfauna (Boersma & Premoli Silva 1991; Pardo et al. 1999; Spezzaferri et al. 2002, etc.). Therefore, their mass productivity at the Eocene/Oligocene boundary was related to climatic cooling, since the cold waters were supersaturated with nutrients. This increase in trophic resources led to a proliferation of spinose taxa, such as subbotinids (Boersma et al. 1995).

Globigerina Marls lack an original isotopic record of seawater temperature due to their common recrystallization and reequilibrium during diagenesis. Therefore, their bulk-sediment isotopic composition attains the more negative δ^{18}O values (–5 ‰). The carbon isotopic composition of the Globigerina Marls ranges within normal δ^{13}C values for marine carbonates (0 to 0. 8 ‰), thus exhibiting no diagenetic overprint or vital effect.

Paleoenvironmental conditions of the Globigerina Marls are also reflected by changes in bottom-water productivity. Benthic foraminiferal frequency varied significantly (P/B = 4–35 %) and tended towards peak abundance and diversity in the *Tritaxia szaboi* horizons. These benthic-rich horizons of the Globigerina Marls imply an enhanced organic matter flux, and bottom-water ventilation with a preference for epifaunal and shallow-infaunal species (e.g. *Lenticulina cultratus*, *Marginulinopsis fragaria*, *Vulvulina haeringensis*, *Gemellides eocaenus*, *Eponides umbonatus*, *Cibicides ungerianus*, *Hanzawia ammophila*, *Gyroidinoides soldani*, *Lagena gracilicosta* and *Uvigerina rippensis*). Seafloor oxygenation by cold water is also expressed by occasional red colouring of the Globigerina Marls, and by their common bioturbation (mainly *Chondrites*-type burrows). In some horizons of the Globigerina Marls, there are no planktonics, and the benthic foraminifers are dominated by agglutinated taxa (e.g. *Ammodiscus polygyrus*, *Rhabdammina discreta* and *Glomospira charoides*). The changing ratios between calcareous and agglutinated foraminifers within the Globigerina Marls indicate the vertical fluctuations of the CCD in the dependance on the productivity, temperature and CO_2 concentration (Fig. 9).

The Calcite Compensation Depth (CCD) declines in the conditions of enhanced productivity like those in the Globigerina Marls, which correspond to a peak of productivity near 33.5 Ma (Diester-Haass & Zahn 1996), and this caused the CCD drop near the Eocene/Oligocene boundary (Thunell & Corliss 1986; Tripati et al. 2005). Following the downslope excursion of the CCD, the Globigerina Marls occur suddenly in carbonate-free deep-water sediments of the Carpathian basins (e.g. the Terchová Formation in the CCPB). Carbonate depletion of the Oligocene sediments in the Carpathian basins could have resulted from shoaling of the CCD due to a higher saturation of cold bottom-water by CO_2. As a consequence, the seawater became more acid and corrosive to the calcareous components. The acidification led to a high solubility of the calcitic microfossils, which disappeared in some horizons of the Globigerina Marls. Small-scale intercalations of non-calcareous black shales with the Globigerina Marls provide evidence of the vertical fluctuations of the CCD, CO_2 concentration and acidification, resulted from the productivity changes, climatic deteriorations and precessional cyclicity (cf. Krhovský 1995; Leszczyński 1997). Above the Globigerina Marls, the carbonate dissolution increased considerably, leading to a non-calcareous deposition of the Menilite Formation.

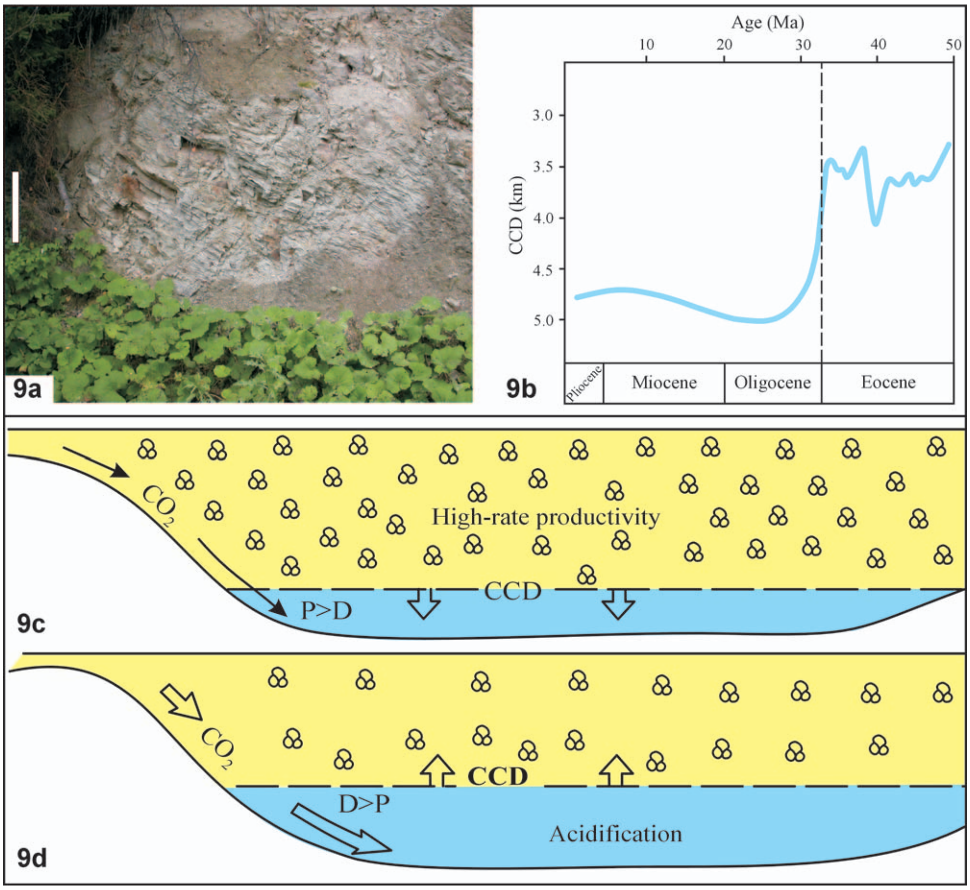

Fig. 9. Productivity driven fluctuation of the CCD during the deposition of the Globigerina Marls (9a — loc. Údol, scale bar = 1 m). High-rate productivity led to the CCD drop (9c), which corresponds to CCD deepening across the Eocene/Oligocene transition (9b — Tripati et al. 2005). Productivity (P) decrease caused a higher rate of dissolution (D), which tend to the CCD rise (9d).

Biotic crises and rapid cooling

Climatic changes culminated in the "Terminal Eocene Event", which corresponds to the global cooling and glacio-eustatic regression related to the Antarctic cryosphere expansion. Since the Late Eocene, a cold climatic phase followed (cryochron sensu Russu 1995) and the ocean temperature fell about 2–5 °C worldwide (Shackleton & Kennett 1975; Pomerol & Premoli Silva 1986). In the CCPB, the paleotemperature drop is indicated by the appearance of cool-water molluscs, such as *Nucula* and *Glycymeris* (e.g. in the Odorín Limestones, Víťaz locality — Marschalko & Volfová 1960; Volfová 1964a). The Late Eocene–Early Oligocene cooling is also manifested in land flora and palynoassociations, recording the onset of Arctotertiary taxa (Snopková 1980; Konzalová et al. 1993). The cool-water influence in the CCPB is also documented by pteropods *Spiratella* (e.g. pteropod shells from the Blatná dolina locality — Orava, Szaflary Beds — Olszewska 1998), which as the polar epiplanktonic fauna expanded to the Paratethyan basins from the Boreal area, the so-called Spiratella Sea (cf. Báldi 1984).

Foraminiferal microfauna of the CCPB indicates a stepwise cooling from the Eocene/Oligocene boundary. Nevertheless, the earliest Rupelian microfauna still contains some temperate-water foraminifers, such as dentoglobigerinids, globoquadrinids and also the last turborotaliids (Fig. 7). These are already associated with cold-temperate and cold-preferring taxa (e.g. paragloborotallids, pseudohastigerinids and chiloguembelinids). Acceleration of cooling in the middle Rupelian is marked by cold-water species (e.g. tenuitellids and globigerinids). The cooling response to the planktonic foraminifers is most evident from their size reduction, reaching almost 80 % in the latest Eocene–Early Oligocene (Fig. 8). This size reduction is know as a dwarfing or Lilliput effect (MacLeod 2006; Wade et al. 2007). Moreover, the Early Oligocene associations of planktonic foraminifers were dominated by sinistral forms (Salaj 1998), and this is a coiling mode of foraminifers in the cold-water environment of high-latitude oceans (Norris & Nishi 2001). Considering this, the Early Oligocene foraminiferal microfauna of the CCPB reveals a dramatic size reduction, oligotaxic character, S coiling mode and an opportunistic life-mode strategy (bloom of chiloguembelinids), which is most likely related to cold-water environments.

