

Paleocene-Eocene boundary events in the northeastern Peri-Tethys

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ABSTRACT

In the northeastern Peri-Tethys epicontinental basin stretching from the Black Sea to Tadjikistan, a sapropelitic bed with total organic carbon (TOC) contents up to 20% and enriched in redox-sensitive elements was deposited during the Initial Eocene Thermal Maximum (IETM), recognized by the occurrence of a negative carbon isotope excursion in organic carbon and carbonate. In the sapropelitic bed, calcareous plankton decreased in abundance, a benthic foraminiferal extinction occurred, dinocysts were abundant, and there were abnormally thick-walled dinocysts and asymmetric calcareous nannoplankton species. We argue that the sapropelitic bed accumulated as a result of high productivity, triggered by an influx of nutrients that reached the basin when organic- and P-rich sediments in low-lying coastal areas were flooded during a rapid transgression. High productivity led to oxygen depletion, decimating benthic foraminiferal assemblages. Organic biomarkers suggest that at least in some places even waters in the photic zone became anoxic. The dysoxia/anoxia might have been exacerbated by stratification, resulting from high runoff in the humid western region, high evaporation in the arid eastern area. Anoxia/dysoxia ended gradually. We suggest that deposition of large amounts of organic matter in the northeastern Peri-Tethys may have been a factor in terminating the IETM.

INTRODUCTION

During the last decade many scientists have investigated events that occurred during the latest Paleocene and earliest Eocene (e.g., Kennett and Stott, 1991; Zachos et al., 1993; Dickens et al., 1995; Thomas and Shackleton, 1996; Schinitz et al., 1996; Aubry et al., 1998; Koch et al., 1992; Wing et al., 1995; Katz et al., 1999). The most prominent event during this period is the Initial Eocene Thermal Maximum (IETM), which has been documented in sections deposited in such different environments as the open ocean, marginal basins, and on the continents. During this extreme warm period marine and continental biota show rapid change, coeval with distinct negative excursions

in oxygen and carbon isotopic values in carbonate and organic matter.

In some areas, sediments deposited during the IETM are enriched in organic matter (e.g., Gavrilov et al., 1997; Speijer et al., 1997). One of these areas is the northeastern Peri-Tethys, where organic-rich sediments can be traced over >2500 km, from the Caucasus to Central Asia (Muzylov et al., 1989; Muzylov and Shcherbinina, 1995). The properties of the organic-rich sediments, called the sapropelitic bed, are relatively constant over this large area, although they accumulated in different environments in different parts of the region.

The origin of the sapropelitic bed has been discussed extensively, after Muzylov et al. (1989) first documented that its

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deposition was synchronous over a large area (e.g., Gavrilov and Muzylov, 1991; Muzylov, 1994; Muzylov et al., 1996; Gavrilov et al., 1997; Stupin and Muzylov, 2001), but no comprehensive discussion of its deposition has been presented. We describe the sedimentological, geochemical, and paleoecological changes during the IETM in the northeastern Peri-Tethys, and propose a scenario for the accumulation of the sapropelitic bed.

METHODS

The mineralogy and geochemistry of the sapropelitic bed and its surrounding sediments were studied in many sections (Fig. 1), six of which are described in this paper, while two more are mentioned. Clay minerals were analyzed using X-ray diffraction. A range of elements and compounds were analyzed in the Laboratory of Analytical Chemistry at the Geological Institute RAS, Moscow: total organic carbon (TOC), total carbonate, S, P, Fe, Ti (wet chemistry), V, Cr, Ni, Co, Pb, Cu, Ga, Ge, Mo, Ag, Zn, Sn (quantitative spectral analysis using an atomic emission spectrometer), Sc, Se, Br, Sb, Cs, Hf, Au, Th, Y, La, Ce, Nd, Sm, Eu, Tb, Yb, Lu (instrumental neutron activation analysis, INAA), and Nb, Zr, Rb, Sr, and Ba (X-ray fluorescence analysis). In two sections (Kheu in the northern Caucasus and Kurpai in Tadjikistan), the organic matter was examined by express pyrolytic analysis using Rock-Eval II, chromatography, molecular mass-spectrometry, and IR-spectrometry.

Nannofossils were studied in the bulk sediment using light microscopy. We counted the relative abundance of major nannofossil taxa. In most samples, 300 specimens were counted, but

in a few samples with a low abundance of nannofossils 100 specimens only could be counted. In some intervals, there were so few specimens of nannofossils (<100 specimens per sample) that no statistically significant counts could be made.

RESULTS

We present results from six most important sections in the northeastern Peri-Tethys, covering a large area in the central and eastern Caucasus, the western shores of the Aral Sea, the western parts of the Kopetdag Mountains, and the Tadjik depression (Fig. 1). Generally, the sapropelitic bed is dark in color and fissile on weathering (Figs. 2, 3). The lower boundary is always sharp, and in some sections erosional, whereas the upper boundary is more gradual. In most sections the sapropelitic bed can be subdivided into three units, with a more calcareous middle unit that is less rich in TOC between lower and upper units with high TOC. Toward the northern margin of the basin, TOC content decreases significantly in the sapropelitic bed. Locally, sediments coeval to the bed contain abundant fish detritus, and in some localities in Central Asia (e.g., the Kyzyl Kum Desert, Tadjik depression) the bed is either associated with or replaced by phosphorite-bearing deposits.

Lithology and paleontology

In the Kheu River section (Central Caucasus) the sapropelitic bed is ~0.5 m thick, and exposed along the riverbank near the village of Gerpegezh (20 km southeast of the city of Nalchik)

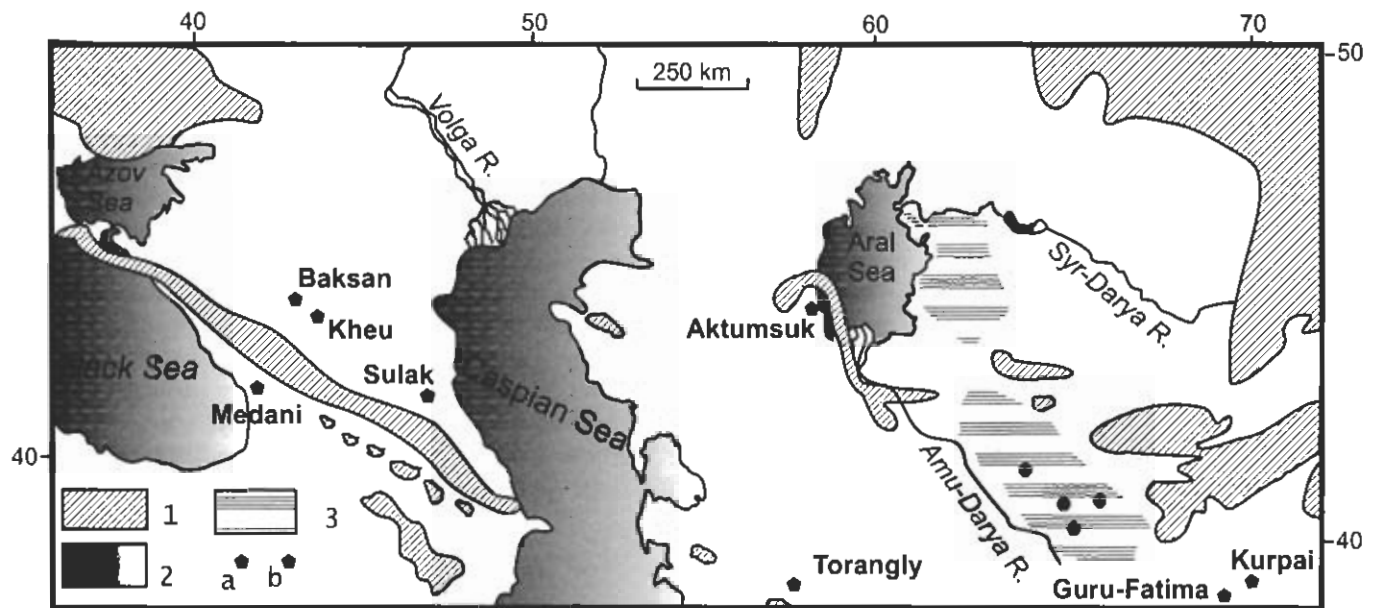


Figure 1. Studied localities in the northeastern Peri-Tethys epicontinental basin. 1—land, 2—sea, 3—area of oil shale distribution, a—localities of the studied sections, b—localities of studied boreholes.

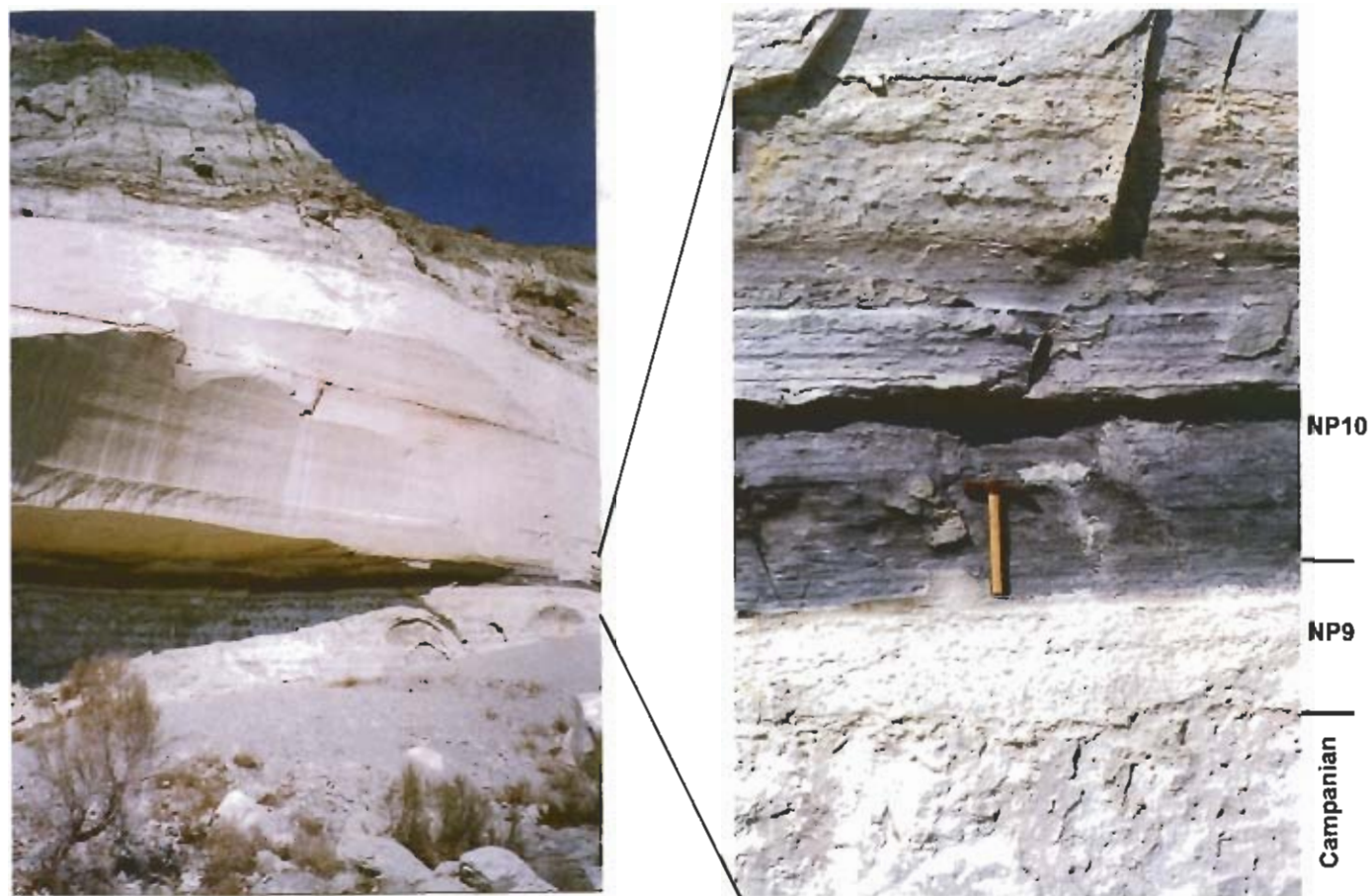


Figure 2. Saproelitic bed in the Paleocene-Eocene boundary sediments of the western Aral Sea region (Aktumsuk Cape).

within bluish-gray calcareous clays. The bed is dark brown (Plate 1, Figures 1–6) and contains abundant fish remains, ostracods, sulfide nodules, and rare pyritized wood fragments. It is underlain by a noncalcareous light brown clay (0.1 m) with scarce nannofossils and no foraminifera (Gavrilov et al., 2000). Within this layer are the first occurrences (FOs) of the calcareous nannofossils *Transversopontis pulcher* and *Campylosphaera eodela* (Fig. 4A, Appendix A). The FO of the latter species marks the bottom of Subzone CP8b in pelagic sections (Okada and Bukry, 1980).

