

New Maastrichtian oxygen and carbon isotope record: Additional evidence for warm low latitudes

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ABSTRACT: The Cretaceous period was generally characterized by greenhouse conditions. Nevertheless, our data on isotopic composition of biogenic carbonates from the Koryak Upland and Sakhalin (Russian Far East) show that during the Maastrichtian, temperatures dropped sharply at high and middle latitudes, with only a slight warming in the early Late Maastrichtian. At the same time, there is contradictory evidence on climatic conditions for low latitude areas during Maastrichtian time. The new and previously published isotopic data on Maastrichtian mollusks in the Western Interior Seaway (North America) (WIS) and some other areas suggest that tropical deep-sea surface temperatures calculated from the oxygen isotopic composition of the majority of investigated Maastrichtian planktic foraminifera are, obviously, underestimated. Unusually low isotopic temperatures were obtained for tropical planktic foraminifera. This probably reflects both local conditions provoked, first of all, by the influence of tropical upwelling zones, and the ability of Maastrichtian planktic foraminifera to migrate within a large vertical interval in the tropical zone in conditions of weakly stratified (well-mixed) ocean. The average tropical deep-sea surface paleotemperature estimates for the Maastrichtian could have been about 26.6–30.2°C, but, apparently, did not reach the level denoted for the Late Albian and Turonian (32±3°C). Negative carbon-isotopic shifts at the end of the early Maastrichtian and the Cretaceous-Tertiary boundary time seem to be connected with the fall of temperature and eventual reduction of oxygen content in the atmosphere and hydrosphere.

Key words: Maastrichtian, invertebrates, isotopic paleotemperatures, carbon-isotopic anomalies, Pacific and Atlantic oceans

1. INTRODUCTION

It is well known that greenhouse conditions prevailed during the Cretaceous and earliest Cenozoic. There is definite evidence showing the sharp decline of seawater temperatures in middle and high latitudes in the Maastrichtian (Lowenstam and Epstein, 1954; 1959; Bowen, 1969; Douglas and Savin, 1973; 1975; Teiss and Naidin, 1973; Boersma

and Shackleton, 1981; Barrera et al., 1987a; D'Hondt and Arthur, 1995; 1996; Huber et al., 1995; MacLeod and Huber, 1996; Zakharov et al., 1996; 1999; Jenkyns et al., 2004). On the basis of data on isotopic composition of calcified remains of various marine organisms (foraminifera, bivalves, nautiloids, ammonoids and belemnites), sea-surface water temperatures in the Antarctic during early Maastrichtian have been shown at 4.0–14.8°C (Barrera, 1994; Pearson et al., 2001). Some authors suggest the presence of early Maastrichtian glaciation in the Southern Hemisphere (Barrera et al., 1987a; Barrera, 1994; Miller et al., 1999; MacLeod and Huber, 2001; Huber et al., 2002).

At the same time, contradictory evidence is now available for the climatic conditions in the low latitudes during the Maastrichtian. For example, paleotemperatures estimated from early and middle Maastrichtian planktic foraminifera in the tropics fluctuate usually from 14.5 to 19.4°C and very rarely reach 20–21°C (Douglas and Savin, 1971; 1973; Boersma and Shackleton, 1981). But D'Hondt and Arthur (1996) and Yiliz and Özdemir (1999) recently obtained higher temperatures from some tropical and subtropical foraminiferal species, and Wilson and Opdyke (1996) also calculated higher paleotemperatures for early Maastrichtian tropics from isotopic compositions of well-preserved (aragonite containing) rudist bivalves.

The problem of “cool tropic paradox” was first discussed in detail by D'Hondt and Arthur (1995; 1996), who considered that the apparent incompatibility of isotopic interpretations with greenhouse climate simulations and some fossil interpretations constitutes this problem. According to D'Hondt and Arthur (1996), Maastrichtian tropical paleo-sea surface temperatures might be lower than those of the present (–6.4–7.4°C below the modern value). Based on the low Maastrichtian latitudinal sea surface temperature gradients for the Maastrichtian, these authors suggested that meridional heat transport was greater during the latest Cre-

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taceous than it is today. Supporting isotopic results, they challenged the applicability of traditional interpretations of “greenhouse” simulations of Late Cretaceous and Paleogene climate conditions.