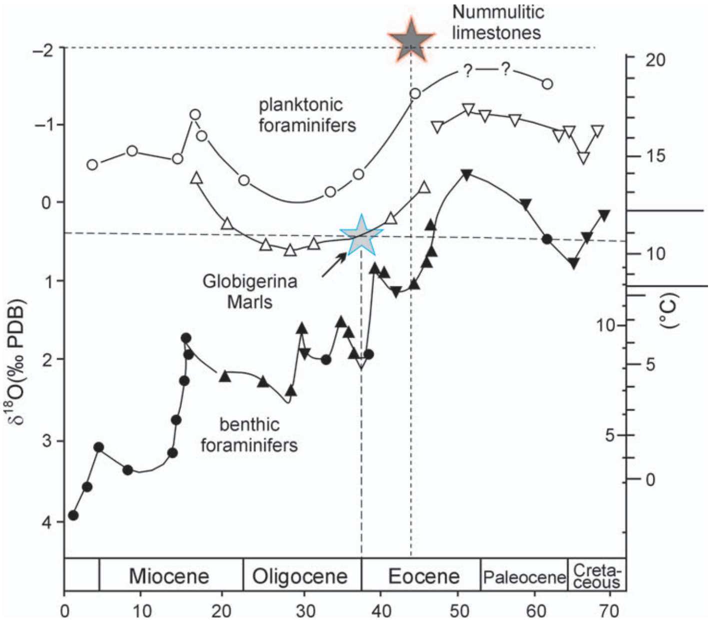

Fig. 10. Paleotemperatures calculated from isotopic composition of the Nummulitic Limestones and the Globigerina Marlstones by use of Craig's equation (Craig 1965). Isotope values $\delta^{18}O$ indicate a sea-surface temperature drop between the Late Middle Eocene (Nummulitic Lmst) and uppermost Eocene (Globigerina Mrls) from about 22 °C to 11 °C.

Early Rupelian cooling in the CCPB reduced the carbonate productivity by elimination of calcareous nannofossils, which occur only as cool-water taxa (e.g. *Isthmolithus recurvus*, *Zigrhablithus bijugatus* and *Reticulofenestra lockeri*). Later, the carbonate productivity changed to phytoplankton and biosilicite productivity, and this represented the productivity maximum in the earliest Oligocene (Diester-Haass & Zahn 2001). Phytoplankton of the Lower Oligocene formations of the CCPB dominated by Wetzelielloideae, whose large-sized dinocysts imply cool-water conditions (Gedl 2000a).

Late Rupelian microfauna became almost monospecific, and limited to globigerinid species of *G. praebulloides*–*G. officinalis* plexus. The small species of *Globigerina praebulloides* is considered to represent the subpolar planktonic foraminifers (Kennett 1982; Pak & Kenett 2002; Bicchi et al. 2003, etc.), or the upwelling-type foraminifers proliferated from the temperature drop and nutrient availability (Boersma et al. 1995). In times of flysch-type deposition in the CCPB (late Rupelian–Chattian), the planktonic foraminiferal productivity almost ceased due to water-mass turbidity, low oxygen concentration and a high sedimentation rate.

The paleotemperature decline in the Early Oligocene is also recorded in the oxygen isotopic composition of the limestones, interposed between the Globigerina Marls and Menilite Shale Formations. Oxygen isotopes exhibit a positive shift of the $\delta^{18}O$ values from –2 ‰ in the Middle Eocene limestones to 0.4 ‰ in the laminated limestones, implying a sharp temperature drop of seawater from 22 °C to about 11 °C (Fig. 10). An identical $\delta^{18}O$ signal was recorded by high-resolution isotope stratigraphy at the Eocene/Oligocene boundary in the ODP Sites (Diester-Haass & Zahn 1996) and type localities (Oberhänsli et al. 1984). The positive $\delta^{18}O$ excursion links with the Oi-1 shift to cooler temperatures (see Abreu & Anderson 1998), which corresponds to maximum productivity increase of surface waters (Salamy & Zachos 1999; Diester-Haas & Zahn 2001), eustatic sea-level fall and tefrostratigraphic events at the Eocene/Oligocene boundary (Fig. 11). The carbon isotopic values of laminated limestones (–4 ‰ δ^{13} C) indicate ^{12}C-enriched CO_2 due to respiration and planktonic foraminiferal metabolism (mainly *Globigerina praebulloides*). The higher respiration rate of these foraminifers resulted in the incorporation of more ^{13}C-depleted metabolic CO_2 during calcification (Bemis et al. 2000).

From the Eocene/Oligocene transition, the rapid cooling and high nutrient supply inhibited calcium carbonate production in the Carpathian Paleogene basins. Low activity of CO_3^{2-} resulted in lower seawater pH and a high organic carbon concentration (Lyle et al. 2005). Therefore, the organic-rich sediments of the overlying beds indicate the conditions of carbonate undersaturation, high solubility of the silicid acid (e.g. bloom of silicoflagellates, presence of silicified woods, etc.), sulphate reduction of organic matter, mobility of the redox-sensitive elements (e.g. Mn, P, Ba) and

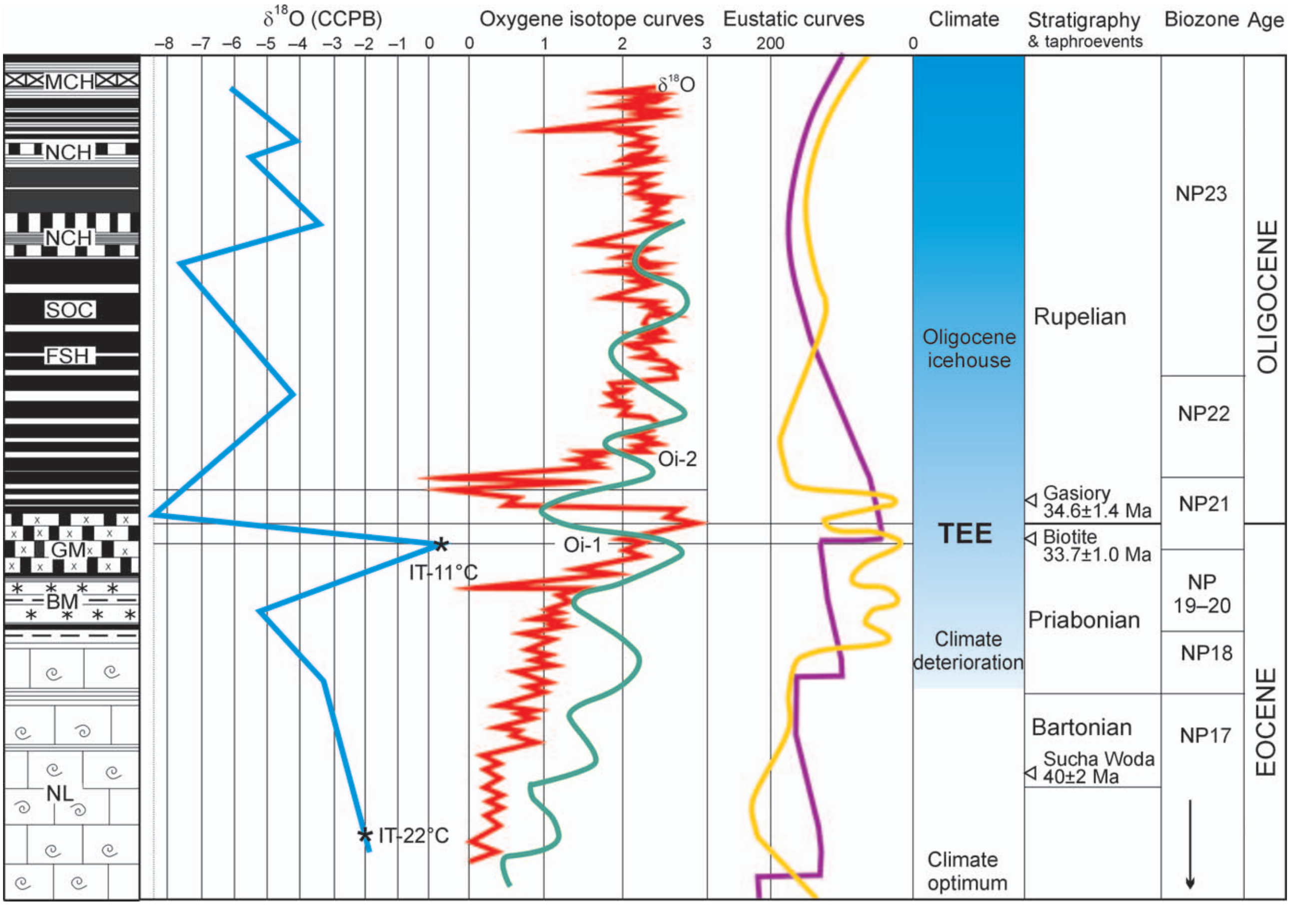

Fig. 11. Correlation of isotopic signals from the CCPB with global isotopic curve, eustatic curve and tafrostratigraphic events. This integrated approach allows us to approximate the Eocene/Oligocene boundary. Oxygen isotope curves after Zachos et al. (1993), Diester-Haass & Zahn (1996), etc.; eustatic curves after Vail et al. (1977) and Haq et al. (1988). Abbreviations see Fig. 4.

methanogenesis. These processes could produce HCO_3^-, increasing the seawater alkalinity and resulting in the formation of the methan- and sulphate-derived carbonates (so-called pelocarbonates).

Anoxia and eutrophication

The Early Oligocene period was a time of widespread anoxia and eutrophication in the Carpathian Paleogene basins, and this led to sapropelitic and biosiliceous deposition of the Menilite facies (cf. Roth & Hanzlíková 1982). Similarly, water-column anoxia, eutrophication, dinoflagellate- and diatom-based productivity, high export flux, elevated chemocline, monsoonal precipitation, interface cycles of coccolithophorid-rich productivity, reduced precipitation and continental influx (Fig. 13) also occurred in the CCPB.

The basin-wide anoxia in the CCPB is indicated both in bottom-water and surface-water environments. Poorly oxygenated bottom sediments were inhabited only by deeper infauna of benthic foraminifers, such as *Virgulinella* (*Loxostomum*) *chalkophilla*, *Chilostomella tenuis*, *Ch. cylindroides*, *Fursenkoina acuta* and *Bulimina pyrula*, etc. The benthic productivity of flysch sediments increased temporarily during the recolonization of the sea-bottom by tubular morphogroups of agglutinated foraminifers, such as *Dendrophrya-Rhabdammina* type associations in the sedimentary formations of the CCPB (Benešova 1962; Blaicher 1973; Samuel in Gross et al. 1999, etc.). Oxygenation of the CCPB improved up to the Huty Formation (NP24), which represents a highstand phase of mud-rich deposition with higher calcareous precipitation and appearance of trace fossils (e.g. *Zoophycos* ichnoceonoses). The shelfal sediments above the oxygen minimum zone OMZ (e.g. in Horehronská kotlina Depression) are much richer in benthic meiofauna (Samuel 1975; Zlinská et al. 2001).