Within the saproelitic bed there is a poorly preserved and low-diversity nannoflora, dominated by various species of *Toweius*. Compared to the underlying clays, the planktic/benthic foraminiferal ratio in the bed is high (Stupin and Muzylov, 2001). Nannofossils are rare and foraminifera are absent in the lowermost 5 cm of the saproelitic bed where the TOC is highest. Slightly higher in the bed, *Pontosphaera* and the calcareous dinocyst *Thoracosphaera* become significant components of the assemblage. In the more calcareous middle part of the bed, the nannofossil content and diversity increase, and the first, rare rhomboasters (*R. bramlettei* s.l.), occur. In the upper, TOC-rich

part of the bed, rhomboasters and discoasters are abundant (including the FO of *Discoaster araneus*). The abundance of benthic foraminifera, mainly calcareous species, is low (Stupin and Muzylov, 2001).

In the Sulak River section (Dagestan, eastern Caucasus, near the Chirkey Hydroelectric Power Station), the saproelitic bed is charcoal-black and occurs intercalated in a series of alternating dark red and green marls (Fig. 3). It is 0.3 m thick, with the lowermost 0.1 m slightly disturbed, possibly as the result of minor slumping. The nannofossil assemblage has a high relative abundance of *Fasciculithus* and *Discoaster* species (Fig. 5A, Appendix B). In some intervals, we observed abundant calcareous detritus and reworked Cretaceous nannofossils.

The underlying homogeneous green sediments are calcareous in their lowermost part, with abundant nannofossils. Toward the top of these sediments the carbonate content decreases and nannofossils disappear. In the lowermost part of the saproelitic bed (0.20 m), there are almost no nannofossils. In the uppermost part, various species of *Toweius* dominate, and rhomboasters, pontosphaerids, *Discoaster araneus* and asymmetric *D. arantios* first appear. The total number of *Thoracosphaera* specimens

Enlarged area



NP10



NP9



Figure 3. Spropelitic bed in the Paleocene-Eocene boundary sediments of Dagestan (Eastern Caucasus, Sulak River).

increases up-section within the sapropelitic bed, and *Campylosphaera eodela* first occurs above the bed. A sharp increase in the abundance of calcareous detritus at the base of the overlying sediments, the reduced thickness of the bed, and the lack of a threefold division suggest that erosion removed the uppermost part of the bed.

In the Aktumsuk Cape section (western Aral Sea, Uzbekistan, at the base of the cliff of the Usturt Plateau), the sapropelitic bed is 1 m thick, dark gray, and intercalated in white limestones (Fig. 2). It is finely laminated and shows significant fluctuations in TOC. Sharp negative excursions in both $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$ are marked in the bed (Bolle et al., 2000). The underlying thin limestone bed, which has been assigned to calcareous nannofossil zone NP9 (Martini, 1971), overlies upper Campanian sediments with a sharp erosional contact (Fig. 4B, Appendix C). The high carbonate concentration is the result of a high content of calcareous detritus (Plate 1, Figures 7, 8), and nannofossils are less common than in the sapropelitic bed. Various pontosphaerids and the first species of rhabdosphaerids (*Rhabdosphaera cf. sola*) occur. As in most other sections, TOC is most enriched in the lowermost part (0.40 m) of the sapropelitic bed where the beginning of carbon isotope excursion and FO of *R. bramlettei* were established. Unique for the studied region, species of *Coccolithus* become more abundant in this section, while species of *Toweius* are rare, as was observed in other regions near the Paleocene-Eocene transition (Tantawy et al., 2000; Monechi et al., 2000).

TOC is lower and the nannofossil abundance increases in the calcareous middle part of the sapropelitic bed, but the nannofossil assemblage does not change significantly. Nannofossils and planktonic foraminifera become more abundant in the organic-rich uppermost part of the sapropelitic bed. The abundance of *Thoracosphaera* and calcareous debris increases sharply in the overlying sediments. The sapropelitic bed has common benthic foraminifera, equally divided between agglutinated and calcareous taxa (Stupin and Muzylov, 2001), although the latter are more common in its middle part.

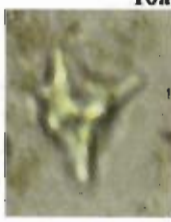
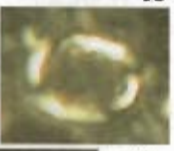
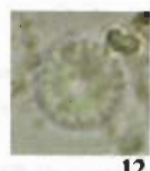
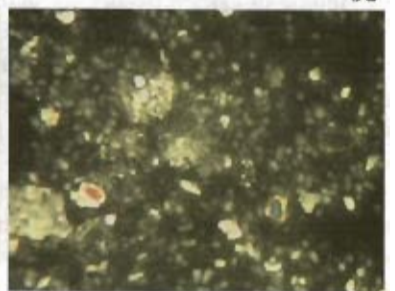
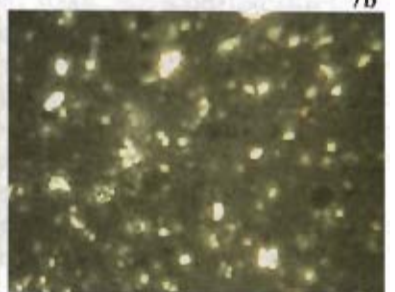
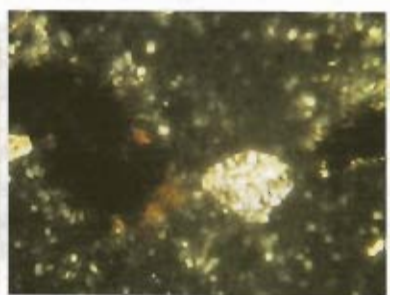
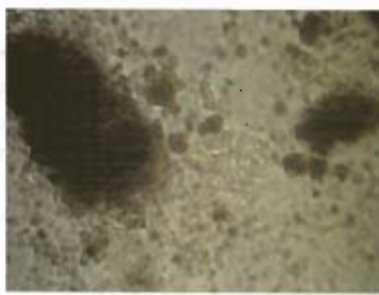
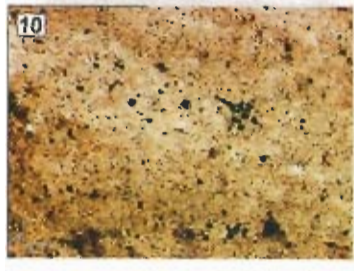
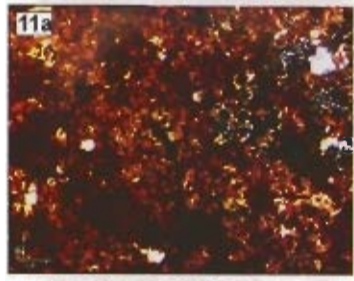
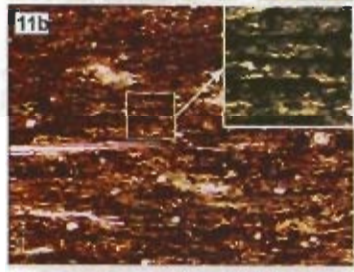
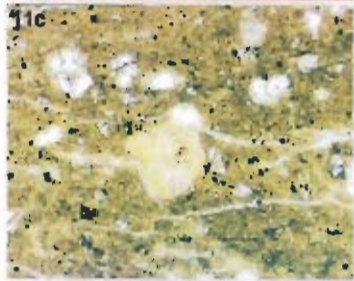
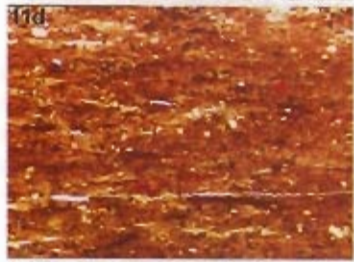
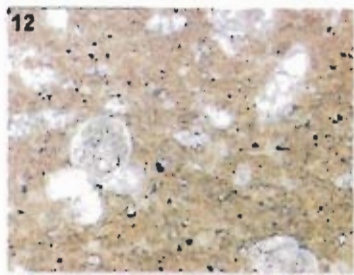
In the Torangly section (western Kopetdag Mountains, Turkmenistan, exposed in the northern slope of the Kurendag anticline), the SB is 1.2 m thick and consists of brown clays, rich in fish scales, intercalated in gray, slightly calcareous clays (Fig. 5B, Appendix D). TOC is relatively low, possibly because the sampled sediment was rather weathered. In this section, the first rhombosphaerids occur in sediments just below the sapropelitic bed, which are not enriched in TOC but have increased concentrations of some elements (e.g., Mo, Zn) characteristic of the sapropelitic bed in other sections (Fig. 7). We suggest that there was an erosional event at the time of deposition of the lowermost part of the sapropelitic bed, so that only the middle and upper parts of the bed are present, as is also the case in the Baksan River section (Muzylov et al., 1996). The abundance of *Fasciculithus*, which is believed to be an oligotrophic taxon, sharply increases in the sapropelitic bed, while planktonic foraminifera are absent.

C. eodela first occurs in the overlying sediments. In the upper part of the bed, nannofossils are very rare, and calcareous species dominate the benthic foraminiferal assemblage (Stupin and Muzylov, 2001).

In the Kurpai section (Tadjik depression, Tadjikistan; Figure 6A), the sapropelitic bed is 1.3 m thick and consists of black, laminated clays with rather poorly preserved and low-diversity nannofossil assemblages dominated by *Fasciculithus* and *Discoaster* species. It is intercalated in slightly calcareous sediments with a significant fraction of terrigenous silty material. In the underlying noncalcareous layers nannofossils are absent and tiny calcareous benthic foraminifera are present. In the lowermost part of the sapropelitic bed, the CaCO_3 content increases, mainly due to the presence of calcareous detritus and, to a minor extent, calcareous benthic foraminifera, but nannofossils are absent. Within the light-colored intercalation nannofossils are present, and *Fasciculithus* is abundant (>50%) together with *Coccolithus*. The uppermost part of the sapropelitic bed is an oil shale with very high TOC, and nannoplankton and foraminifera are absent. In the sediments overlying the bed, nannofossils including pontosphaerids are rare and calcareous detritus is common.

In the Guru-Fatima section (Tadjik depression) the sapropelitic bed is a 1.3 m thick, black oil shale with TOC >15%, intercalated in low-carbonate gray clays (Fig. 6B). There is a prominent erosional surface at the NP8/NP9 boundary below the bed. In the lowermost part of the bed, where TOC is highest, nannofossils are rare and their diversity is low. *Braarudosphaera bigelowii* (57%), a long-ranging nannoplankton species (Lower Cretaceous-Quaternary) which is abundant in sporadic blooms, is dominant. Calcareous detritus is abundant. Nannoplankton are not common but diverse (though species of *Toweius* dominate the assemblage) higher in the sapropelitic bed, despite the low CaCO_3 content, and oxic benthic foraminifera are common (Stupin and Muzylov, 2001). Up-section, TOC decreases gradually, and calcareous nannofossils become rare or are absent. In the uppermost part of the bed, rare nannofossils occur, and the abundance of *Toweius* sharply decreases while various *Fasciculithus* and *Discoaster* species become more abundant. The overlying sediments contain rare various nannofossils and planktic foraminifera only.

In the Kheu river (N.I. Zaporozhets, 2002, personal commun.), Medani (Transcaucasia, Georgia) and Torangly (Akhmetiev and Zaporozhets, 1996) sections, organic-walled dinocysts show good preservation without reworking within the sapropelitic bed and surrounding sediments. Scarce *Apectodinium* spp. are found in the sediments surrounding the bed. Within the bed, however, *Apectodinium homomorphum*, *A. augustum*, *A. parvum*, *A. quinquelatum*, *A. hyperacanthum*, *A. paniculatum*, and *A. sumissum*, are abundant, and co-occur with *Kenleyia*, *Rhombodinium*, *Hystriochokolpoma*, and *Fibrocysta*. This assemblage was assigned to the dinocyst *Apectodinium hyperacanthum* zone (Powell, 1992). The sapropelitic bed thus corresponds to the *Apectodinium* acme observed in



many parts of the world during the IETM (e.g., Bujak and Brinkhuis, 1998; Crouch et al., 2000; Steurbaut et al., 2000; Crouch et al., 2001).

The sapropelitic bed contains unusual forms of the genus *Apectodinium* with extremely thick-walled peri-cysts, which constitute up to 60% of the dinocyst assemblage. These forms may be endemic to the northeastern Peri-Tethys and indicate eutrophic conditions (Akhmetiev and Zaporozhets, 1996). Organic-walled plankton remains are abundant in the bed, but their total amount is only a part (25% in Kheu river section; 65% in Medani section) of the bulk amorphous matrix, which consists of solid organic matter and other algal remains.