However, one explanation (Schrag, 1999; Pearson *et al.*, 2001) is that unusual oxygen isotopic composition of tropical deep-sea surface water during some Cretaceous ages, Late Eocene and Oligocene is due to diagenesis of microfossils (recrystallization in the cold sea-floor environments). Wilson and Norris (2001) and Wilson *et al.* (2002) proposed the possibility that Albian and Turonian deep-sea surface temperatures in the tropical Atlantic underwent significant variations with peak temperatures about $32\pm 3^\circ\text{C}$ for the Late Albian and Turonian.

This paper focuses on the paleotemperature fluctuations during the Maastrichtian to estimate equatorial deep-sea surface temperatures using some oblique evidences: recently published (Cochran *et al.*, 2003) and original mollusk $\delta^{18}\text{O}$ results from the southern Western Interior Seaway (WIS) of North America and also our new and previously published isotopic data in other regions including the Koryak Upland, South Sakhalin and areas covered by Deep Sea Drilling Project works. We also use previous results based on some Recent data from Albian, Turonian and Late Maastrichtian foraminifera (D’Hondt and Arthur, 1996; Wilson and Norris; 2001; Wilson *et al.*, 2002) and Eocene foraminifera and bivalves (Kobashi *et al.*, 2001; 2004).

2. METHODS

The following were used to determine diagenetic alteration in the biogenic carbonates investigated: (1) visual signs, (2) percentage of aragonite in the structure when analyzing shells originally composed of 100% aragonite (Zakharov *et al.*, 1975); (3) degree of integrity of microstructures determined under a scanning electron microscope (SEM) when ammonoid shell aragonite and foraminiferal test calcite were investigated. To find signs of alteration in well-preserved foraminiferal tests and brachiopod shells, a preliminary luminescent test using a JXA-5A microanalyzer was used, as recommended by Grossman *et al.* (1991).

The results of X-ray diffraction analyses and microscopic observations reveal that most of the aragonitic mollusk samples from the Cretaceous of the Koryak Upland, Sakhalin, Tennessee and South Dakota, carefully selected after visual inspection, seem to be suitable for distinguishing their original oxygen and carbon isotopic records. New data on Recent and fossil photosymbiotic foraminifera (Houston and Huber, 1998; Houston *et al.*, 1999) show that their $\delta^{18}\text{O}$ values are more negative and their $\delta^{13}\text{C}$ are more positive than in existing asymbiotic taxa; accordingly, we have not used known photosymbiotic species in our analyses.

Oxygen and carbon isotope measurements were made using a Finnigan MAT-252 mass spectrometer at the Ana-

lytical Center of the Far Eastern Geological Institute, Vladivostok. The laboratory gas standard was calibrated relative to calcite NBS (National Bureau of Standards) 19 and equals $1.8\pm 0.10\text{‰}$ for oxygen relative to VPDB (Vienna Pee Dee belemnite) and $-0.75\pm 0.10\text{‰}$ for carbon. Reproducibility of replicate standards was always better than 0.10‰ . In calculating the temperatures, since icecaps were not present during most of the Cretaceous Period, a $\delta^{18}\text{O}$ of -1.2‰ VPDB (equivalent to -1.0‰ Standard Mean Ocean Water [SMOW]) was thought to be appropriate (Savin, 1977). Two scales were used for paleotemperature calculation: those of Anderson and Arthur (Epstein *et al.*, 1953; Anderson and Arthur, 1983) for calcite material and Grossman and Ku (1986) for aragonitic material. A total of 97 samples of Maastrichtian biogenic carbonates were investigated using mass spectrometer. X-ray analyses were carried out using a DRON-3 diffractometer following the method of Davis and Hooper (1963).

3. RESULTS

Our isotopic results on Maastrichtian fossils and some published data on this topic were obtained from semipelagic and pelagic (ammonoids and belemnites), benthic (benthic foraminifera, brachiopods, bivalves, and gastropods) and planktic (planktic foraminifera) forms. Ammonoids and belemnites, as it is known, are fossil cephalopod groups and their ecology is not so well investigated now. According to new isotopic evidence, Late Cretaceous ammonoid shells were most likely secreted in near-bottom conditions where the animals spent most of their lives in shallow marine basins (Smyshlyaeva *et al.*, 2002, Zakharov *et al.*, 2003a; 2005; Moriya *et al.*, 2003). On the other hand, belemnites (extinct coleoids), like Recent *Nautilus*, engaged in significant short-term vertical migrations in the water column (Smyshlyaeva *et al.*, 2004). Dating of individual sections described below was made mainly on the basis of known published data on ammonoid and foraminiferal biochronology.