The expansion of the OMZ to surface-water is inferred from the impoverishment of planktonic foraminifers, and in proliferation of euryoxibiont forms. Lower oxygen conditions are expressed by blooms of chiloguembelinids, which occurred in the CCPB during the early Rupelian. *Chiloguembelina*-dominated assemblages are indicative of hypoxic to anoxic conditions (Pardo et al. 1999), and this reveals life stress in the productive zone (similar to the *Guembelitria* blooms near the K/T boundary — Keller & Pardo 2004, etc.). Evidence of similar conditions is provided by wetzeliellacean dinoflagellates, diatoms and bacterioplankton.

Eutrophication of the Carpathian Paleogene basins led to replacement of calcareous productivity by organic-rich and biosiliceous productivity (Fig. 6), because the dinoflagel-

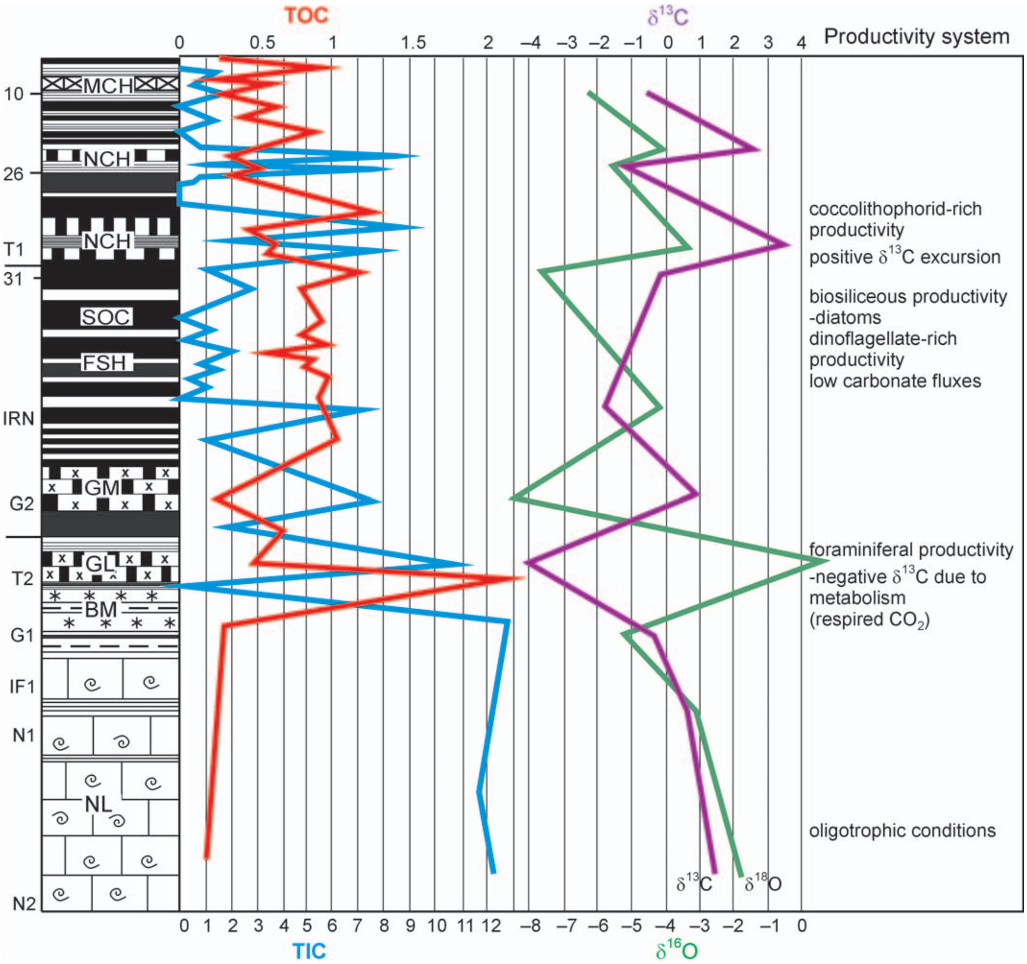

Fig. 12. Chemostratigraphic data from the Central-Carpathian Paleogene sequence ploted in the TOC, TIC and isotopic curves. Abbreviations see Fig. 4.

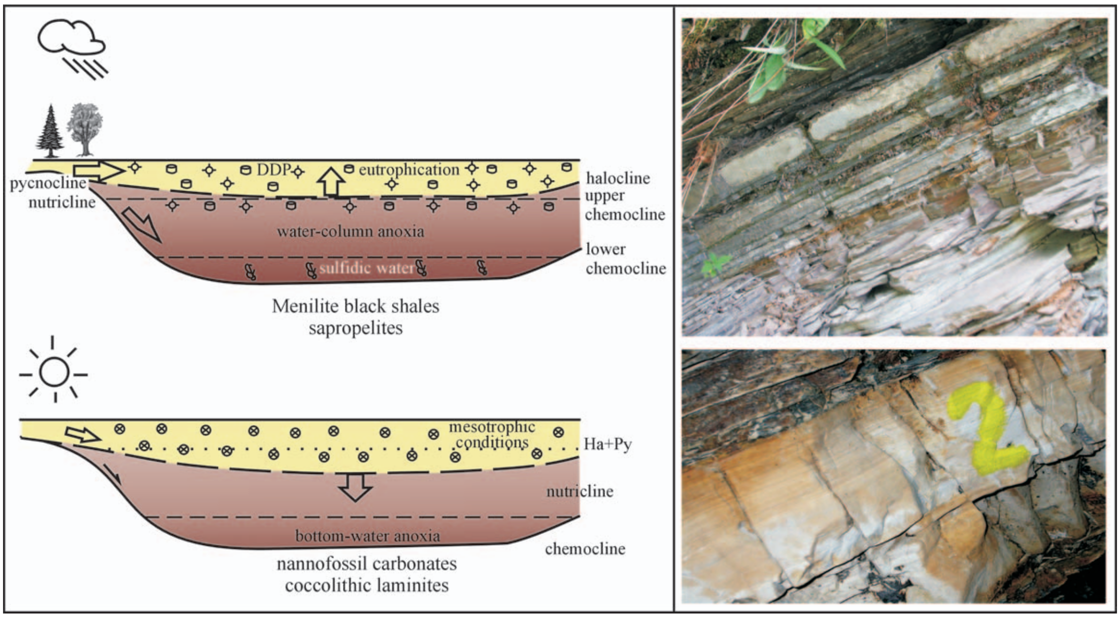

Fig. 13. A model of sapropelitic deposition in the Carpathian Paleogene basins. Menilitic sapropelites were deposited under a high continental runoff, increased nutrients (eutrophication and elevated nutricline), dinoflagellate and diatom-based productivity, oxygen depletion (anoxia and elevated chemocline), etc. Calcareous layers is considered to represent the nannofossil limestones, deposited under a high rate of coccolithophorid productivity, mesotrophic conditions (deeper nutricline), weakening anoxia (deepened chemocline), etc.

lates and diatoms utilized nutrients more efficiently than foraminiferids and coccolithophorids. Therefore, the eutrophication in the CCPB was responsible for dinoflagellate blooms which were dominated by peridionids (Gedl 2000a,b). This trend culminated in catastrophic eutrophication and the mass expansion of diatoms (Menilite cherts in upper NP23 — Nagymarosy 2000). The diatomite origin of the Menilite cherts was documented by discoveries of frustules several times (e.g. Kaczmarska & Kilarski 1979; Krhovský 1981b, etc.), and by molecular record of the C_{25} highly branched isoprenoids (Rospondek et al. 1997). Pyritized diatoms are also frequently detected in the supra-menilite formations of the CCPB (e.g. Šambron Beds, Huty Formation, Brzegi Beds). As well as the evidence provided by diatoms, the increase in seawater $Si(OH)_4$ is marked by the appearance of spicules, benthic sponges and silicified woods in sediments of the Menilite facies (Starek 2001; Soták et al. 2007). Silica availability in the CCPB resulted from the riverine input of nutrient-rich water from adjacent continents. Consequently, the diatomites from the Menilite Formation consist predominantly of fresh-water species, such as *Melosira distant*, *M. islandica*, etc. (Řeháková in Krhovský 1981a). The marine-continental interference in the CCPB is documented by tidal-marsh foraminifers (mainly *Trochammina* species) in the coastal plain sediments of the Spiš area, abundant presence of land flora and riparian-type vegetation (leaves, woody fragments) and even by rare avifauna (finding of a bird leg at the Bystré locality, Eastern Slovakia — Fig. 14). Salinity decrease in the surface-water layer is also indicated by *Deflandrea*-dominated associations of dinocysts in the Early Rupelian sediments of the CCPB (Gedl 2000a). Nevertheless, the silica production from a volcanogenic source could also not be excluded, because the Menilite Shales in the CCPB are occasionally intercalated by tefra horizons (Gross 1981) or biotite-rich layers (Soták et al. 2007). The increase in primary productivity from volcanic-derived nutrients is considered to be important at the Eocene/Oligocene boundary (e.g. Massignano section — Monechi et al. 2000), Early/Middle Miocene boundary (e.g. Styrian Basin — Spezzaferri et al. 2002, etc.). Therefore, the excess of nutrients may have also been responsible for the impoverishment of the Oligocene foraminiferal microfauna in the CCPB.