Mineralogical and geochemical characteristics of the sapropelitic bed

We analyzed the clay fraction in samples from all sections. Clay mineral assemblages consist of varying proportions of smectite, illite, mixed-layered illite-smectite, chlorite, and rarely kaolinite. In most cases there is no significant difference between the clay mineral assemblages in the sapropelitic bed and those in the surrounding sediments. Kaolinite is absent in most sections, and found only in the Aktumsuk section (northwestern Uzbekistan) and, to a lesser extent, in the Medani section (Western Georgia). In these sections, the sapropelitic bed and surrounding sediments contain kaolinite, but it is slightly more abundant in the bed.

The oil shale collected near the Amu-Daria River (Fig. 1) contains an admixture of palygorskite, which is a common mineral in the Paleocene-lower Eocene sediments along the eastern margin of the basin (Fergana embayment, eastern Uzbekistan), indicating that the climate was arid in these regions. In the Torangly (Turkmenistan) and Sulak River (Dagestan) sections Paleocene-Eocene sediments contain illite with a low percentage of expanding interlayer clays (5–10%) and chlorite, as a result of deep burial.

The TOC in the sapropelitic bed is highly variable (Fig. 7). In the Amu Darya region (Suzak oil shale zone; Figure 1), the TOC content is higher than 20%, but in the Lower Volga region

it does not exceed 1%. In the areas of predominantly carbonate sedimentation (Caucasus and Aral Sea region), where the sapropelitic bed is organic-rich, the TOC is usually negatively correlated with the CaCO₃ content, but this correlation does not hold in the regions of essentially terrigenous sedimentation (Central Asia).

Fe, P, S are enriched in the sapropelitic bed whereas Mn is slightly depleted. The minor elements can be divided into two groups. V, Cu, Mo, Ni, Ag, Au, Se, Co, Zn, Cr, Sn, Ga, Ge, Br, Re, Te, Tl, Bi, Sb, Cd are generally enriched to a variable extent in the bed. Some of these elements show enrichment throughout the area, others in some sections only. Ti, Zr, Nb, Rb, Sr, Sc, Cs, Hf, and the REE have a very low concentration in the sapropelitic bed (Gavrilov et al., 1997). The pattern of elemental enrichment in the bed is similar to that in the Quaternary sapropels of the Black Sea and Mediterranean (e.g., Fomina and Volkov, 1970; Volkov, 1973; Calvert, 1990; Pruyssers et al., 1991; Sutherland et al., 1984).

In the Kheu section, the organic matter in the sapropelitic bed is enriched in ¹²C ($\delta^{13}\text{C}$ –30.3‰ to –30.8‰) relative to the organic matter in the surrounding sediment (–27‰ to –28‰; Figure 8). The carbonates show the same isotopic excursion (Kodina et al., 1995; Gavrilov et al., 1997), as also reported for the Aktumsuk section (Bolle et al., 2000).

Organic geochemical analyses show a strong increase in the abundance of marine, planktonic organic matter in the sapropelitic bed relative to the surrounding sediments, but the sources of the organic matter vary by location (Gavrilov et al., 1997). In the Kheu River section, organic matter is mainly derived from marine plankton, but there is a minor amount of terrestrial organic material. Terrestrial organic matter is absent in the Kurpai section. This variability in the composition of organic matter may reflect the regional variability in setting, run off, and climate. Climate was more humid in the western region, and a terrestrial organic matter was derived from the vegetated coastal areas, whereas the eastern regions were much more arid, and there was minimal input from vegetation.

Biomarkers derived from the anaerobic photosynthetic green sulphur bacteria *Chlorobium* (Summons and Powell,

Plate 1. Figures 1–6. Thin sections of different layers of sapropelitic bed and surrounding sediments in the Kheu river section. Numbers in the top left corner correspond to numbers of samples in Figure 7: (1) "presapropelitic" clay (10 cm); (2) lowermost sapropelitic layer (5 cm) rich in organic matter in the major extent; (3) thinly laminated organic-rich layer (20 cm); (4) calcareous interlayer (10 cm); (5) uppermost organic-rich layer (15 cm); (6) overlying marl. Figures 7–9. Smear slides of sapropelitic bed and underlying sediments of Aktumsuk section: (7–8) underlying sediments; (9) uppermost part of sapropelitic bed, sample 58 (similar states of discoaster preservation and a high concentration of calcareous detritus can be seen in all these samples). 7: *Thoracosphaera operculata*, ×400; a—transmitted light, b—cross-polarized light. 8: *Discoaster multiradiatus*, ×400; a—transmitted light, b—cross-polarized light. 9: *Discoaster multiradiatus*, ×400; a—transmitted light, b—cross-polarized light. Figures 10–18. Selected nannofossil species in the sapropelitic bed of the Kheu River section: (10) *Rhomboaster bramlettei*, ×2000, sample 11a, transmitted light; (11) *Rhomboaster bramlettei*, ×2000, sample 11b, transmitted light; (12) *Discoaster lenticularis*, ×2000, sample 11b, transmitted light; (13) *Fasciculithus thomasi*, ×2000, sample 11b, cross-polarized light; (14) *Rhomboaster bramlettei*, ×2000, sample 11a, a—transmitted light, b—cross-polarized light; (15) *Neochiastozygus concinnus*, ×2000, sample 11b, a—transmitted light, b—cross-polarized light; (16) *Rhomboaster bramlettei*, ×2000, sample 11b, a—transmitted light, b—cross-polarized light; (17) *Fasciculithus involutus*, *Toweius callosus*, ×2000, sample 11b, cross-polarized light, (18) *Placozygus signoides*, ×2000, sample 11b, a—transmitted light, b—cross-polarized light.

1986, 1987) were detected in the organic matter of the sapropelitic bed in the Kurpai section (Kodina et al., 1995). These biomarkers indicate anaerobic conditions and the presence of H_2S in the photic zone of the water column, as in the present Black Sea (Sinninghe Damste et al., 1993; Repeta et al., 1989; Repeta and Simpson, 1991). Organic geochemical data thus show that relative to the surrounding sediments the organic matter in the sapropelitic bed is more dominated by marine planktonic organic matter, and at least in some parts of the basin waters were truly anoxic in the photic zone, and H_2S was present.

Biostratigraphy and environments (planktonic and benthic microfossils)

Recent studies (Muzylöv, 1994; Muzylöv et al., 1989, 1996; Muzylöv and Shcherbinina, 1995) documented that the sapropelitic bed corresponds in age to the boundary between late Paleocene Subzones CP8a/CP8b of Okada and Bukry (1980), based on the FO of *Rhombaster cuspis* within the sapropelitic bed. According to Martini (1971), the FO of *R. bramlettei* (base of Zone NP10) marks the Paleocene-Eocene boundary. Recently, however, the taxonomy of rhombaster has been revised, with the proposition to establish two subspecies of *Rhombaster bramlettei*, *R. bramlettei bramlettei* and *R. bramlettei cuspis* (Bybell and Self-Trail, 1994, 1997; von Salis et al., 2000). In this revision, the base of Zone NP10 is close to that boundary between Subzones CP8a/CP8b. In addition, the base of the global Carbon Isotope Excursion has now been proposed as the Paleocene-Eocene (P-E) boundary by the International Subcommission on Paleogene Stratigraphy (Luterbacher et al., 2000). If we follow this revision, the P-E boundary falls at the base of the sapropelitic bed. We do not have isotope data in all studied sections, and in the figures for this paper have still placed the P-E boundary at the NP9/NP10 zonal boundary.

Consistent biotic changes in our sections include a major change from mainly calcareous-walled phytoplankton in sediments below the sapropelitic bed to mainly organic-walled plankton remains within the bed, separated in some sections (Kheu, Sulak, Aktumsuk, and Kurpai) by a thin layer of sediment just below the bed which lacks both planktic and benthic calcareous microfossils. Some nannofossil taxa (pontosphaerids and rhabdosphaerids) which are common in epicontinental sediments in subsequent epochs, evolved at this time as seen in some sections (Figs. 4, 5, and Muzylöv, 1994).

The sapropelitic bed itself contains both nannofossils and planktic foraminifers, but the nannofossil assemblages within

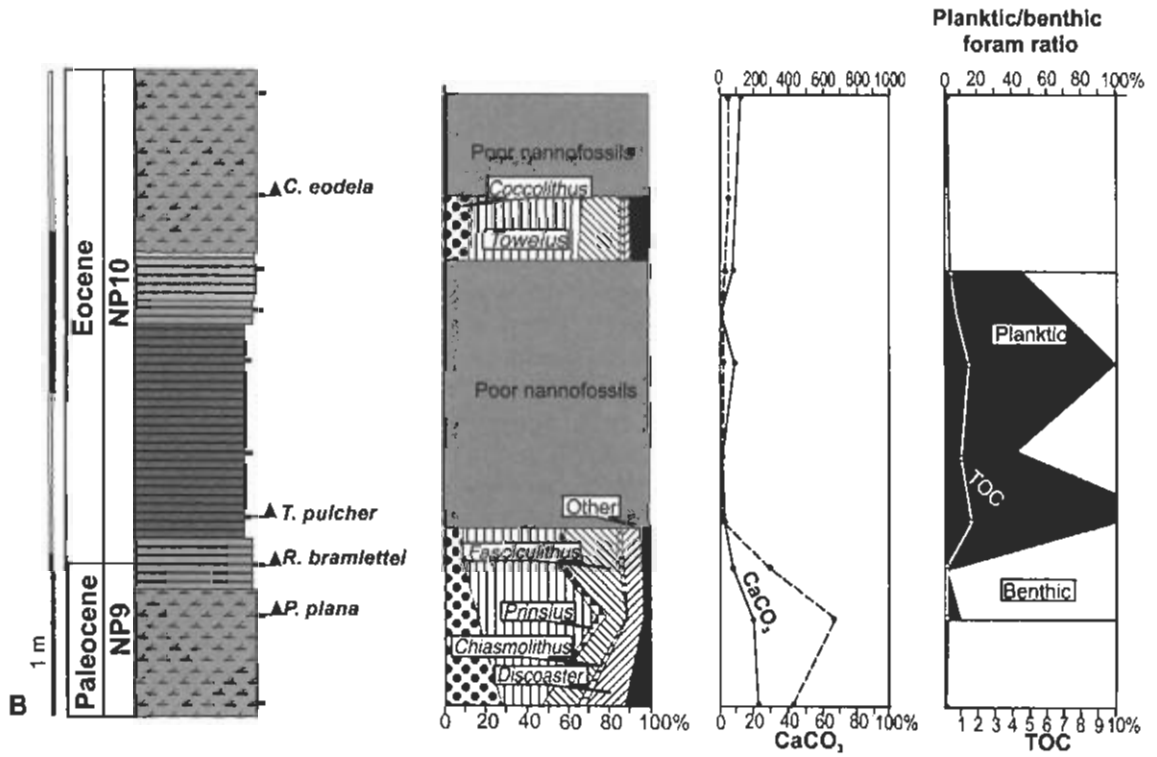
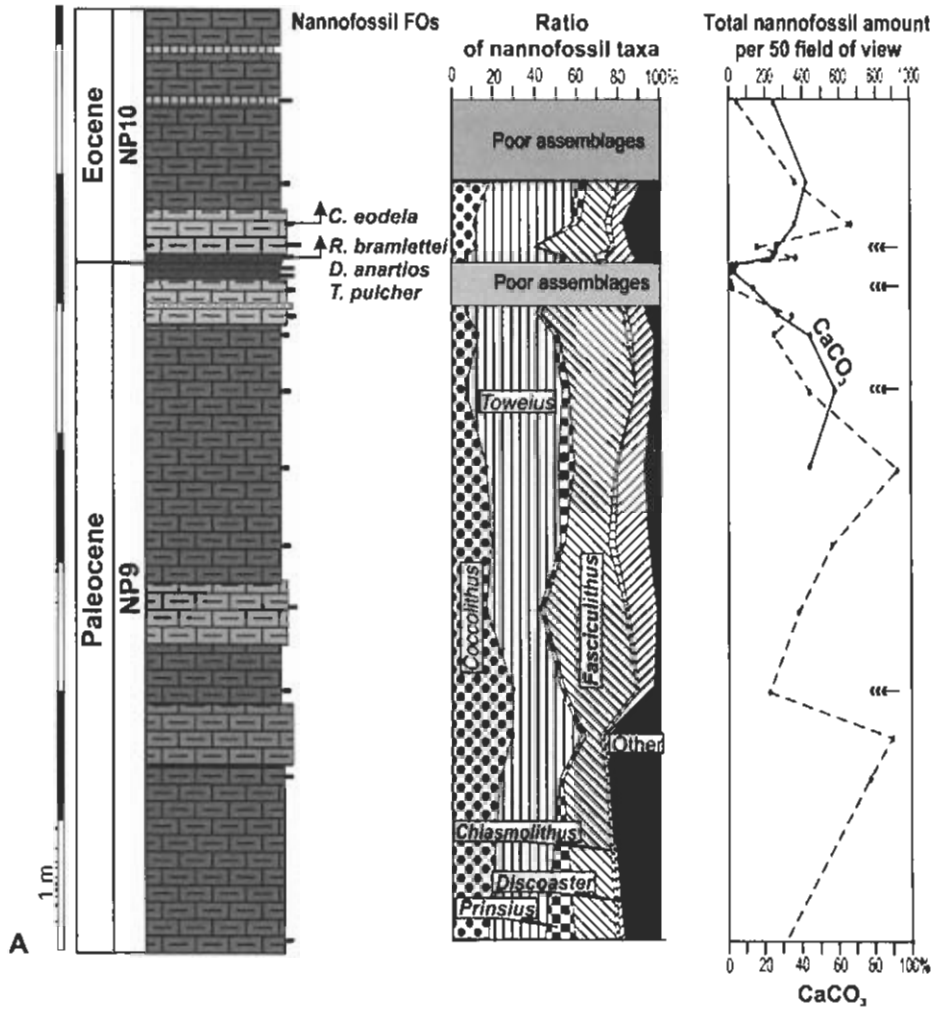
the bed differ from those in the surrounding sediments. Generally, within the sapropelitic bed the nannofossil diversity is low, and one or two taxa dominate. Sections with carbonate sediments (Kheu, Aktumsuk) show a strong dominance of *Toweius*, and those with mostly terrigenous sediments (Torangly, Kurpai, and Guru-Fatima sections) contain assemblages dominated by *Fasciculithus*, probably reflecting shallow water depth and oligotrophic environments.