3.1. Tennessee State, USA

The age of Coon Creek fauna is considered to be Late Campanian now (Larson, 2003). The presence of *Cirroceras conradi* (Morton) more precisely places the age at the equivalent age of *Baculites reesesidei* thought *Baculites jensenii* Zones of the Western Interior (K. Tanabe, pers. communication) although “*Jeletzkytes*” (=Hoploscapites) and *Solenoceras* are common for both the Campanian and the Maastrichtian. Samples from the uppermost Campanian of the WIS were analyzed for comparative purposes. $\delta^{18}\text{O}$ values in the three analyzed latest Campanian aragonitic bivalve shells of *Linotrignia (Linotrignia) thoracica* (Morton) from the Coon Creek Formation (Landman and Waage, 1993) (K. Tanabe’s collection) vary between -1.3 and -0.3‰ ,

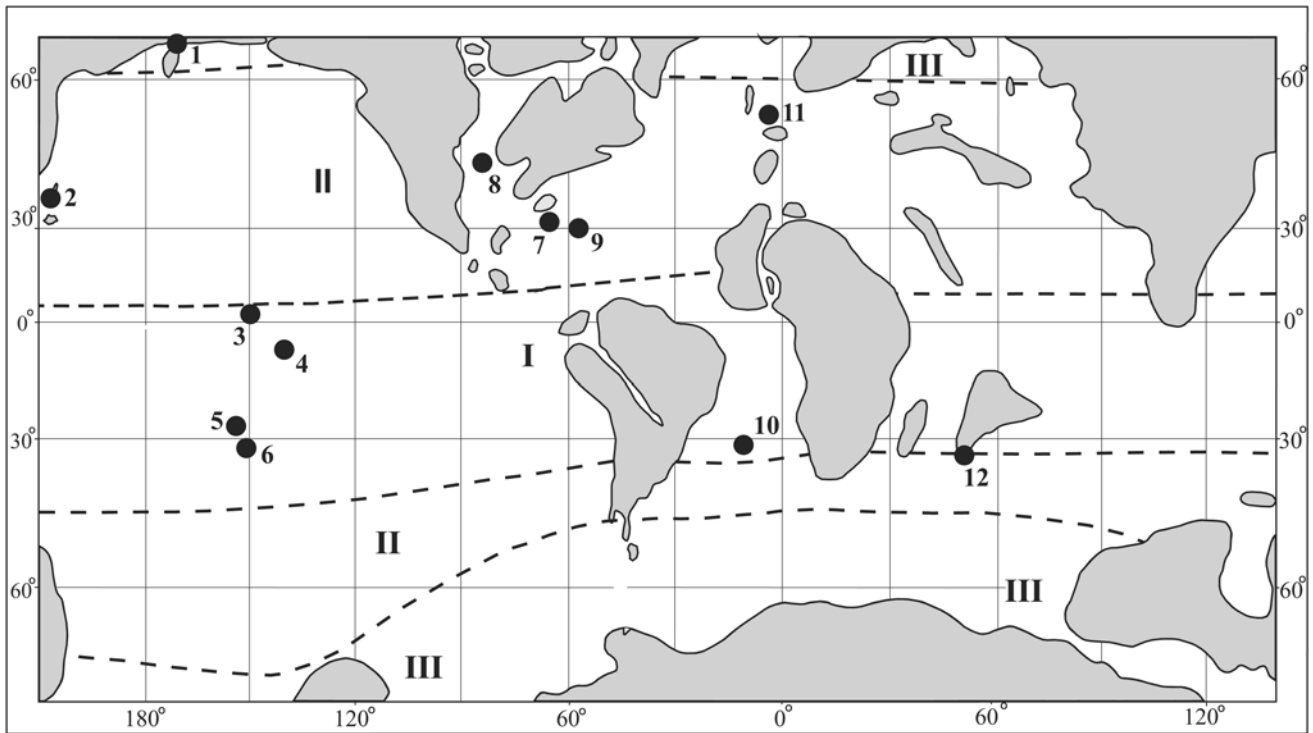


Fig. 1. Paleogeographic map for the Latest Cretaceous showing sampling locations in: (1) Tundrovaya River (Koryak Upland) (2) Naiba River (Sakhalin), (3) DSDP Hole 305(Shatsky Rise, Central Pacific), (4) Fedorov guyot, Magellan Rise, (5) DSDP Hole 289 (Ontong Plateau, South Pacific), (6) DSDP Hole 288A (Ontong Plateau, South Pacific), (7) Tennessee (USA), (8) South Dakota (USA), (9) South Netherlands and (12) Ariyalur, Trichinopoly area (Southern India). Climatic Zones I- Tropical-Subtropical, II- Subtropical, III- Warm-Temperate.