Primary productivity in the CCPB culminated in hypertrophication and oxygen depletion. The euxinia of the photic zone is considered to be an important factor in the deposition of organic-rich sediments, such as sapropels (Koopmans et al. 1996; Bosch et. al. 1998). Sapropelitic character of the Menilite-type facies in the CCPB is indicated by lowered

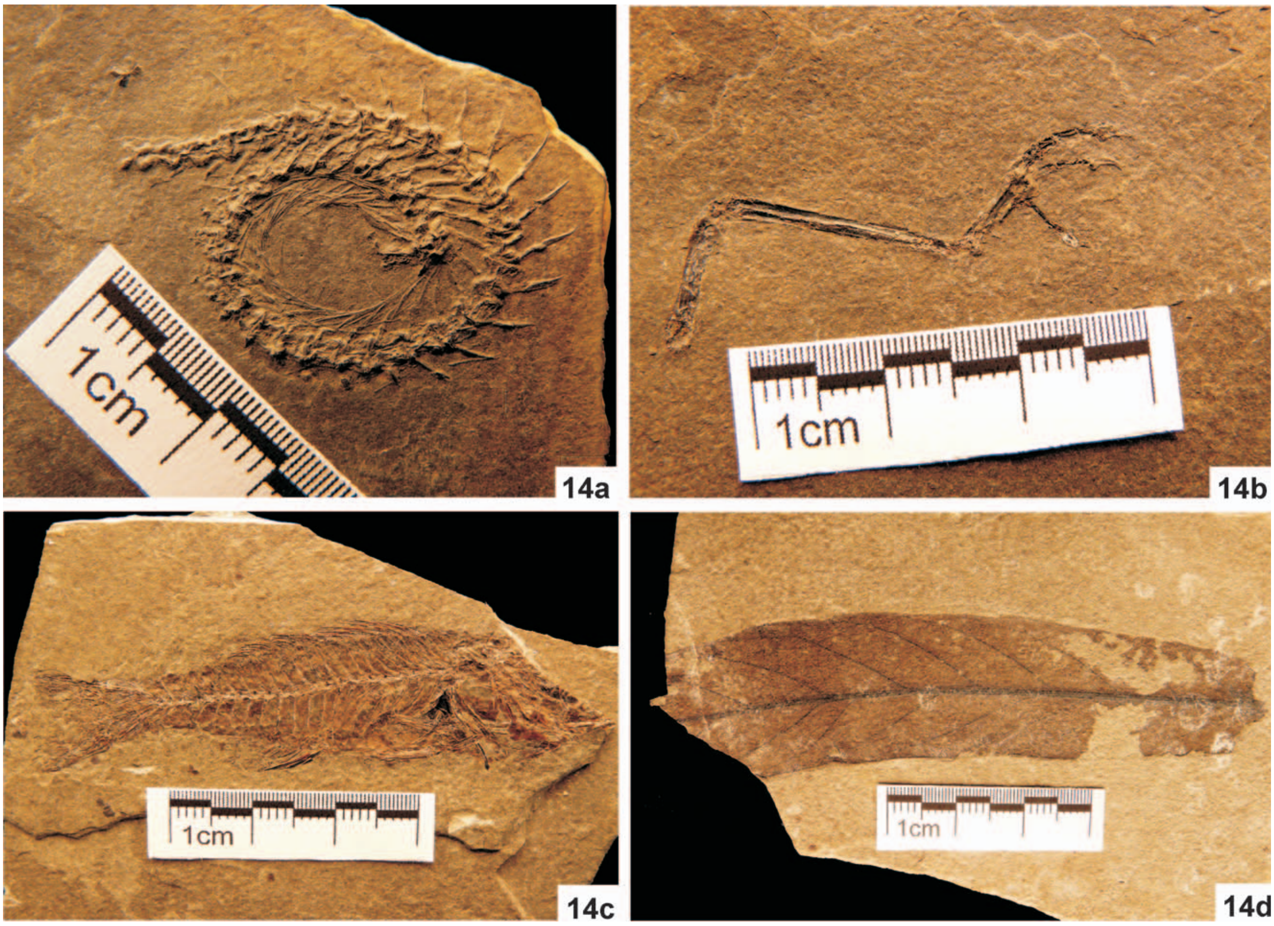

Fig. 14. Rich fossil fauna of fishes (Pl. 12a,c), bird's limb (Pl. 12b — first discovery) and leaves (Pl. 12d) from the Menilite shales of the Central-Carpathian Paleogene Basin (Bystré locality). Excellent preservation of fossils is caused by bottom-water anoxia.

TIC and elevated TOC contents (Fig. 12). They approach up to 8 wt. % TOC, and the organic matter yields a high content of bituminite and alginites (Kotulová 2004). Anaerobic conditions in the photic zone prevented the expansion of bacterioplankton, which required light and hydrogen sulphide. The photosynthetic activity of green sulphur bacteria in the anoxic sediments is revealed by isorenierante derivates (see Koopmans et al. 1996; van Breugel et al. 2005), and their presence was identified in the sediments of the Menilite facies (e.g. Schulz et al. 2003). In the CCPB, the anoxygenic water column and microbial photosynthesis is indicated by carbon isotope composition of the Menilite Shales, which $\delta^{13}C$ values are negative (–0.8 to –7.9 ‰) or extremely low close to isorenierante (up to –21 ‰). Moreover, the abundance of pyrite framboids in the Menilite Shales implies a sulphidic water column (see Wilkin et al. 1997). Sulphate-reducing bacteria react with ferrous iron to form monosulphides and pyrite. Anoxic conditions promoted bacterial decomposition of organic matter and provided a source of N. Nitrification in the CCPB elevated the ammonium concentration which supported phytoplankton biomass and eutrophication.

The nannofossil crisis in the Menilite facies was interrupted by episodes of coccolithophorid-rich productivity, which is indicated by the presence of laminated pelagic limestones (Tylawa Limestone) and related carbonates with an almost complete dissolution and overgrowth of coccoliths due to recrystallization and dolomitization. The increase in coccolithophorid productivity was connected with a decrease in diversity (Bubík 1978). More abundant nannoplankton species in these limestones represent high-nutrient taxa (e.g. *Cyclicargolithus floridanus*) and Paratethys-endemic or even salinity-reduced taxa (e.g. *Reticulofenestra ornata*, *Transversopolis fibula* and *T. lata*). Their monospecific associations are recognized as Solenovian (Polbinian) nannoflora of the Central and Eastern Paratethys (Nagymarosy & Voronina 1992; Nagymarosy 2000; Melinte 2005). Nannoplankton bloom in the NP23 Zone was probably prevented by short-term increase in eustasy, salinity, trophic resources and renewed circulation (cf. Krhovský & Djurasinovič 1992; Nagymarosy & Voronina 1992; Schulz 2003; Schulz et al. 2004). Calcareous vs. organic-walled plankton productivity depends most likely on the salinity and precipitation. During wet climates, the increased water runoff led to silicate concentration, reduction of salinity and eutrophication of the surface water in the CCPB. These conditions proliferated a high productivity of diatoms and dinoflagellates in the Menilite Shales. However, drier phases with reduced runoff and enhanced evaporation were responsible for nutrient depletion, improvement in water clarity, an increase in salinity and the dense blooms of calcareous nannofossils (Fig. 13).

Sapropelitic sediments in the CCPB (Menilite Shales) exhibit the carbon and oxygen isotope lightening ($\delta^{13}C$ –7.9 ‰, $\delta^{18}O$ –7.5 ‰) probably due to bacterial oxidation of organic matter. Such light carbon-isotope values in sediments of the Menilite Formation can also suggest a strong influence of fresh water (e.g. Schmiedl et al. 2002). Contrary to this, the coccolithic limestones exceeded the background values of the Menilitic formation, attaining a more positive isotopic values. These limestones correspond to short-lived positive $\delta^{13}C$ excursions (–0.5 to +3.4 ‰), coupled with less pronounced positive shifts of $\delta^{18}O$ (–3.4 ‰). Calcareous nannoplankton flourished in surface water with a high content of dissolved inorganic carbon ($\sum CO_2$). Consequently, the carbon isotopes became heavier in coccolith-rich productivity, and this led to a positive shift in surface-water $\delta^{13}C$ (Mitchell et al. 1996). Coccolithophorid productivity depends on atmospheric CO_2, availability of which increased during the greenhouse periods.

Widespread anoxia in the CCPB is also inferred from the manganese sedimentary ores considering their redox-controlled deposition (Roy 2006). Manganese became a mobile component in anoxic deep-water environments, whereas Mn-enriched waters precipitated in oxic environments on the shallow continental shelves (Nijenhuis 1999). The manganese ore basins of the Paratethys spread over South Ukraine, Georgia, Azerbayan, northwest Turkey, and northeast Bulgaria, etc. (Stolyarov & Kochenov 1995; Öztürk & Frakes 1995; Varentsov 2002; Efendiyeva 2004, etc.). The Early Oligocene deposition in the CCPB, like that in the Paratethyan basins (e.g. Maikop Basin, Tard Sea, Ileada Sea, Melleta Sea, etc.), took place under basin-wide anoxia, which produced metalliferous deposits (e.g. manganese ore beds near Kišovce and Švábovce in the CCPB).

The Menilite Shales of the CCPB contain rich ichthyofauna (*Clupeidae*, *Serranidae* — Chalupová 2000; Gregorová & Fulín 2001) — Fig. 14. Fish mortality resulted most likely from the toxicity of H_2S-rich water, which surfaced from the Oxygen Minimum Zone due to water-column stratification or via upwelling (such as “en masse” extinction of *Spiratella* fauna — Báldi et al. 1984). H_2S is known to be an extremely potent neurotoxin causing a cessation of pulmonary function (Oschmann 1995). The toxic crush of fish fauna in the Menilite Shales is indicated by their curved backbones, unattached mouths and mostly adolescent forms (Pokorný 1992). The alternation of fish shales with barren sediments in the lamina scale points to seasonal changes in water circulation or upwelling activity (cf. Vetö 1987).