The nannofloral changes in the northeastern Peri-Tethys sections thus differ from those in sections in Spain (Trabakua Pass section), where the IETM occurs in a dissolution horizon (Orue-Etxebarria et al., 1996), and is marked by a low abundance of the cool water species of the genera *Chiasmolithus* and *Prinsius* and a high abundance of *Fasciculithus*. An abrupt decline in abundance of *Fasciculithus* and the appearance of abundant *Zygrhablithus* above the base of the IETM have been observed in epicontinental and oceanic sections (Aubry et al., 1996; Orue-Etxebarria et al., 1996; Monechi et al., 2000; Bralower and Premoli-Silva, 2002). In our sections, *Zygrhablithus* is generally absent and *Fasciculithus* are common, showing decreased abundance above the sapropelitic bed in the Guru Fatima section only.

In most sections, the lowermost part of the sapropelitic bed which contains the highest TOC content is characterized by a poor nannoflora, and an increased abundance of *Thoracosphaera*. A bloom of *Braarudosphaera* is found only in the Guru-Fatima section, situated in the arid eastern part of the basin. These two taxa are abundant after the Cretaceous-Paleogene boundary, when other nannofossil taxa were rare. Occasional *Braarudosphaera* blooms have been recognized in mid-Oligocene sediments from the South Atlantic, possibly due to voluminous freshwater influx resulting from the melting of Antarctic ice. In Miocene and Pliocene sediments in the Mediterranean and Black Sea, *Braarudosphaera*-rich sediments are associated with sapropelitic layers (L.A. Golovina, 2001, personal commun.). Bukry (1974) concluded that *B. bigelowii* may thrive at relatively low salinities, but Perch-Nielsen (1977) speculated that the taxon is euryhaline, and thus more common where salinity fluctuations limit the occurrence of other species. Experimental data on modern *Thoracosphaera heimii* in culture show a maximum growth rate and productivity at 27 °C in a nutrient-rich medium (Karwath et al., 2000), suggesting that the sapropelitic bed accumulated at relatively high temperatures, in agreement with the occurrence of abundant warm-water discoasters within the bed. Besides, this species has its maximal productivity during periods of stratification of the upper water column and a well-developed thermocline (Vink et al., 2002).

The microfossil assemblages thus indicate that conditions were warm during the deposition of the sapropelitic bed, and locally salinity variations occurred. In addition, the occurrence of some nannofossil forms such as rhombasters, asymmetric discoasters, morphotypes of *Fasciculithus* with a very high curved cone, and extremely thick-walled *Neochiastozygus* sp. testify to the occurrence of unusual environmental circumstances.

Figure 4. Lithological and microfossil changes near the Paleocene-Eocene boundary. A: Kheu River section (Central Caucasus). B: Aktumsuk Cape section (western Aral Sea). Arrows indicate levels of enrichment in calcitic detritus. Data on foraminifera after Stupin and Muzylöv, 2001. Note that the Paleocene-Eocene boundary was drawn at the NP9/NP10 zonal boundary.



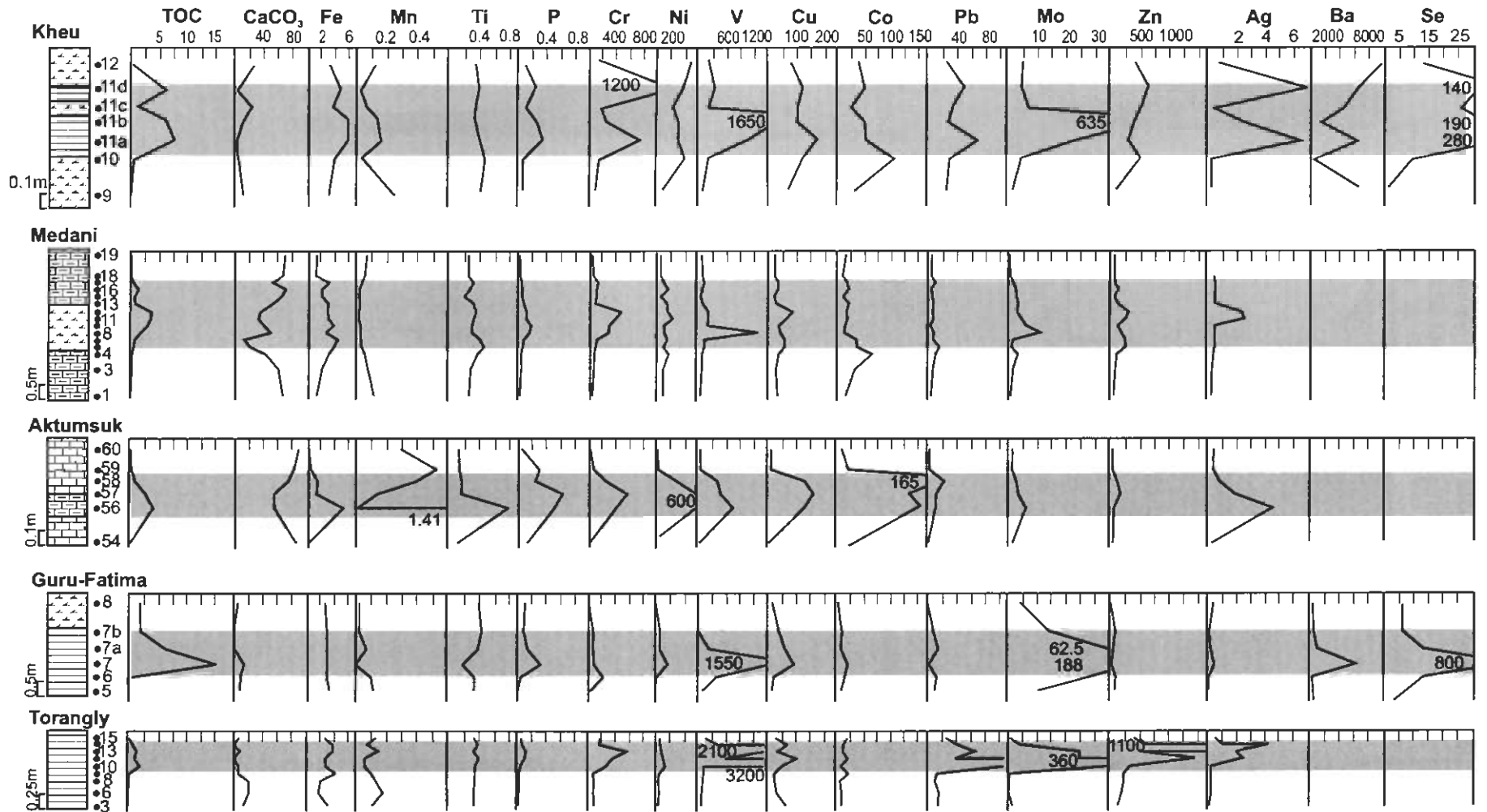


Figure 7. Concentrations of various compounds and elements in the sapropelitic bed in selected sections. The concentrations of C_{org}, Ca CO₃, P, Ti, Fe, Mn, are given in percent, those of the other elements in ppm.

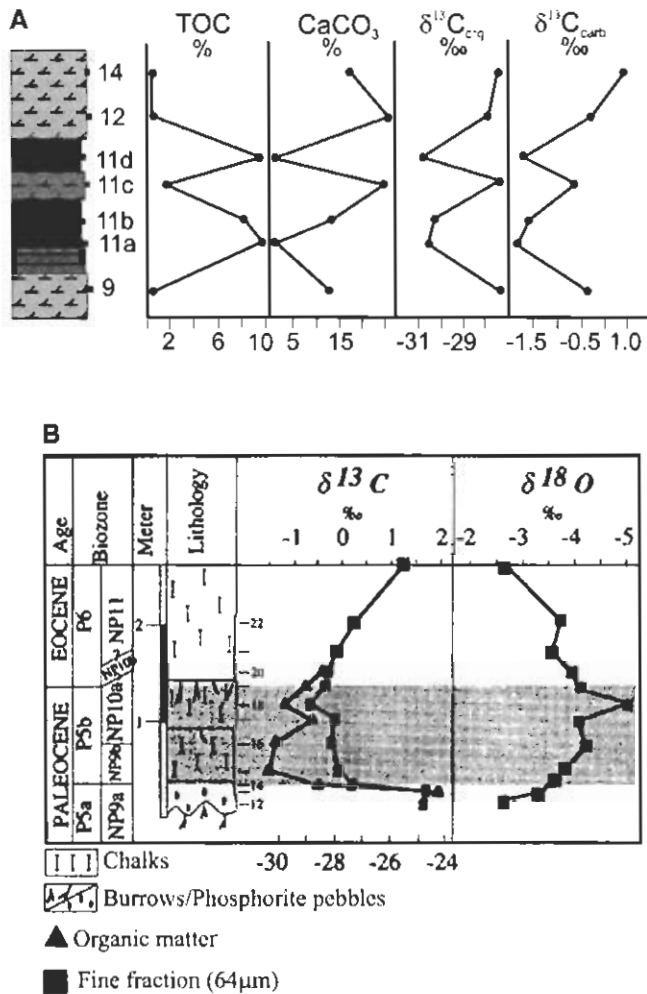


Figure 8. Oxygen and carbon isotopes in the sapropelitic bed: A: Kheu River section. B: Aktumsuk section (Bolle et al., 2000).

The planktic/benthic foraminifera ratio increased sharply in the sapropelitic bed, with benthic biota represented mostly by dysoxic species of benthic foraminifera, indicating varying degrees of dysoxia/anoxia within the basin (Stupin and Muzylöv, 2001). The benthic fauna recovered partially in the middle part of the sapropelitic bed, concurrent with the recovery of calcareous plankton and a decrease in TOC, suggesting that dysoxia/anoxia weakened temporarily. The changes in benthic foraminiferal assemblages in the sapropelitic bed of the northeastern Peri-Tethys thus resemble those documented in coeval sediments in Egypt (Speijer et al., 1997; Speijer, 2003). In the Torangly section, buliminids, generally thought to indicate a high food supply, were abundant (Speijer et al., 1997).

The calcareous microplankton apparently recovered during the deposition of the sapropelitic bed, after the "presapropelitic" crisis (thin interval without microplankton). We do not consider that carbonate dissolution caused the absence of calcareous plankton in this layer; in fact, dissolution appears to have been

somewhat more important in the sapropelitic bed. Remains of organic-walled plankton dominated during the accumulation of the bed. The acme in dinoflagellate *Apectodinium* cysts in sediments deposited during the IETM has usually been explained as the result of high temperatures as well as high nutrient levels (e.g., Crouch et al., 2001). Dinoflagellates form a major part of primary planktonic production in oceans, and their blooms usually occur in warm, eutrophic waters. Dinoflagellates are also known for producing toxins in the so-called "red tide" blooms, causing fish and plankton kills (e.g., Taylor and Seliger, 1979; Steidinger and Baden, 1984).