(9) DSDP Hole 390A, (10) DSDP Hole 516F, (11) South Netherlands

Table 1. Carbon and oxygen isotope analyses of aragonite- preserved trioniid bivalve shells from the Upper Campanian of the Tennessee State, USA (K. Tanabe’s collection)

Sample Species	Formation	Location (H, in mm)	Diagenetic alteration			$\delta^{13}\text{C}$ (VPDB), ‰	$\delta^{18}\text{O}$ (VPDB), ‰	T, C
			Original aragonite, %	Admixture, %	Colour			
T1-1 <i>Linotrigonia (L.) thoracica</i> (Morton)	Coon Creek	45	100	0	White	2.0	1.0	19.8
T1-2 <i>Linotrigonia (L.) thoracica</i> (Morton)	Coon Creek	20	100	0	White	1.7	-0.3	16.7
T1-3 <i>Linotrigonia (L.) thoracica</i> (Morton)	Coon Creek	30	100	0	White	1.4	-1.3	21.2

which correspond to paleotemperatures of 16.7–21.2°C (Fig. 1, Table 1). $\delta^{13}\text{C}$ values are positive (1.4–2.0‰).

3.2. Tundrovaya River, Koryak Upland

The middle Pillalvayam Formation of the Mamet River basin in the Koryak Upland is characterized by a Maastrichtian ammonite *Pachydiscus japonicus* Matsumoto (Perghament, 1961). Isotopic analyses were carried out on silvery white shells of rhynchonellid brachiopod with well-preserved fibrous structure recovered from the lower portion of the Maastrichtian exposed at the right bank of the Tundrovaya River (61°42’57’’N, 164°12’33’’E)(Fig. 1). They were discovered in medium-grained sandstone (80 m

in thickness) associated with corals and a bivalve *Acila (Truncacila)* sp. (Fig. 2) (Zakharov et al., 2002). The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values for rhynchonellid brachiopod samples range from -1.2 to 0.5‰ (which corresponds to paleotemperatures of 10.2–16.9°C) and from 0.6 to 1.8‰, respectively. Higher temperatures were obtained from mollusk shells found in the underlying sediments (Zakharov et al., 2002).

3.3. Naiba River basin, Sakhalin

Well-preserved Maastrichtian ammonoids and brachiopods from the Krasnoyarka Formation of the Naiba River basin (Naiba, Krasnoyarka and Sary Rivers), South Sakhalin were used for stable isotopic analysis (Fig. 1) (Zakharov