Widespread anoxic and eutrophic conditions in the CCPB most likely resulted from a high runoff and a positive water balance similar to that in the Black-sea type basin (cf. Schulz et al. 2005). Dissolved silicate, nitrate, ammonium, manganese may have been supplied from continental sources. Trophic resources could also have increased via upwelling, which preferentially regenerated nutrients from organic matter under anoxic conditions (cf. Ingall & Jahnke 1997; Murphy et al. 2000, etc.). Both, this runoff and upwelling resulted in eutrophication of surface-waters. Biomass productivity of surface waters in the Paratethyan basins resembles a drifting flora and conditions similar to those in the Sargasso Sea (Jerzmanska & Kotlarcyzk 1976; Kvaček & Bubík 1990, etc.). This algal biocoenosis is dominated by *Sargassum maikopicum* (Efendiyeva 2004).

Eustatic and cyclostratigraphic events

Paleoenvironmental changes in the CCPB could correspond to eustatic cycles of sea-level history (Haq et al. 1988) — Fig. 15. The basin began to develop from the late Lutetian

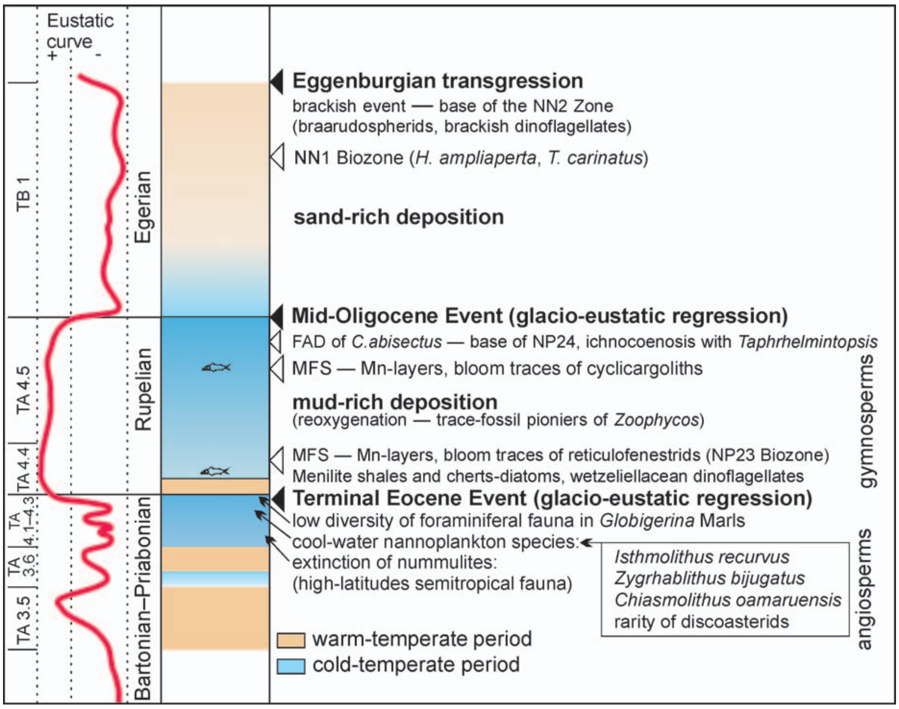

Fig. 15. Paleoenvironmental changes in the Central-Carpathian Paleogene Basin related to the global eustatic and climatic changes. Sedimentary sequence records the global eustatic events (Terminal Eocene Event and Mid-Oligocene Event sensu Aubreu & Anderson 1998), and provides the evidence of the 3rd order cyclicity (TA 3.5–TA 4.5) and alteration of thermochrons and cryochrons.

transgression (SBZ16), and this was followed by two 3rd order cycles of shallow-marine deposition (TA 3.5–TA 3.6). Compositional changes of large foraminiferal associations in the upper SBZ17 Zone indicate a relative cooling event near the TA 3/TA 4 boundary (Bartholdy et al. 1999). At that time, the sea-level fall resulted in channel incision and formation of fan delta systems (e.g. the Pucov Fan). New transgression of the TA 4.1 cycle started during the early Priabonian climatic optimum (last nummulites in the SBZ19 Zone, warm-temperate porticulosphaerids in P15 Zone), which preceeded very unstable conditions during the late Priabonian (TA 4.2–TA 4.3 cycles). The highstand of the TA 4.3 cycle is marked by the productivity maximum of the Globigerina Marls, which corresponds to the Terminal Eocene Event. The eustatic cycle TA 4.4 began with the largest sea-level fall near the Eocene/Oligocene boundary (50–65 m after Pekar 2003; Wade & Pälike 2004). In the CCPB, this eustatic event fits well with the positive $\delta^{18}O$ excursion, which marked the Eocene/Oligocene boundary on a world-wide scale (Cavelier et al. 1981; Poore & Matthews 1984; Kennet & Barker 1990; Miller 1992; Zachos et al. 1996). Boundary sections in the CCPB imply the erosional unconformities with a mass redeposition of morozovellids and acarininids (e.g. in Pucov section). Such types of deep-sea hiatuses, with reworked Eocene species, have been observed near the Eocene/Oligocene boundary in the DSDP sites (Keller 1985). These hiatuses were formed during the global cooling and sea-level lowstand, resulting from changes in oceanic deep circulation and bottom current erosion (Keller 1983). Considering this, the erosive events of muricate foraminifers in the CCPB indicate a lowstand phase between TA 4.3–TA 4.4 cycles at the Eocene/Oligocene boundary. This eustatic event was associated with carbonate depletion of the boundary clays, which are almost non-calcareous and are enriched by siliceous microfauna (e.g. sponge microsklerites in Pucov section).

Early Oligocene transgression in the CCPB is documented by a stepwise onlap of the ravinement surfaces of the Borové Formation (Bartonian ramp-type formation) by mudstone formations (NP22–NP23 Zones). The TA 4.4 cycle tended to a higher sea level in the horizons of coccolithic limestones, with the appearance of open-marine biota (e.g. pteropods). Forced regression of the TA 4.4 cycle culminated in incision of submarine channel systems of the CCPB during the Early Oligocene (e.g. Tokáreň Fan). The sea-level lowstand of the TA 4.4 cycle coincides with the beginning of the Paratethyan isolation which is marked by endemic nannofossil species and biosiliceous horizons (Menilitic facies). The transgression of the TA 4.5 cycle is expressed by the more calcareous and mud-rich deposition of the Huty Formation, which recorded the FAD and ACME of *Cyclicargolithus abisectus* in the NP24 Biozone. The falling stage of the TA 4.5 cycle links with the most pronounced Mid-Oligocene regression, which introduced the sand-rich turbidite systems in the CCPB. Late Oligocene deposition of the CCPB persisted in a lowstand eustasy of the TB 1 cycle, which is inferred from progradational sequences, an increased abundance of braarudosphaerids and reduced salinity. The sand-rich deposition should have terminated by the Early Miocene, recorded by the FAD of *Discoaster drugii* (23.8 Ma) and the appearance of rich uvigerinid species in the topmost part of the Biely Potok Formation which occurred earlier than the Eggenburgian transgression of the Prešov Formation.

Syngenetic volcanism

Transitional beds around the Eocene/Oligocene boundary in the Carpathian Paleogene basins provide evidence of syngenetic volcanism (Fig. 11). The first evidence of volcanic activity was recorded by the vitritic-crystal tuffs in the basal conglomerates beneath the Globigerina Marls (Glazek et al. 1998). Based on this, the tuffs correspond to the taphrostratigraphic event near 40 ± 2 Ma (Sochaczewski 2000). The most frequent tuffs of the CCPB occur above the Globigerina Marls (e.g. Prosiek, Huty, Pucov, Bystré), and this may correspond to the Gasiory Tuff (34.6 ± 1.4 Ma — Wieser 1985) and tuffs intercalated in the upper part of the Buda Marls and lower part of the Tard Clay (Báldi 1984, 1986). Tuffs from the Buda Basin have been approximated to the Eocene/Oligocene boundary (33.7 ± 1.0 Ma — Balogh & Pécskay 2001). As well as tuffs, there are also tuffaceous sandstones containing a great portion of biotite and volcanic quartz in the Orava region (e.g. Veličná, Pucov), and this appears to be equivalent of biotitic sandstones from the uppermost part of the Buda Marls in Hungary (33.7 ± 1.0 Ma — Balogh & Pécskay 2001). Similar biotite-rich layers are widely distributed (e.g. in Messignano GSSP section — Montanari et al. 1985; Monechi et al. 2000; Possagno section — Oberhänsli et al. 1984), and they record volcanic activity at the Eocene-Oligocene boundary. Biotite-rich layers in the CCPB are significantly enriched in content of biosiliceous microfossils (e.g. sponge spicules), and therefore the syngenetic volcanism may have enhanced the nutrient flux from products of coeval pyroclastic activity (cf. Spezzaferri et al. 2002; Vetö et al. 2007). Younger taphrostratigraphic events in the CCPB correspond to the NP24 Biozone (Uhlík et al. 2002), or even to the 14 marker horizons in the Chocholow Beds of the Podhale region (Westwalewicz-Mogilska 1986).

Volcanogenic activity is commonly recorded around the Eocene/Oligocene boundary. At the same time-horizon, there exists a lot of impactoclastic evidence, such as spherule-rich layers, iridium anomalies, shocked quartz, Ni-rich spinels, etc. (e.g. Montanari et al. 1993; Wei 1995; Clymer et al. 1996; Pierrard et al. 1998). Therefore, the Late Eocene climatic deterioration may have resulted from impact-related cooling (Vonhof et al. 2000).