When the sapropelitic bed accumulation gradually came to an end, the nannofloras recovered slowly in abundance and diversity. In the Torangly section the benthic foraminifera also show a gradual recovery (Speijer et al., 1997), as they did at ODP Site 1051 (Katz et al., 1999) and at several other sites globally (Thomas, 1998). Low-diversity nannofloras persisted for a long time after the accumulation of the sapropelitic bed, although the relative abundances of the most important taxa recovered to pre-sapropelitic bed levels. *Discoaster mahmoudi* appeared just after the deposition of the sapropelitic bed, with a FO apparently synchronous over the Tethys, as well as in a number of sections in Spain and Egypt (Moneschi et al., 2000).

Depositional scenario for the sapropelitic bed

The deposition of the sapropelitic bed in the northeastern Peri-Tethys was the regional expression of the global IETM event in the unique setting of a large epicontinental basin. The regional features developed in response to the global events (climatic warming, carbon isotope excursion [CIE], benthic extinction event) documented to have occurred on land, in open ocean, and shelf areas. The sapropelitic bed is characterized by a light carbon isotopic composition of organic matter and carbonates (Kodina et al., 1995), corresponding to globally light values of carbon isotopes in the ocean-atmosphere system during the IETM (e.g., Kennett and Stott, 1991; Thomas and Shackleton, 1996), which probably resulted from the large-scale dissociation of gas hydrates (e.g., Dickens et al., 1995). The CIE in the northern Peri-Tethys section is similar in magnitude to that in open ocean sections, and we thus assume that values were probably not influenced by regional factors to a major degree.

The duration of the accumulation of the sapropelitic bed can be only approximately estimated by comparison with rates of sediment deposition elsewhere. Sapropelites may accumulate rapidly; for example, a sapropelitic bed of similar thickness as the sapropelitic bed accumulated in the Black Sea over a period of ~4–5 k.y. (Vinogradov et al., 1962). Mediterranean sapropels of similar thickness are also thought to have been accumulated over several thousands of years, and the accumulation rate of sapropels is 3–5 times greater than that of surrounding sediments (Sutherland et al., 1984). A lower Eocene succession of sapropelitic bed (10–15 cm) accumulated within 2 k.y. maximum (Oberhänsli and Beniamovskii, 2000). An approximate es-

timate obtained by assuming that the lamination of the sapropelitic bed represents annual laminae (Plate 1, Figure 3) indicates that the whole of the bed accumulated in ~10–12 k.y. However, correlation of the carbon isotope record of the sapropelitic bed (Fig. 8) with records in oceanic sections suggests that the main part of the bed formed during the peak of the CIE, which lasted ~60 k.y. (3 precession cycles; Röhl et al., 2000). We can presently not explain this disagreement.

The formation of organic-rich sediments is usually thought to result from increased biological productivity (e.g., Pedersen and Calvert, 1991). We agree with these authors that high productivity was the most important factor in the accumulation of the sapropelitic bed, while stratification of the water column may have played a role, but was not the main cause of low-oxygen conditions. Models proposed for the genesis of the Pliocene-Pleistocene eastern Mediterranean sapropels (e.g., Kidd et al., 1978; Calvert, 1983; Sutherland et al., 1984) and the Black Sea (Strakhov, 1971) are probably not relevant to the deposition of the sapropelitic bed, because the northeastern Peri-Tethys was a very large, epicontinental, relative shallow-water basin, separated from the open ocean by a system of archipelagoes, and thus had an oceanographic setting very different from that of the eastern Mediterranean and the Black Sea.

The sapropelitic bed accumulated during a rapid transgression, as documented by its sharp, sometimes erosional lower boundary. This transgression was preceded by a short regression (Muzylöv et al., 1996; Gavrilov et al., 1997), very similar to interpretations of the setting of coeval TOC-rich sediments at the southern Peri-Tethyan margin (Speijer, 2003; Speijer and Morsi, 2001). These authors argue that both increased productivity and increased preservation of organic matter were important factors in the deposition of the organic-rich sediments, and that the high productivity was driven by increased upwelling of nutrient-rich waters through offshore Ekman transport.

In contrast to these authors, we think that upwelling cannot have been efficient throughout the large, shallow epicontinental basin of the northern Peri-Tethys, and argue that high productivity throughout this vast region was mainly caused by an increased supply of nutrients into the basin from coastal regions during a rapid transgression (Fig. 9) (Gavrilov, 1994; Gavrilov and Kopaevich, 1996; Gavrilov et al., 1997).

This increased nutrient flux resulted from a rather complex series of events. During the rapid regression just before the IETM, large lowland areas with little relief formed along the periphery of the basin and around archipelagoes in a relatively shallow epicontinental sea. Non- or weakly lithified sediments in the newly exposed, widespread coastal plains underwent soil formation, and the soils became enriched in nutrients. In humid areas, such as the Caucasus and adjacent areas, lakes and water-logged bogs developed rapidly in the extensive coastal plains. In present-day peat bogs, P is very soluble and becomes enriched in surface waters (Kovalev, 1985). Water-logged basins contain abundant dissolved and particulate organic matter from the decomposition of vegetation.

We argue that a rapid transgression flooded the coastal plains at the beginning of the formation of the sapropelitic bed (Fig. 9). Because most of the northeastern Peri-Tethys was a peneplain formed in the Late Cretaceous and earliest Paleogene, even minor sea-level fluctuations resulted in a major change in basin outline. The sea-level rise was at least several tens of meters (Haq et al., 1987; Gavrilov et al., 1997) in agreement with estimates by Speijer and Morsi (2001) and Speijer (2003) for the southern margin of Peri-Tethys. During the sea-level rise, particulate and dissolved organic matter was transported into the sea from the drowned peat bogs and soils. The particulate organic matter was partially redeposited, enhancing enrichment in TOC in some regions (e.g., the Kheu river section). The dissolved organic matter and nutrients (including P which is enriched in most sapropelitic bed sections) were transported into the basin, triggering a strong increase in productivity of organic-walled and possibly prokaryote phytoplankton. The blooms of organic-walled phytoplankton may have adversely affected other plankton groups and fish, as observed in present-day "red tides" in eutrophic coastal regions.

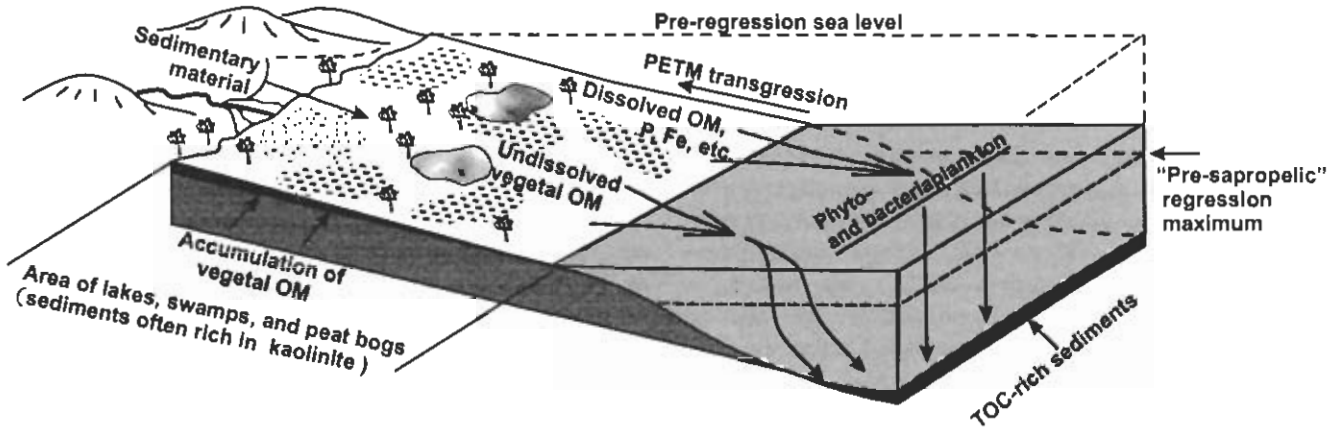
The hypothesis that the transgression played a major role in triggering increased productivity was also invoked to explain the origin of Pennsylvanian black shales (Wenger and Baker, 1986). In addition, data on events during the filling of reservoirs behind dams, which simulates a transgression, support this hypothesis: During the initial filling of reservoirs, phytoplankton productivity is extremely high because large amounts of nutrients are washed from the flooded soils into the basin (Petrova, 1990).

In some sections in the eastern arid part of the basin (e.g., the Kurpai section) only a negligible amount of terrestrial organic matter accumulated in the sapropelitic bed, but the bed is closely associated with phosphate-bearing rocks and phosphorites. The lowermost part of the Suzak Formation (underlying the oil shale unit) contains abundant phosphate concretions. We argue that a significant amount of P, concentrated in soils and produced by weathering in pre-sapropelitic bed time, was supplied to the basin during the transgression, triggering phytoplankton blooms and causing high organic matter concentrations.

The high productivity of organic-walled plankton supplied most of the organic material present as TOC in the sapropelitic bed, while productivity of calcareous plankton was low, although we may underestimate calcareous plankton productivity as a result of slight dissolution. This high productivity was the major cause of the anoxia/dysoxia in the water column, as indicated by the impoverished benthic foraminiferal faunas, sediment lamination, and organic biochemical tracers. In addition, it has been argued that open-ocean waters were overall low in oxygen during the IETM (e.g., Thomas, 1998), and these global low oxygen levels could have influenced oxygen levels in marginal basins (e.g., Speijer, 2003).

We cannot, however, completely exclude the possibility that the anoxia/dysoxia was exacerbated by stratification and resulting stagnation. Increased runoff of fresh water could have caused stratification in western parts of the basin. The only ev-

A



B

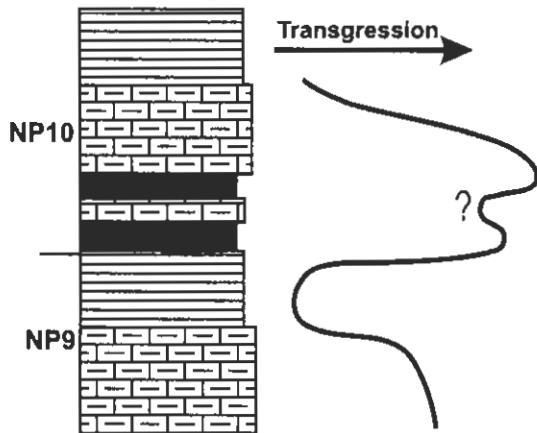


Figure 9. Suggested model of formation of the sediment rich in organic matter in the northeastern Peri-Tethys (A) and suggested sea-level changes (B).

idence for such an increase in runoff can be seen in the low values of oxygen isotopes in bulk carbonates in the sapropelitic bed in the Aktumsuk section (Fig. 8). But the thick-walled dinocyst abundance, the blooms of the calcareous dinocyst *Thoracosphaera*, and diverse nannofossil assemblages indicate eutrophic environments and normal salinity, and the biota thus do not indicate lowered salinity. At other times, the biota very clearly documented increased runoff. For instance, well-documented episodes of high freshwater influx in the northeastern Peri-Tethys (e.g., in the middle Oligocene part of the Maykop Formation) are characterized by the occurrence of abundant monospecific nannofossil assemblages associated with small thin-walled dinocyst remains (M. Akhmetiev, 2002, personal commun.). We thus have no solid evidence for increased runoff resulting in lowered salinity, and conclude that deposition of the sapropelitic bed was probably not caused by stagnation resulting from stable stratification as a result of increased runoff, even in the more humid western parts of the basin.

In the eastern part of the area conditions were arid, as documented by the higher palygorskite content in sections in the Fergana embayment and the Amu-Daria region, Uzbekistan. Over the whole basin, and not just in the western regions where climate was humid, the transgression was followed by the deposition of TOC-rich sediments, with maximal TOC concentrations in arid areas (e.g., in the Kurpai and Guru-Fatima sections). In these arid eastern and southeastern part of the basin, increased stratification might have resulted from higher evaporation rates during the warm IETM, similar to the coeval episode of extensive evaporation documented in Spanish sections (Schmitz and Pujalte, 2001). Hypersaline lagoons and lakes may have formed during pre-sapropelitic bed regression, and supplied saline water to the marine basin during the following transgression, possibly increasing local salinity and causing density stratification. Stable density stratification thus may have been a contributing factor to the high concentrations of organic matter in sediments (e.g., TOC >20% in the Suzak oil shales).

The accumulation of organic matter in the sediments caused sulfate-reduction and the generation of H_2S that diffused into the water, accumulated close to the bottom, and locally reached the photic zone, as shown by the biomarkers. The development of anoxic and dysoxic conditions caused extinction of benthic fauna, the appearance of anomalous planktonic taxa and possibly impoverished planktonic assemblages where these conditions reached the photic zone. The intensity of dysoxia/anoxia varied in different parts of the basin, as indicated by differences in benthic foraminiferal assemblages (Stupin and Muzylöv, 2001).