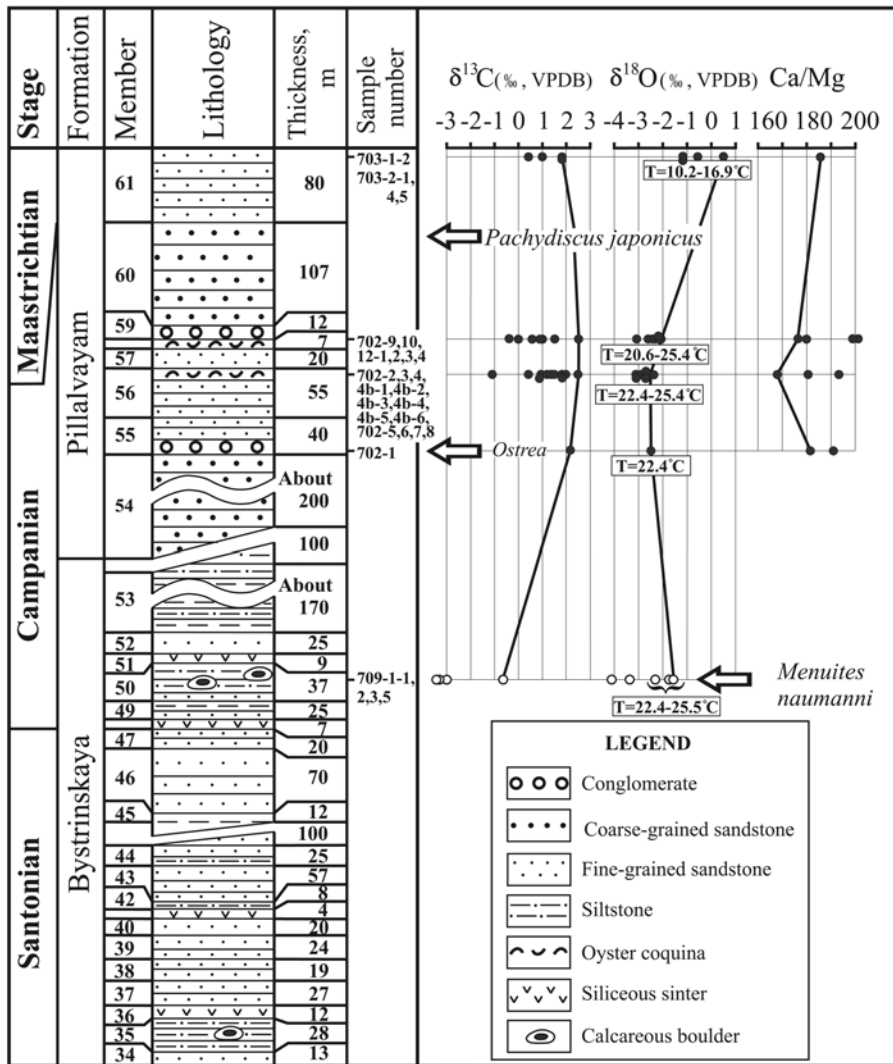


Fig. 2. The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of early Early Maastrichtian brachiopod shells from the Pillalvayamskaya Formation of the Tundrovaya River section, Penzhinskaya Guba. Open circles indicate calcite, filled ones aragonite.

et al., 1984a, 1996, 1999).

3.3.1. Naiba River

The following Maastrichtian succession toward the top of the sequence is exposed along the Naiba River, south Sakhalin, Far East Russia.

Pachydiscus - Pleurogrammatodon bykovensis Beds

7. Grayish green, medium- and fine-grained sandstone, 94–104 m thick and contains a Maastrichtian palynological assemblage (Kalishevich et al., 1981).

Unexposed interval (over 50 m in thickness).

6. Interbedding of coarse-, medium- and fine grained tuffaceous sandstone, rarely siltstone, 132 m thick, and yields foraminifera, bivalves, ammonoids (*Gaudryceras* sp., *Diplomoceras* sp., *Pseudoxybeloceras lineatum* (Gabb), *Pachydiscus* (*Pachydiscus*) sp., *P.(P.)* aff. *gollevillensis* d’Orbigny and *P. (Neodesmoceras) japonicus* Matsumoto) (Vereschagin, 1977; Kalishevich et al., 1981; Zakharov et al., 1984a).

5. Dark gray sandy siltstone with calcareous-marly boulders, 6.5 m thick (bed 5–4: samples ?KL111–1,2,3). It contains corals, brachiopods, bivalves, scaphopods, gastropods, ammonoids (*Phyllopachyceras forbesianum* (d’Orbigny)) and fish (scales) (Zakharov et al., 1984a, b).

4. Dark gray mudstone with calcareous-marly boulders and lenses, and also wood remains, 25 m thick (bed 5–6: sample 141-952-65). It yields brachiopods, ammonoids (*Tetragonites popetensis* Yabe, *Gaudryceras tenuiliratum* Yabe and *Polyptychoceras pseudogaultinum* (Yokoyama)) and plant remains.

3. Dark gray mudstone with calcareous-marly boulders and lenses, 23 m thick (bed 5–7: samples 105-952-32, 106-952-35, 106-952-42; 5-9, 114-952-50). It is rich in corals, brachiopods, bivalves, gastropods and ammonoids (*Pachydiscus* (*Pachydiscus*) sp., *Pachydiscus* (*Neodesmoceras*) *japonicus* Matsumoto) (Zakharov et al., 1984a).