Alpine orogenesis and Paratethyan isolation

Synorogenic basins, such as the CCPB, should reflect an important role of the Alpine tectonics. However, there is a problem due to the synchroneity of the Alpine tectonic phases and global eustatic events. The initial transgression in the Central Western Carpathians corresponds well with the Middle Eocene transgression in the Mediterranean Tethys, but also with the Illyrian tectogenesis. The Pyrenean tectonic activity changed the distribution of lands and seas in Central Europe, which was still a part of the North Peri-Tethys (Popov et al. 2004) and archipelago during the Late Eocene (Steininger & Vessely 2000). Unlike the high rate of Alpine subduction, the mountain uplift was less intensive and the Alps were mostly submerged (Frisch et al. 1998). Therefore, the isolation of the Paratethyan basins was also facilitated by the largest eustatic fall at the Eocene/Oligocene boundary (54 m — Pekar et al. 2003). The subsequent isolation of the Paratethyan basins led to the widespread distribution of marker horizons (e.g. planktonic pteropods, fish shales, coccolithic chalks, diatoms and manganese deposits — Fig. 18), which most likely indicates the predominance of global eustasy over regional tectonics. The late Pyrenean phase at the beginning of the NP24 Zone provided the strongest compressional event in the Carpatho–Pannonian basins (Tari et al. 1993). Since the Late Rupelian, the Carpathian flysch basins were supplied with sand-rich turbidite systems, implying reorganization from the passive-margin to active-margin fans (Soták et al. 2001). Therefore, a higher source activity should increase in response to tectonics. However, the Rupelian/Chattian boundary links with the most pronouced change in global eustasy, which was significantly lowered at the time of the first major glaciation in Antartica. Considering this, the eustatic force of sand-rich deposition in the Central Western Carpathians (incl. Krosno Facies of the Outer Western Carpathians) should also be important (cf. Krhovský & Djurasinovič 1992).

The isolation of the Paratethys resulted from progressive rise of the collisional wedge in the Alpine-Dinaric-Balkan internids (Báldi et al. 1984). Late Eocene–Early Oligocene uplift in the Dinarides, which became a landmass barrier, is recorded by terrestrial and fluviatile-lacustrine deposits (Tari & Pamić 1998). At that time, the Alpine collision achieved the Pyrenean phase, which is considered to be the main orogenic phase in the Alps (Trümpy 1980). The Alpine Foreland Basin lost connection with the Tethys, and this is recorded by the appearance of the endemic fauna in Solenovian time (Steininger & Wessely 2000). The Alpine collision induced a large-scale continental escape of the Carpatho-Pannonian terranes. The Pannonian Basin was detached from the Slovenian Basin and emplaced to the Intra-Carpathian area along the Insubric–Balaton and Mid-Hungarian Lineaments (Csontos et al. 1992; Fodor et al. 1999; Schmiedl et al. 2002). Accordingly, the escape tectonics seem to be responsible for the first isolation of the Carpathian Paleogene basins. The first isolation in the CCPB was already indicated by the zoogeographical separation of the Spiš Basin, where nummulitids are absent, and mollusc fauna indicate an affinity to the Transylvanian Basin and Caucasus — Aral area (Volfová 1962). Endemic fauna of Solenovian-type molluscs (*Ergenica*? sp., *Janschinella*? sp.) and cerites (*Tympanotonus* sp.) are also recognizable from the Paleogene sediments of the Hron Valley Depression (Volfová 1964b).

The Carpathian Paleogene basins were differentiated in sea-floor topography, bathymetry and water exchange. During the Eocene/Oligocene transition, the differentiation was recorded by the Globigerina Marls, deposition of which was controlled by anti-estuarine circulation, high rate calcareous productivity and lowered depth of calcium carbonate dissolution (CCD) — Fig. 16. The CCD declines in the conditions of enhanced productivity, which correspond to a peak of productivity near 33.5 Ma (Diester-Haass & Zahn 1996), and this caused the CCD drop near the Eocene/Oligocene boundary (Thunell & Corliss 1986; Tripati et al. 2005). Following the downslope excursion of the CCD, the Globigerina Marls and red-coloured marls occur suddenly in carbonate-free deep-water sediments of the Carpathian Paleogene basins. Successive cooling resulted in shoaling of the CCD due to a higher saturation of cold bottom-water by CO_2. As a consequence, the seawater became more acid and corrosive to the calcareous components. Above the Globigerina Marls, the carbonate dis-

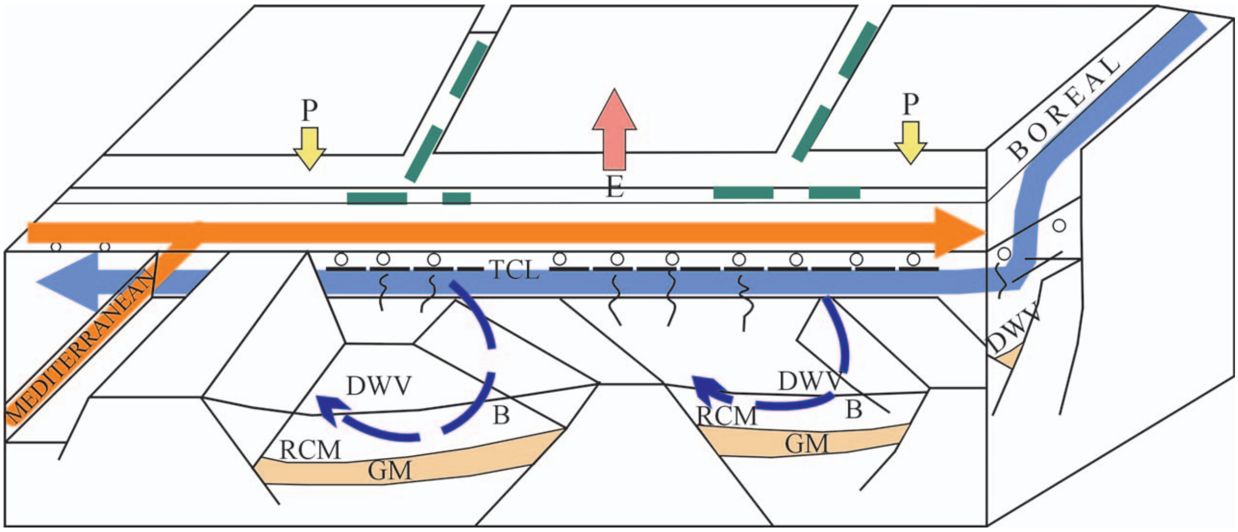

Fig. 16. Upper Eocene basins with the Tethyan–Boreal communication, which resulted in cold-temperate productivity of the Globigerina Marls (GM), thermal water mass stratification (TCL — thermocline), anti-estuarine circulation, deep-water oxygenization (DWV), proliferation of benthic life (B), red-coloured marl deposition (RCM), enhanced evaporation (E), lowered precipitation (P), reduced runoff, etc.

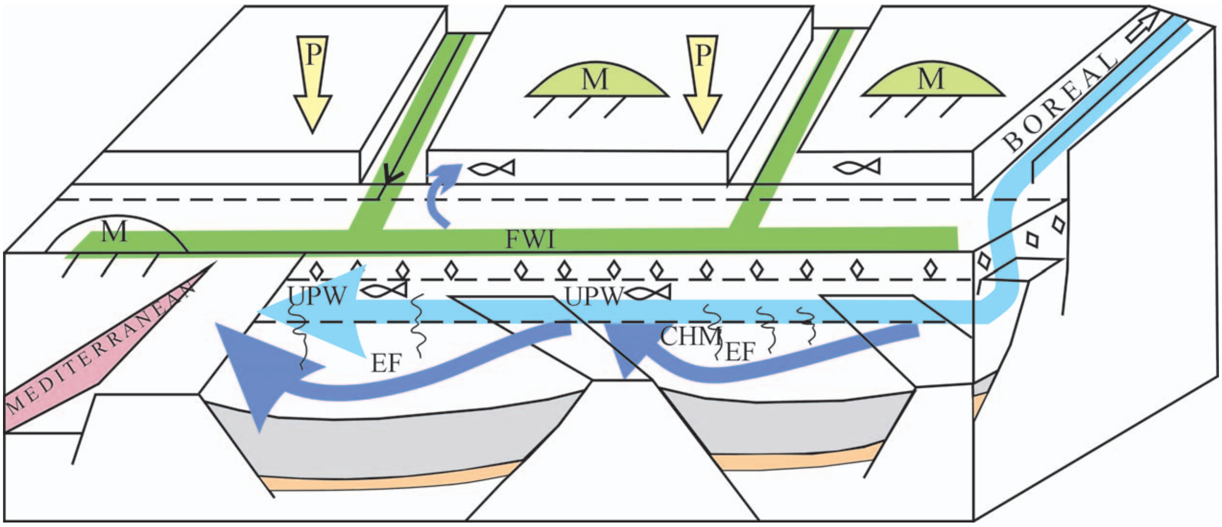

Fig. 17. Lower Oligocene basins with the Black-sea type hydrography, estuarine circulation, inflow of Boreal waters, surface-water overflow (FWI — fresh water influx), closure of the Mediterranean seaway, semi-isolation, water column stratification (CHM — chemocline), deep-water stagnation, bottom-water anoxia, upwelling (UPW), eutrophication (EF), dinoflagellate- and diatom-based productivity, export flux, fish fauna intoxication, precipitation (P), humidity, monsoonal activity (M), etc.

solution increased considerably, leading to non-calcareous deposition of the Menilite Formation.