The presence of H_2S affected the geochemical signature of the sediments, too. In all sections, the sapropelitic bed is strongly enriched in Mo with respect to the surrounding sediments (Fig. 7), and enrichment patterns are similar for Se and

Re (Gavrilov et al., 1997). These elements (Mo, Se, Re) are redox-sensitive and form insoluble sulfides in the presence of H_2S in the water column. This process caused the observed enrichment in the sapropelitic bed, in combination with other mechanisms for the concentration of minor elements, such as sorption by clay particles and organic matter, and possibly diagenetic enrichment.

In some sections, we observed that biotic changes started below the base of the sapropelitic bed, where a thin layer of sediment contains neither plankton nor benthos. We have no lithological or geochemical evidence of major taphonomic changes during diagenesis in this layer (e.g., calcite redistribution). We suggest that a dramatic decrease in productivity may have occurred during the regression, which occurred before the rapid transgression that led to deposition of the sapropelitic bed. During this regression, the waters had a high turbidity in some parts of the basin, as a result of erosion of weakly lithified sediments, as documented by the occurrence of detrital carbonate and reworked Cretaceous nannofossils in some sections. The increased turbidity would have decreased light penetration in the water column, thus reducing primary productivity, and influencing floral and faunal composition.

After the peak of the transgression, the rich supply of nutrients gradually decreased with the deposition of the organic matter in the sapropelitic bed, and the intense blooms of various forms of plankton ended, ending the dysoxic to anoxic conditions. Besides, global ocean waters became more oxygenated (e.g., Thomas, 1998), which may have assisted in helping to alleviate the environmental conditions in the marginal basins (e.g.,

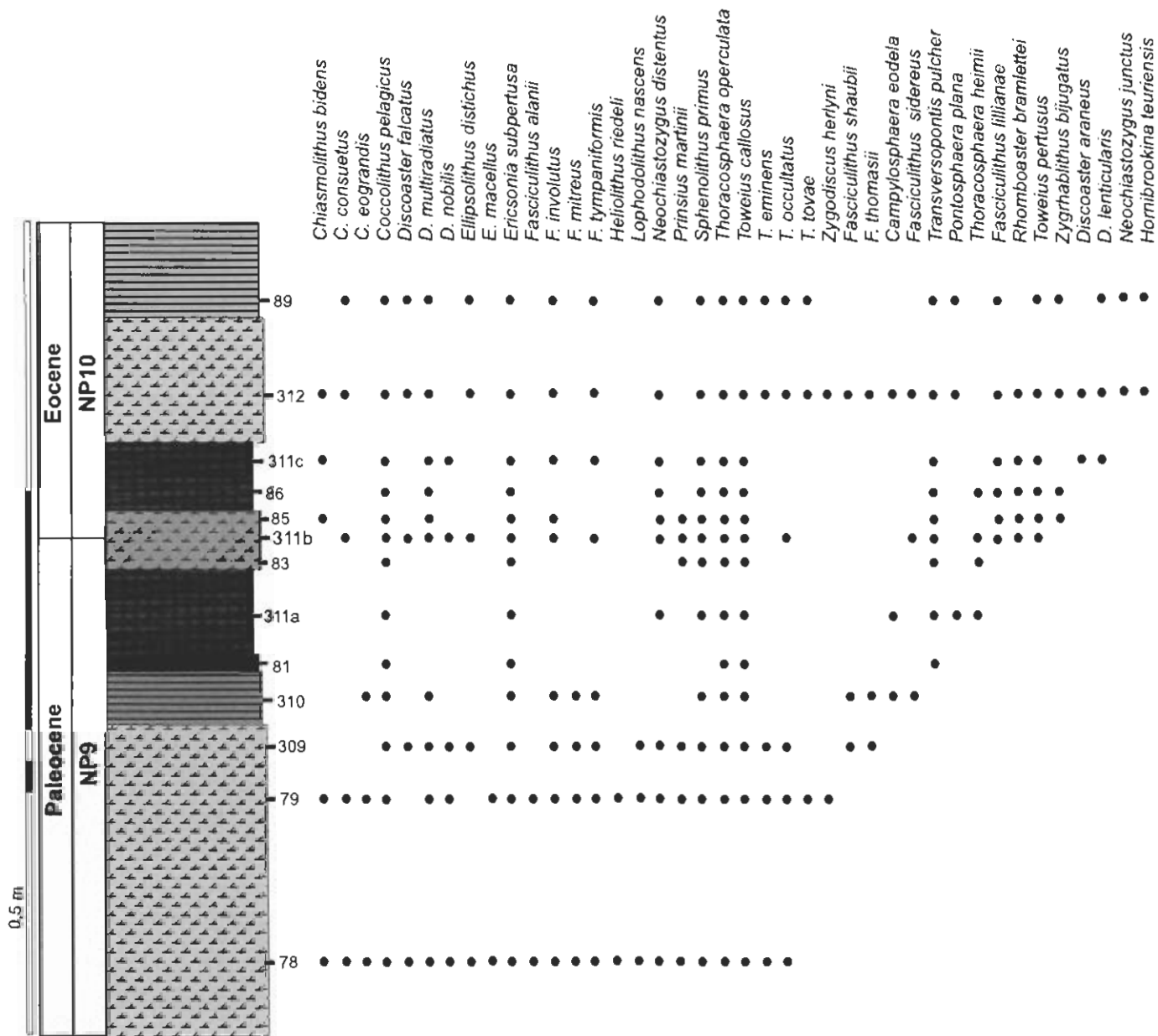
Speijer, 2003). In addition, the world cooled gradually (e.g., Zachos et al., 2001), which made it possible to dissolve more oxygen in the waters.

There has been much speculation regarding the cause of the end of the IETM, and the prevention of a runaway greenhouse effect. It has been argued that increased productivity on land (Beerling, 2000) or in open ocean (Bains et al., 2000) was responsible for the removal of carbon dioxide from the atmosphere, thus ending the IETM. We suggest that the deposition of huge amounts of organic material in the sapropelitic bed in the vast, northeastern Peri-Tethys may very well have contributed to the ending of the IETM and restoration of more normal environmental conditions.

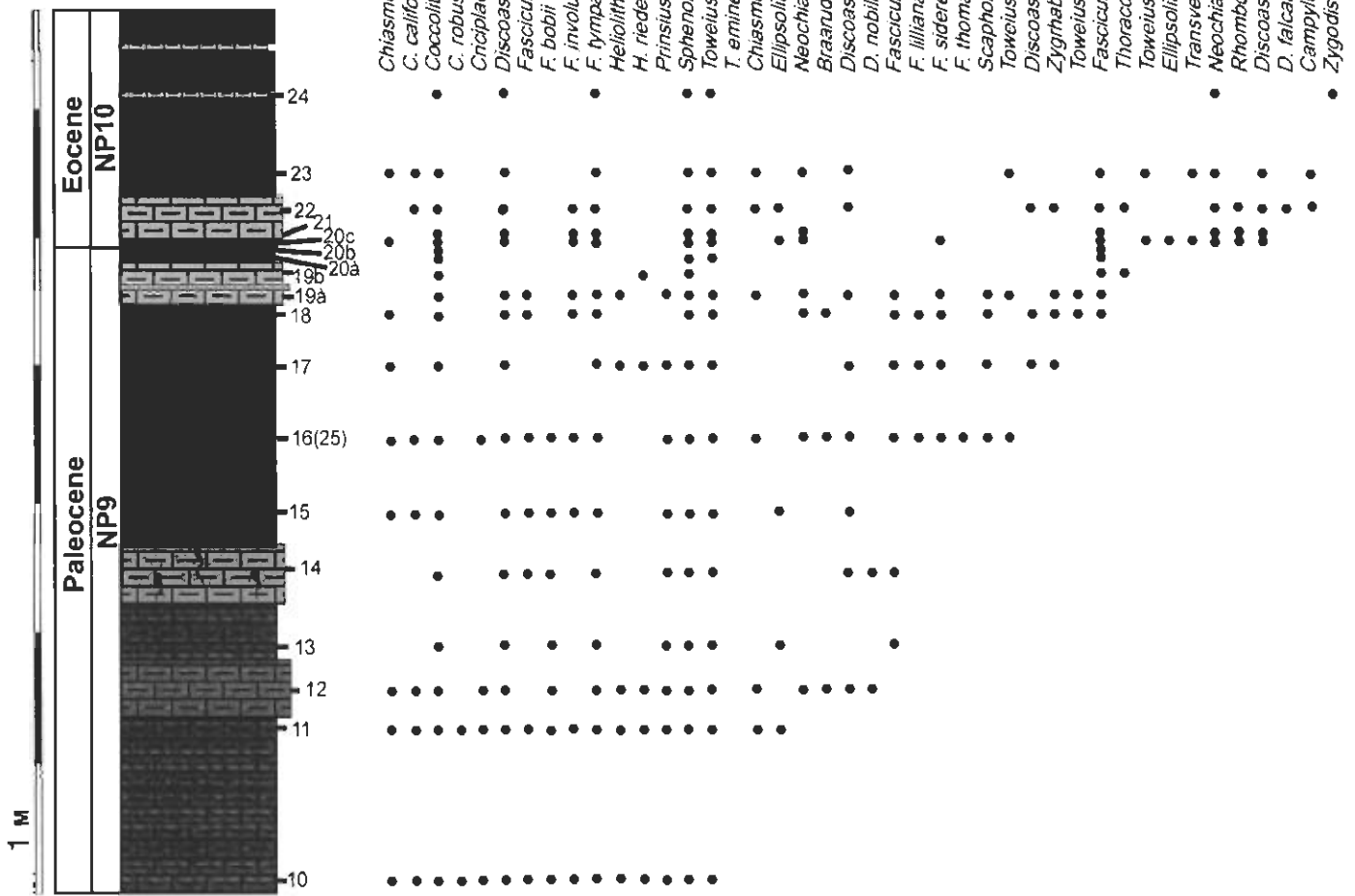
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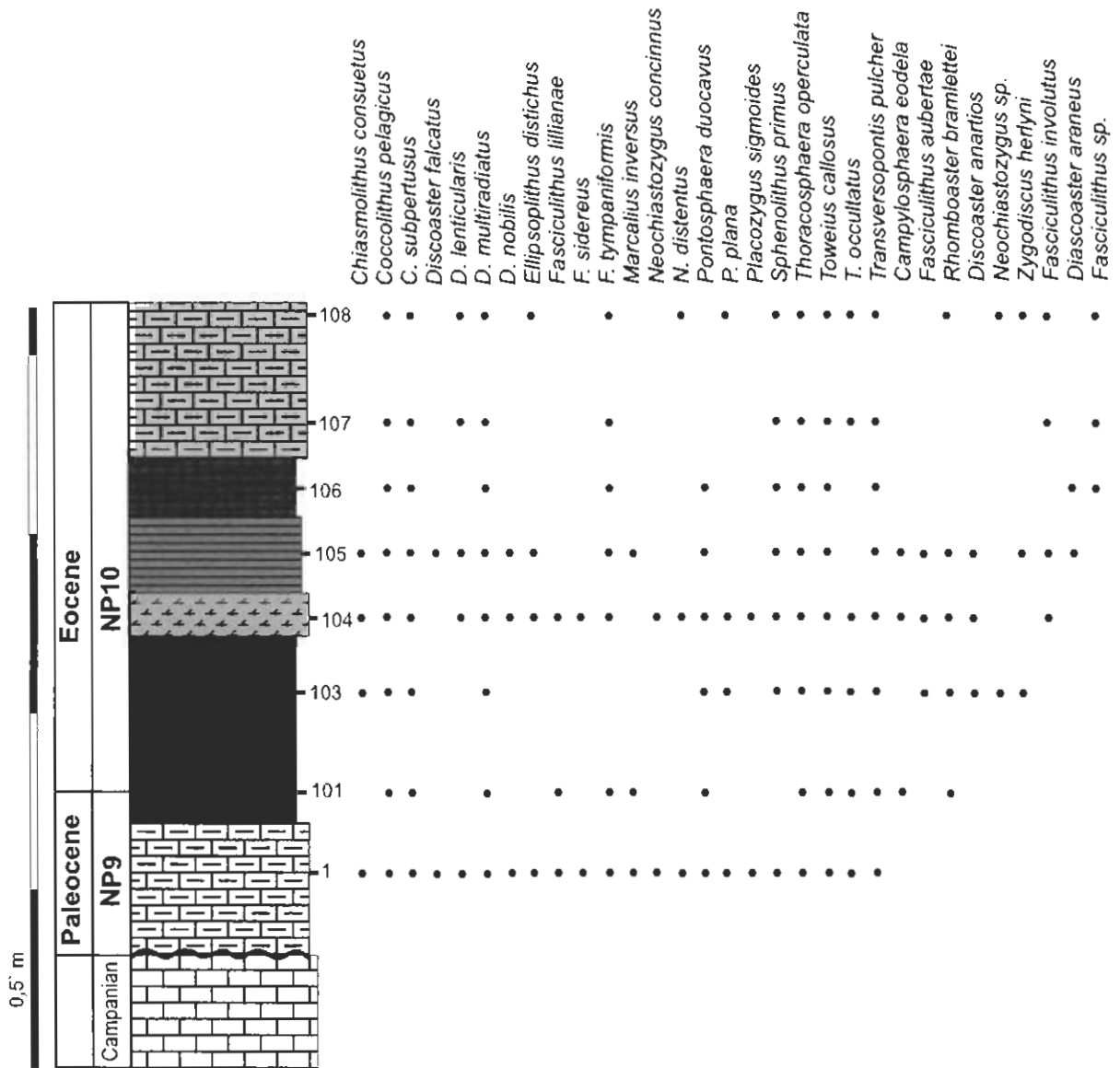
This research was supported by the Russian Foundation for Basic Research, projects no. 00-05-64593 and 01-05-64805 and INTAS grant no. 93-2509.