Carpathian and Hungarian Paleogene basins record a different rate of the Paratethyan isolation or Tethyan communication. While the Buda Basin reveals a full-marine deposition, richness of foraminiferal microfauna (*Tritaxia szaboi* beds, Buda Marls, Kiscell Clays), later termination of low-latitude habitats (e.g. latest nummulites, lepidocyclinids and globoquadrinids), the hiatuses ("infra-Oligocene" regression) and proximity of volcanogenic activity, the Central-Carpathian Paleogene Basin implies restricted environments, impoverishment of foraminiferal microfauna (Huty Formation, Zuberec Formation, Brzegi Member), predominance of oxygen-deficient taxa, sapropelitic deposition and biosilicite productivity (Menilite black shales and cherts), calcareous nannoplankton blooms, manganese beds and tuffaceous admixture, etc. Considering this, the Oxygen Minimum Zone (OMZ) implied the southward weakening of anoxia towards the Tard Sea in the Hungarian Paleogene Basin and its northward widening towards the Carpathian Flysch Sea (Fig. 19). The northern part of the CCPB seems to be more isolated, humid, eutrophicated, oxygen depleted and cooled by Boreal waters. On contratry, the southern part of this basin represents the neritic and near-shore zone of the North Buda Paleogene Basin (Gross 1978;

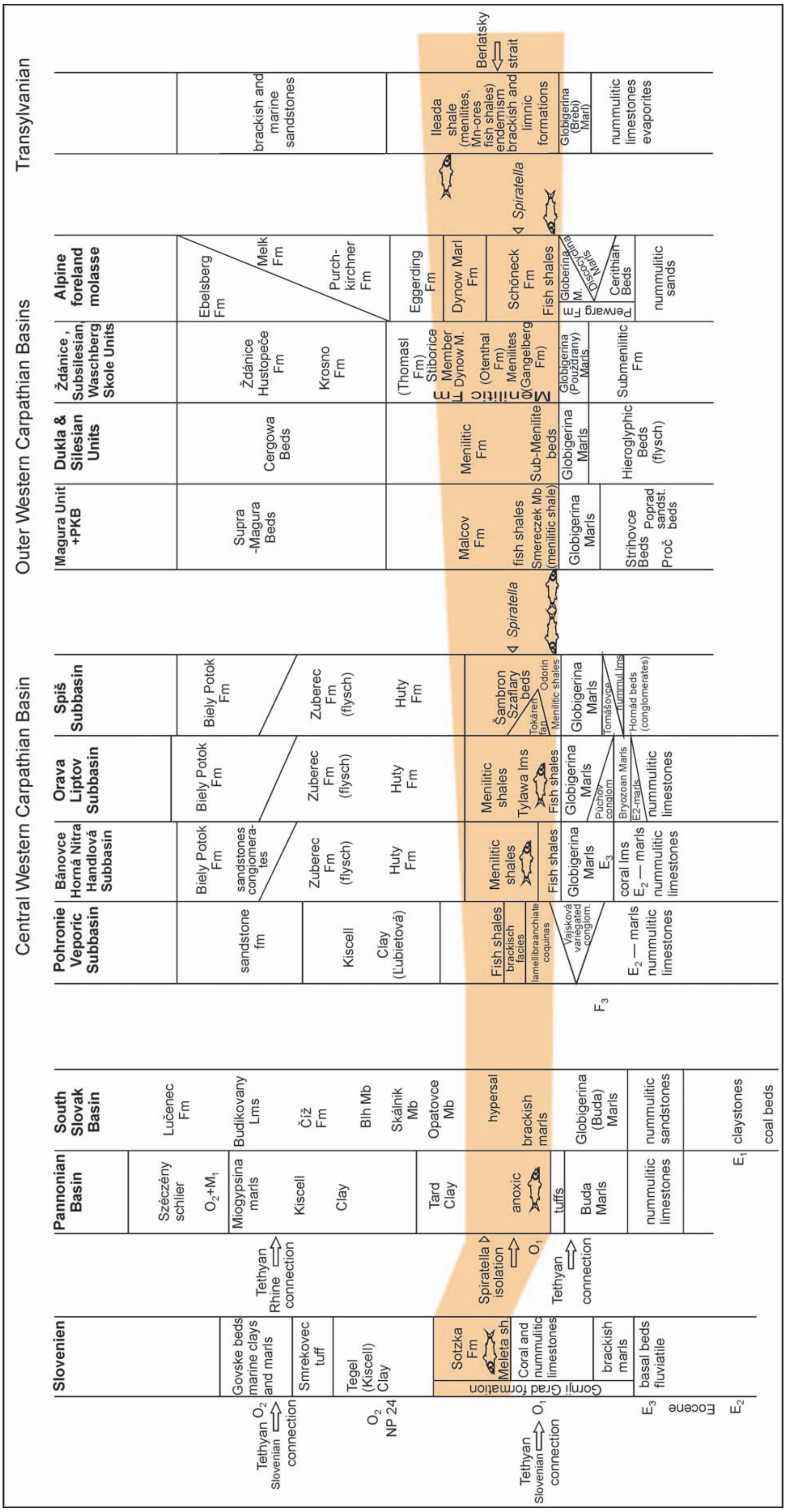

Fig. 18. Distribution of anoxic facies, fish shales, biosilicite horizons, fish shales, laminated limestones, Spiratella horizons, endemic fauna and other indications of the Proto-Paratethyan isolation in the Carpatho-Pannonian basins at the beginning of the Oligocene. Abbreviations: **PKB** — Pieniny Klippen Belt.

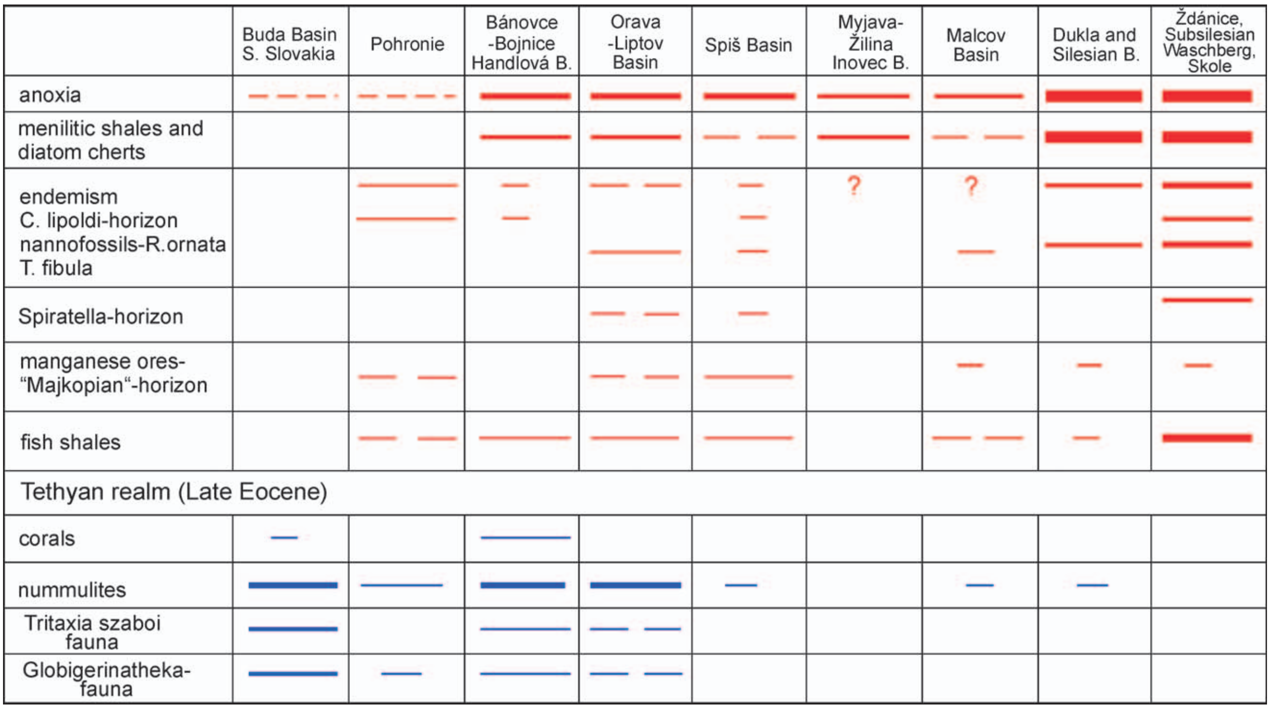

Fig. 19. Tethyan vs. Paratethyan marker facies in the Carpathian Paleogene basins. The figure exhibits northward increase in the intensity of the Paratethyan isolation and southward amplification of the Tethyan influence.

Nagymarosy 1990), influenced by the Tethyan waters. The circulation model of the Paratethyan sea (Dohman 1991; Schulz 2003) presumed a mixture of the Boreal deep water and Tethyan surface water. The restricted seaway connection with the Mediterranean Tethys during the Early Oligocene resulted in fresh water overflow in the Paratethyan basins. The high runoff and separation of the CCPB by the intrabasinal high (Low Tatra Highland), indicates an estuarine-type circulation with an inflow of marine bottom waters and an outflow of fresh surface waters (Fig. 17). Here, the CCPB revealed features of the Black Sea-type basins similar to other basins in the Paratethys Sea (Schulz et al. 2005).