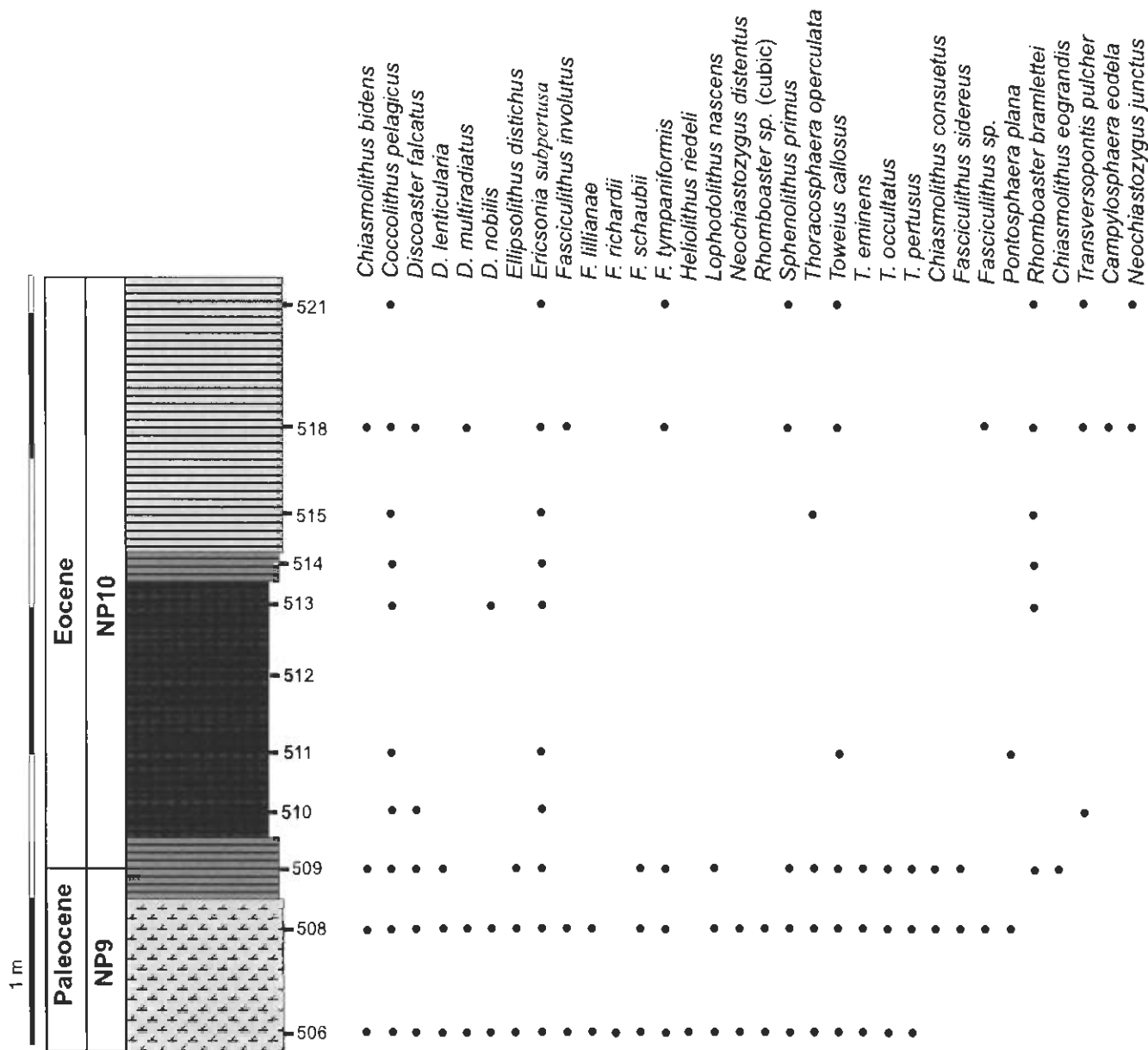
Appendix A. Nannofossil range chart of the sapropelitic bed of Kheu River section. Note that the Paleocene-Eocene boundary was drawn at the NP9/NP10 zonal boundary.



Appendix B. Nannofossil range chart of the sapropelitic bed of Sulak River section. Note that the Paleocene-Eocene boundary was drawn at the NP9/NP10 zonal boundary.



Appendix C. Nannofossil range chart of the sapropelitic bed of Aktumsuk section. Note that the Paleocene-Eocene boundary was drawn at the NP9/NP10 zonal boundary.



Appendix D. Nannofossil range chart of the sapropelitic bed of Torangly section. Note that the Paleocene-Eocene boundary was drawn at the NP9/NP10 zonal boundary.

REFERENCES CITED

- Akhmetiev, M.A., and Zaporozhets, N.I., 1996. Changes in dinocysts assemblages as reflecting in ecosystem transformations in the Russian Platform, Crimea, Caucasus, and Turanian plate during Paleogene and early Miocene, in Kuznetsova, K.I., and Muzyl'ov, N.G., eds., Fossil microorganisms as the basis of the Phanerozoic stratigraphy, correlation, and paleobiogeography: Moscow, GEOS, p. 55–69 (in Russian).
- Aubry, M.-P., Berggren, W.A., Stott, L., and Sinha, A., 1996. The upper Paleocene–lower Eocene stratigraphic record and the Paleocene–Eocene boundary carbon isotope excursion: Implications for geochronology, in Knox, R.W.O'B., et al., eds., Correlation of the early Paleogene in northwestern Europe: Geological Society [London] Special Publication 101, p. 353–380.
- Aubry, M.-P., Lucas, S.G., and Berggren, W.A., editors, 1998. Late Paleocene–early Eocene climatic and biotic events in the marine and terrestrial records: New York, Columbia University Press, 513 p.
- Bains, S., Norris, R.D., Corfield, R.M., and Faul, K., 2000. Termination of global warmth at the Paleocene/Eocene boundary through productivity feedback: *Nature*, v. 407, p. 171–174.
- Beerling, D., 2000. Increased terrestrial carbon storage across the Paleocene–Eocene boundary: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 161, p. 395–405.
- Bolle, M.-P., Pardo, A., Adatte, T., Tantawy, A.A., Hinrichs, K.-N., von Salis, K., and Burns, S., 2000. Climatic evolution on the southern and northern margins of the Tethys from the Paleocene to the early Eocene: *GFF*, v. 122, p. 31–32.
- Bralower, T.J., and Premoli-Silva, I., 2002. Leg 198 Summary: Bralower, T.J., et al., eds., Ocean Drilling Program, Preliminary Report, Leg 198. http://www.dp.tamu.edu/publications/198_IR/VOLUME/CHAPTERS/IR_198_01.PDF.
- Bujak, J.P., and Brinkhuis, H., 1998. Global warming and dinocyst changes across the Paleocene/Eocene epoch boundary, in Aubry, M.-P., et al., eds., Late Paleocene–early Eocene biotic and climatic events in the marine and terrestrial records: New York, Columbia University Press, p. 277–295.
- Bukry, D., 1974. Coccoliths as paleosalinity indicators: Evidence from Black Sea, in Degens, E.T., and Ross, D.A., eds., Black Sea: Geology, chemistry, and biology: Tulsa, Oklahoma, American Association of Petroleum Geologists Publishing, p. 353–363.
- Bybell, L.M., and Self-Trail, J.M., 1994. Evolutionary, biostratigraphic, and taxonomic study of calcareous nannofossil from a continuous Paleocene/Eocene boundary section in New Jersey: U.S. Geological Survey Professional Paper 1554, 36 p.
- Bybell, L.M., and Self-Trail, J.M., 1997. Late Paleocene and early Eocene calcareous nannofossils from three boreholes in an onshore-offshore transect from New Jersey to the Atlantic continental rise, in Miller, K.J., and Snyder, S.W., eds., Proceedings of the Ocean Drilling Program, Scientific Results: Texas A & M University, College Station, Texas, USA, v. 150X, p. 91–110.
- Calvert, S.E., 1983. Geochemistry of Pleistocene sapropels and associated sediments from the Eastern Mediterranean: *Oceanologica acta*, v. 6, no. 3, p. 255–267.
- Calvert, S.E., 1990. Geochemistry and origin of the Holocene sapropel in the Black Sea, in Ittekkot, V., et al., eds., Facets of modern biogeochemistry: Berlin, Springer-Verlag, p. 326–352.
- Crouch, E., Bujak, J.P., and Brinkhuis, H., 2000. Southern and Northern Hemisphere dinoflagellate cyst assemblage changes in association with the late Paleocene thermal maximum: *GFF*, v. 122, p. 40–41.
- Crouch, E.M., Heilmann-Clausen, C., Brinkhuis, H., Morgans, H.E.G., Rogers, K.M., Egger, H., and Schmitz, B., 2001. Global dinoflagellate event associated with the late Paleocene thermal maximum: *Geology*, v. 29, p. 315–318.
- Dickens, G.R., O'Neil, J.R., Rea, D.K., and Owen, R.M., 1995. Dissociation of oceanic methane hydrate as a cause of the carbon isotope excursion at the end of Paleocene: *Geology*, v. 25, p. 259–262.
- Fomina, L.S., and Volkov, I.I., 1970. Rare earth elements in the Black Sea sediments: *Litologia i poleznye iskopaemye*, no. 2, p. 148–160 (in Russian).
- Gavrilov, Yu.O., 1994. The possible causes of sediment rich in organic matter accumulation in relation with eustatic sealevel changes, in Semikhatov, M.A., and Rozanov, A.Yu., eds., Ecosystem restructures and the evolution of biosphere: Moscow, Nedra, p. 305–311 (in Russian).
- Gavrilov, Yu.O., and Kopaevich, L.F., 1996. Geochemical, biochemical, and biotic consequences of eustatic fluctuations: *Stratigraphy and Geological Correlation*, v. 4, no. 4, p. 3–14.
- Gavrilov, Yu.O., and Muzyl'ov, N.G., 1991. The geochemistry of sapropelitic interbeds in the Paleogene of Central Caucasus: *Lithology and Mineral Resources*, v. 6, p. 84–98.
- Gavrilov, Yu.O., Kodina, L.A., Lubchenko, I.Yu., and Muzyl'ov, N.G., 1997. The Late Paleocene anoxic event in epicontinental seas of Peri-Tethys and formation of the sapropelitic unit: *Sedimentology and geochemistry: Lithology and Mineral Resources*, v. 5, p. 492–517.
- Gavrilov, Yu.O., Sheherbinina, E.A., and Muzyl'ov, N.G., 2000. A Paleogene sequence in central North Caucasus: A response to paleoenvironmental changes: *GFF*, v. 122, p. 51–53.
- Haq, B.U., Hardenbol, J., and Vail, P.R., 1987. Chronology of fluctuating sea levels since Triassic: *Science*, v. 235, p. 1156–1167.
- Karwath, B., Janofske, D., Tetjen, F., and Willems, H., 2000. Temperature effects on growth and cell size in the marine calcareous dinoflagellate *Thoracosphaera heimii*: *Marine Micropaleontology*, v. 39, no. 1–4, p. 43–51.
- Katz, M.E., Pak, D.K., Dickens, G.R., and Miller, K.G., 1999. The source and fate of massive carbon input during latest Paleocene thermal maximum: *Science*, v. 286, p. 1531–1533.
- Kennett, J.P., and Stott, L.D., 1991. Abrupt deep-sea warming, paleoceanographic changes, and benthic extinction at the end of Paleocene: *Nature*, v. 353, p. 225–229.
- Kidd, R.B., Cita, M.B., and Ryan, W.B.F., 1978. Stratigraphy of Eastern Mediterranean sapropel sequences recovered during DSDP Leg 42A and their paleoenvironmental significance, Hsu, K.J., et al., eds., Initial reports of the Deep Sea Drilling Project: Washington, D.C., U.S. Government Printing Office, v. 42, p. 421–443.
- Koch, P.L., Zachos, J.C., and Gingerich, P.D., 1992. Coupled isotopic changes in marine and continental carbon reservoirs at the Paleocene–Eocene boundary: *Nature*, v. 358, p. 319–322.
- Kodina, L.A., Huang, Y., Gavrilov, Yu.O., Jones, M., and Eglinton, G., 1995. Environment of upper Paleocene black shale deposition in southern Russia and adjacent regions as revealed by isotope and biomarker study, in Grimalt, J.O., and Dorronsoro, C., eds., Organic geochemistry: Developments and applications to energy, climate, environment and human history: Selected papers from the 17th International Meeting on Organic Geochemistry, Donostia-San Sebastian, The Basque Country, 4th–8th September 1995, p. 192–194.
- Kovalev, V.A., 1985. Bolotnye mineralogo-geokhimicheskie systemy (Mineralogical and geochemical systems of bogs): Minsk, Nauka i Tekhnika, 327 p. (in Russian).
- Luterbacher, H.P., Hardenbol, J., and Schmitz, B., 2000. Decision of the voting members of the International Subcommittee on Paleogene Stratigraphy on the criterion for the recognition of the Paleocene–Eocene boundary: *Newsletter of the International Subcommittee on Paleogene Stratigraphy*, Tübingen, November 2000, v. 9, p. 13.
- Martini, E., 1971. Standard Tertiary and Quaternary calcareous nannoplankton zonation, in Farinacci, A., ed., Proceedings of the 2nd Planktonic Conference, Rome, Italy, 1970: *Tecnoscienza*, v. 2, p. 739–785.
- Monechi, S., Angori, E., and Spejjer, R., 2000. Upper Paleocene stratigraphy in the Mediterranean region: Zonal markers, diachronism, and preservational problems: *GFF*, v. 122, p. 108–110.
- Muzyl'ov, N.G., 1994. Paleocene–middle Eocene anoxic events, in Semikhatov, M.A., and Rozanov, A.Yu., eds., Ecosystem restructures and the evolution of biosphere: Moscow, Nedra, p. 151–159 (in Russian).
- Muzyl'ov, N.G., and Sheherbinina, E.A., 1995. The main geological events of Thanetian–lower Ypresian of the South of the Former USSR: IGCP (International Geological Correlation Program) Project 308 meeting Paleocene–Eocene boundary events, Beer Sheva, Israel, 8–15 January 1995, p. 6–7.

- Muzylöv, N.G., Benyamovskii, V.N., and Tabachnikova, I.P., 1989, Sapropel interlayers in lower Paleocene deposits of the southern USSR: *Izvestia Akademii Nauk SSSR, Seria Geologicheskaya*, v. 11, p. 117–119 (in Russian).
- Muzylöv, N.G., Benyamovskii, V.N., Gavrilov, Yu.O., Shcherbinina, E.A., and Stupin, S.I., 1996, Paleontological and geochemical features of the Central Caucasian upper Paleocene sapropelite, in Kuznetsova, K.I., and Muzylöv, N.G., eds., *Fossil microorganisms as the basis of the Phanerozoic stratigraphy, correlation, and paleobiogeography*: Moscow, GEOS, p. 117–127.
- Oberhänsli, H., and Benyamovskii, V.N., 2000, Dysoxic bottom water events in the Peri-Tethys during the late Ypresian: A result of changes in the evaporation/precipitation balance in adjacent continental regions: *GFF*, v. 122, p. 21–123.
- Okada, H., and Bukry, D., 1980, Supplementary modification and introduction of code numbers to the low-latitude coccolith biostratigraphic zonation (Bukry, 1973, 1975): *Micropalaeontology*, v. 5, no. 3, p. 321–325.
- Orue-Etxebarria, X., Apellaniz, E., Baceta, J.I., Coccioni, R., Di Leo, R., Dinares-Turell, S., Galeotti, S., Monechi, S., Nunez-Beteln, K., Pares, J.M., Payros, A., Pujalte, V., Samsó, J.M., Serra-Kiel, J., Schmitz, B., and Tosquella, J., 1996, Physical and biostratigraphic analysis of two prospective Paleocene-Eocene boundary stratotypes in the intermediate-deep water Basque Basin, western Pyrenees: The Trabakua Pass and Ermua sections: *Neues Jahrbuch für Geologie und Paläontologie, Abhandlungen*, v. 200, p. 1–64.
- Pedersen, T.F., and Calvert, S.E., 1991, Anoxia vs. productivity: What controls the formation of organic-carbon-rich sediments and sedimentary rocks?: Discussion: AAPG (American Association of Petroleum Geologists) Bulletin, v. 75, p. 500–501.
- Perch-Nielsen, K., 1977, Albian to Pleistocene calcareous nannofossils from the western South Atlantic, in Stupko, P.R., ed., *Initial reports of the Deep Sea Drilling Project*: Washington, D.C., U.S. Government Printing Office, v. 39, p. 699–824.
- Petrova, N.A., 1990, Phytoplankton successions related to the antropogenic eutrophic processes in large lakes: Leningrad, Nauka, 199 p. (in Russian).
- Powell, A.J., 1992, Dinoflagellate cysts of the Tertiary System, in Powell, A.J., ed., *A stratigraphic index of dinoflagellate cysts*: London, Chapman and Hall, p. 155–251.
- Repeta, D.J., and Simpson, D.J., 1991, The distribution and recycling of chlorophyll, bacteriochlorophyll and carotenoids in the Black Sea: *Deep-Sea Research*, v. 38, Supplementary 2, p. 969–984.
- Repeta, D.J., Simpson, D.J., Jorgensen, B.B., and Jannasch, H.W., 1989, Evidence of anoxygenic photosynthesis from the distribution of bacteriochlorophylls in the Black Sea: *Nature*, v. 342, p. 69–72.
- Röhl, U., Bralower, T.J., Norris, R.D., and Wefer, G., 2000, New chronology for the late Paleocene thermal maximum and its environmental implications: *Geology*, v. 28, p. 927–930.
- Schmitz, B., and Pujalte, V., 2001, Correlating the Paleocene-Eocene boundary in the Pyrenees, Spain, by carbon isotopes: A transect from deep marine to continental environments, in Ash, A.W., and Wing, S.L., eds., *Climate and biota of the early Paleogene*: International Meeting July 3–8, 2001, Northwest College, Powell, Wyoming, USA, Abstract volume, p. 81.
- Schmitz, B., Heilmann-Clausen, C., King, C., Steurbaut, E., Andreasson, F.P., Corfield, R.M., and Cartledge, J.E., 1996, Stable isotope and biotic evolution in the North Sea during the early Eocene: The Albaek Hoved section, Denmark, in Knox, R.W.O'B., et al., eds., *Correlation of the early Paleogene in northwestern Europe*: Geological Society [London] Special Publication 101, p. 275–306.
- Sinninghe Damste, J.S., Wakeham, S.G., Kohnen, M.E.L., Hayes, J.M., and de Leeuw, J.W., 1993, A 6000-year sedimentary molecular record of chemocline excursions in the Black Sea: *Nature*, v. 362, p. 827–829.
- Speijer, R.P., and Morsi, A.M., 2001, First detailed ostracod records indicating sea-level change associated with the LPTM (Eastern Desert, Egypt), in Ash, A.W., and Wing, S.L., eds., *Climate and biota of the early Paleogene*: International Meeting July 3–8, 2001, Northwest College, Powell, Wyoming, USA, Abstract volume, p. 88.
- Speijer, R.P., and Wagner, T., 2001, Black shale deposition and sea-level change in the Tethys during the LPTM, in Ash, A.W., and Wing, S.L., eds., *Climate and biota of the early Paleogene*: International Meeting July 3–8, 2001, Northwest College, Powell, Wyoming, USA, Abstract volume, p. 89.
- Speijer, R.P., Schmitz, B., and van der Zwaan, G.J., 1997, Benthic foraminifera extinction and repopulation in response to latest Paleocene anoxia: *Geology*, v. 25, p. 683–686.
- Steidinger, K.A., and Badeu, D.G., 1984, Toxic marine dinoflagellates, in Specator, D.L., ed., *Dinoflagellates*: New York, Academic, p. 201–261.
- Steurhaut, E., De Conink, J., Dupui, C., and Kiug, C., 2000, Dinoflagellate cyst events and depositional history of the Paleocene-Eocene boundary interval in the southern North Sea basin: *GFF*, v. 122, p. 154–155.
- Strakhov, N.M., 1971, Geochemical evolution of the Black Sea in Holocene: *Litologia i Poleznie iskopaemye*, no. 3, p. 3–17 (in Russian).
- Stupin, S.I., and Muzylöv, N.G., 2001, The late Paleocene Ecologic crisis in epicontinental basins of the eastern Peri-Tethys: Microbiota and accumulation conditions of sapropelitic bed: *Stratigraphy and Geological Correlation*, v. 9, no. 5, p. 501–507.
- Summons, R.E., and Powell, T., 1986, Chlorobiaceae in Paleozoic seas revealed by biological markers, isotopes, and geology: *Nature*, v. 319, p. 763–765.
- Summons, R.E., and Powell, T., 1987, Identification of aryl isoprenoids in source rocks and crude oils: Biological markers for the green sulphur bacteria: *Geochimica and Cosmochimica Acta*, v. 51, p. 557–566.
- Sutherland, H.E., Calvert, S.E., and Morris, R.J., 1984, Geochemical studies of the recent sapropel and associated sediment from the Hellenic Outer Ridge, eastern Mediterranean Sea: Mineralogy and chemical composition: *Marine Geology*, v. 56, no. 1–4, p. 79–92.
- Tantawy, A.A., Ouda, Kh., von Salis, K., and Saad El-Din, M., 2000, Biostratigraphy of Paleocene sections in Egypt: *GFF*, v. 122, p. 163–164.
- Taylor, D.L., and Seliger, H.H., editors, 1979, *Toxic dinoflagellate blooms*: New York, Elsevier, 528 p.
- Thomas, E., 1998, The biogeography of the late Paleocene benthic foraminiferal extinction, in Aubry, M.-P., et al., eds., *Late Paleocene–early Eocene biotic and climatic events in the marine and terrestrial records*: New York, Columbia University Press, p. 214–243.
- Thomas, E., and Shackleton, N., 1996, The Paleocene-Eocene benthic foraminiferal extinction and stable isotope anomalies, in Knox, R.W.O'B., et al., eds., *Correlation of the early Paleogene in northwest Europe*: Geological Society [London] Special Publication 101, p. 401–441.
- Vink, A., Brune, A., Hall, C., Zommerveld, K.A.F., and Willems, H., 2002, On the response of calcareous dinoflagellates to oligotrophy and stratification of the upper water column in the equatorial Atlantic Ocean: *Palaeogeography, Palaeoclimatology, Palaeogeography*, v. 178, no. 1–2, p. 53–66.
- Vinogradov, A.P., Grinenko, V.A., and Ustinov, V.I., 1962, Isotopic composition of sulfur in the Black Sea: *Geokhimiya*, no. 10, p. 851–873 (in Russian).
- Volkov, I.I., 1973, Main regularities in the distribution of chemical elements in deepwater sediments of the Black Sea: *Litologia i poleznye iskopaemye*, no. 2, p. 3–22.
- von Salis, K., Monechi, S., Bybell, L.M., Self-Trail, J.M., and Young, J., 2000, Remarks on calcareous nanofossil markers Rhomboaster and Tribrahiastus around the Paleocene-Eocene boundary: *GFF*, v. 122, p. 138–140.
- Wenger, L.M., and Baker, D.R., 1986, Variation in organic geochemistry of anoxic-oxic black shale-carbonate sequences in the Pennsylvanian of the Midcontinent, USA: *Organic Geochemistry*, v. 10, no. 1–3, p. 85–92.
- Wing, S.L., Alroy, J., and Hickey, L.J., 1995, Plant and mammal diversity in the Paleocene to early Eocene of the Bighorn Basin: *Palaeogeography, Palaeoclimatology, Palaeoecology*, v. 115, p. 117–156.
- Zachos, J.C., Lochman, K.C., Walker, J.C.G., and Wise, S.W., Jr., 1993, Abrupt climate change and transient climates during Paleogene: A marine perspective: *Journal of Geology*, v. 101, p. 191–213.
- Zachos, J., Pagani, M., Sloan, L., Thomas, E., and Billups, K., 2001, Trends, rhythms, and aberrations in global climate 65 Ma to present: *Science*, v. 292, p. 686–693.