During the Early Oligocene, the oxygen crisis affected the whole Central Paratethys, including the Hungarian Paleogene Basin (Tard Clays), Slovenian Basin (Fish Shale — Tegel Unit), Carpathian Paleogene basins (Menilite Beds) and Austrian foreland basin (Schöneck–Ottenthall Formations), Transylvanian Basin (Ileanda Shale), etc. Nevertheless, some of the Paratethyan basins were reconnected with the Mediterranean Tethys, as is indicated by the Lower Oligocene bioehermal limestones containing nummulitids (e.g. *N. vascus*) in the Slovenian and Hungarian Paleogene Basins (Gornji Grad Beds — Nebelsick et al. 2000; Szépevölgy Limestones — Korpás et al. 1999), but not/or very rarely in Central-Carpathian Paleogene Basin (Buček & Filo 2004). On the other side, the *Spiratella*-rich fauna of the Hungarian Paleogene Basin (Tard Clay) provide evidence of the cold-water influence of the Boreal Sea. This implies that Tethyan-Boreal communication, most likely via the Mid-Hungarian corridor and Slovenian Strait, which is indicated by the mixed mollusc fauna in the Kiscellian Clay (Báldi 1984). In contrast to other basins in the Central Paratethys, the Transylvanian Paleogene Basin was connected rather with the Eastern Paratethys, considering immigrations of mollusc fauna (e.g. *Lenticorbula sokolovi*, *L. helmerseni* — Báldi 1980; Moisescu 1995), evaporitic and lagoonal-lacustrine deposition (Proust & Hosu 1996), oolitic ironstones such as those in the Maikop Basin (Popov & Stolyarov 1996) and even migrant Asian mammals (Baciu & Hartenberger 2001). The affinities of the Transylvanian Paleogene Basin to the Solenovian Sea and Boreal Sea (*Nucula compta* Level) proved its more northern position within the Intra-Carpathian basins.

The Intra-Carpathian system of the Paleogene basins was disturbed during plate-tectonic reorganization of the ALCAPA terranes (Figs. 20, 21). The Slovenian and Hungarian Paleogene basins were accommodated more southerly, and later on they have been shifted to their present position (Csontos et al. 1992). The Hungarian and Central-Carpathian Paleogene Basins are quite different (epicontinental-type basin vs. marginal basin of the Carpathian Flysch Sea), exhibiting no direct paleogeographical connection between them. The northern limit of the Hungarian Paleogene Basin is inferred in the Šahy Antiform, which represents a nearshore zone of the Kiscellian Sea containing sebhka-type facies (Vass 2003). The Veporic borderland was submerged in some places by transversal depressions (Soták et al. 2004). The sediments of the Hungarian Paleogene Basin in the Štúrovo, Lučenec and Rimavská kotlina depressions show a closer proximity to the Mediterranean Sea, revealed by their full-marine facies, tide-influenced deposition, abundance of foraminiferal microfauna, presence of lepidocyclinids and miogypsinids, etc. (Samuel & Vaňová 1967; Vaňová 1975; Sztanó 1995; Holcová-Šutovská 1996;

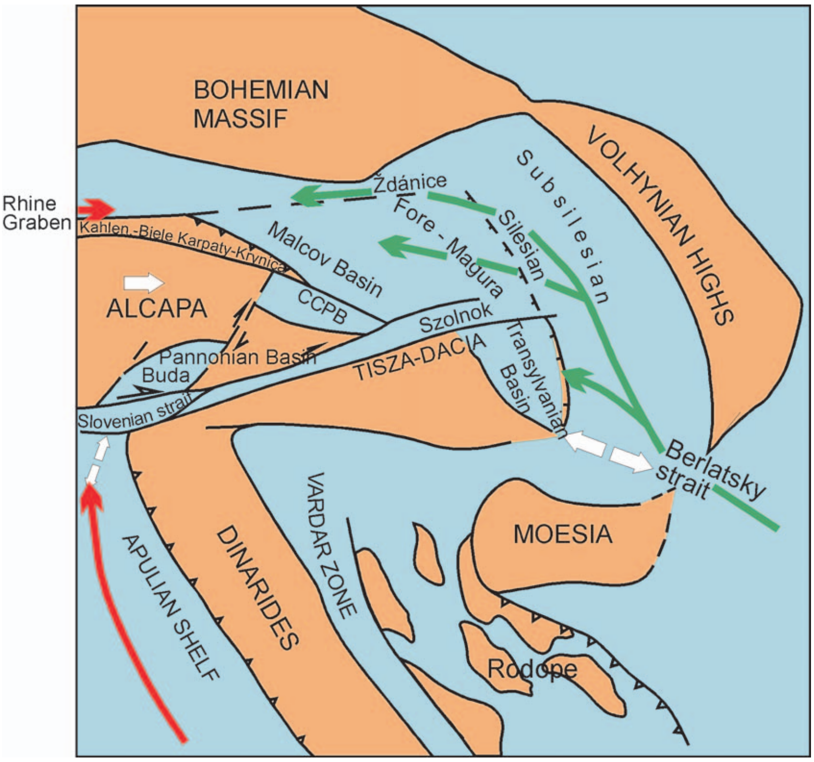

Fig. 20. Palinspastic scheme of the Peri-Tethyan domain in the Late Eocene. Basinal systems show a plate-tectonic reorganization due to the escape tectonics of the ALCAPA Unit. They were full-marine basins with southward connection to the Mediterranean Sea and northward connection to the Boreal Sea.

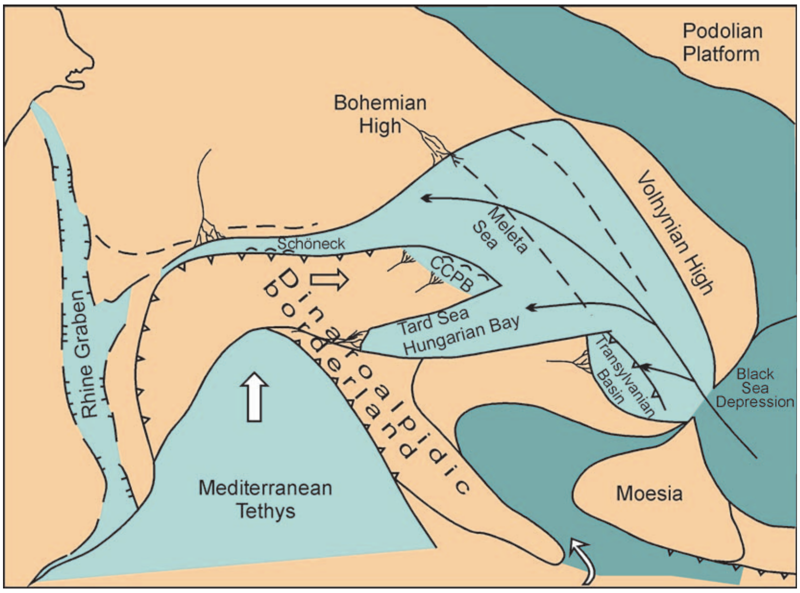

Fig. 21. Palinspastic scheme of the basinal systems of the Proto-Paratethys in the beginning of the Oligocene. Paleogeography was changed in response to the Mesoalpine tectogenesis, which led to the uplift of the Dinaroalpidic borderland and first isolation of the Alpine foreland, Carpatho-Pannonian and Transylvanian basins.

Holcová 2001, etc.). It is most likely that the Hungarian and Central-Carpathian Paleogene Basins came into tectonic juxtaposition due to the NE-directed displacement of the Pelso Unit (Nagymarosy 1990). This unit attained its present position by tectonic and rotational movement from the Late Oligocene to Early Miocene (Vass et al. 1996). This is also the case in the Hungarian and Transylvanian Paleogene basins, which were jointed by large-scale tectonic movement of the ALCAPA and TISIA-DACIA blocks.

Summary

Sedimentary sequences of the Central-Carpathian Paleogene Basin provide evidence of following paleoenvironmental and climatic events related to the Eocene/Oligocene boundary:

EBi-1: growth potential of carbonate platform progressively weakened due to drowning, temperature drop and overfeeding stress of oligotrophic organisms (e.g. nummulitids);

EPi-1: demise of the carbonate platform in response to climatic deterioration and mesotrophication in the Carpathian basins during the Late Eocene;

EPi-2: temperate- to cool-water productivity of Bryozoan Marls and Globigerina Marls, indicating a rise of nutrients from runoff or via upwelling. Plankton productivity of the Globigerina Marls implies the changes in life-mode strategy, diversity, temperature preferences and water-depth habitats;

TEE: peak abundance of bloom-forming globigerinids corresponds to climate-productivity maximum at 33.7 Ma, which reveals the increase of δ^{18}O (cooling) and the decrease of δ^{13}C (CO_2 respired due to metabolism); — the positive shift of oxygen isotopes indicates a decrease in seawater temperature at the Eocene/Oligocene boundary (12–15 °C); — contemporaneous TA 4.4 sea-level fall and increase in carbonate dissolutional rate; — taphrostratigraphic events near 33.7 ± 1.0 Ma;

Oi-1: the major climatic turnover took place in the Early Oligocene; — cooling and humidity led to temperate/wet conditions; — fresh water discharge, water-column stratification, brackish surface water, dinoflagellate-rich productivity, estuarine circulation, oxygen depletion, sulphidic bottom water and seasonal upwelling activity (H_2S intoxication and a high mortality);

Oi-2: eutrophication, expansion of OMZ, photic zone anoxia and stagnant regime of the CCPB during the mid-Early Oligocene; — sapropelitic deposition of menilite shales, fish shales and manganese ores; — episodical blooms of calcareous nannoplankton in the period of runoff stillstand, anti-estuarine circulation and higher carbonate resources (Tylawa-type limestones); — coccolithophorid-rich productivity links with a positive $\delta^{13}C$ excursions, indicating short-lived greenhouse events and/or decrease photosynthetic rate of ^{12}C.

Earliest Paratethyan stage: the first isolation event in the NP23 Biozone, the appearance of Paratethys-endemic nannofossils, diatom-based productivity and biosiliceous deposition; differentiation of basin-floor topography and paleogeographic communication; — plate-tectonic reorganisation of the intra-Carpathian basins during the Alpine orogenesis.

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