Zelandites varuna Beds

2. Dark gray mudstone with calcareous-marly boulders, 5

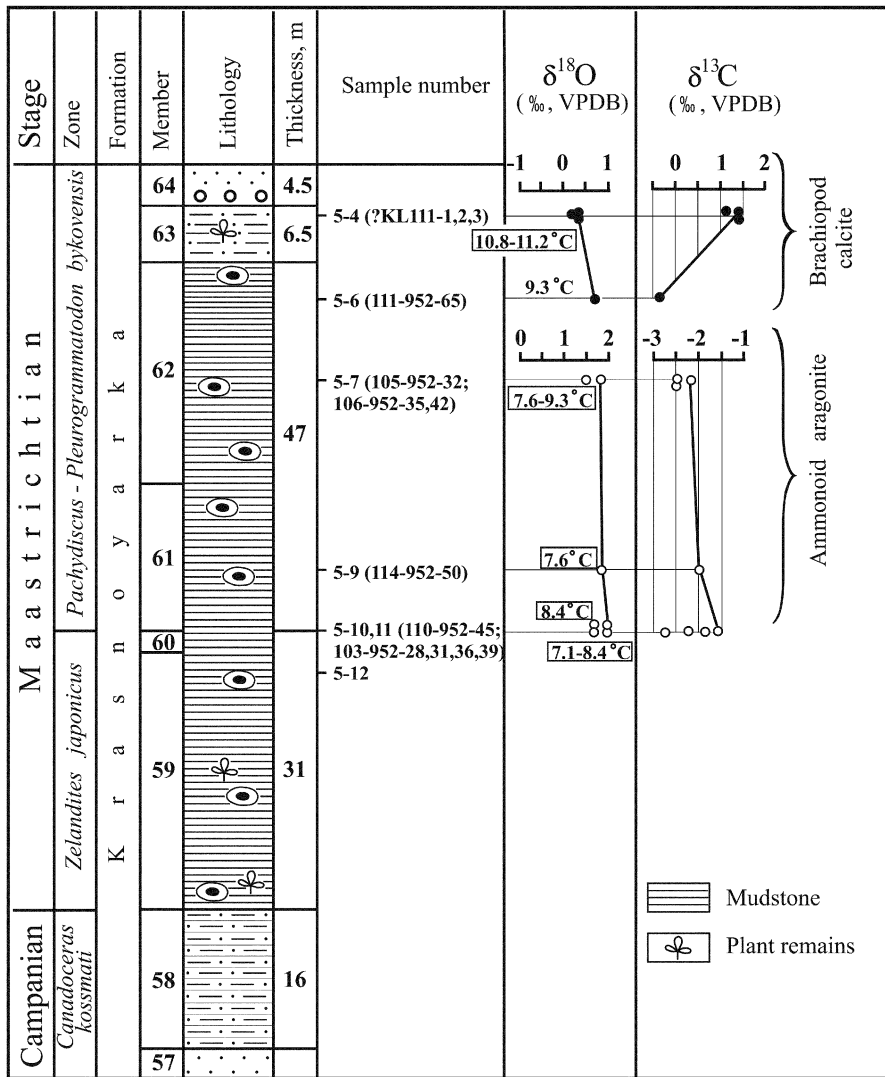


Fig. 3. The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of Maastrichtian brachiopod and ammonoid shells from the Krasnoyarsk Formation of the Naiba River section, South Sakhalin. Open circles indicate calcite, filled ones aragonite. Other designation as in Fig. 2.

m thick (bed 5–10: sample 110-952-45; bed 5-11: samples 103-952-28,31,36,39) and contain foraminifera and ammonoids (*Zelandites japonicus* Matsumoto and *Pachydiscus* (*Pachydiscus*) cf. *gollevillensis* d'Orbigny) (Zakharov et al., 1984a, b).

1. Dark gray mudstone with calcareous-marly boulders and lenses and rare layers of fine-grained sandstone and gravelstone, 26 m thick, and yields bivalves, gastropods, ammonoids (*Zelandites japonicus* Matsumoto) and plant remains (Zakharov et al., 1984a, b).

The thickness of the Maastrichtian in the Naiba River section is about 370 m.

A few gigantic Lower Maastrichtian ammonoid shells were used for isotopic analyses (Fig. 3) were collected from two horizons in the upper part of the *Zelandites varuna* Beds, beds 5–11 and 5–10, along the Naiba River (47°23'14"N, 142°29'23"E). From the lower horizon, three aragonitic (99±1%) *Pachydiscus* (*Pachydiscus*) cf. *gollevillensis* (d'Orbigny) shells with no $\alpha\text{-SiO}_2$ were collected. Three samples (103-952-28,31,36) were taken from the lateral wall and only one

(108-952-39) from a septum. Their $\delta^{18}\text{O}$ values are very high: fluctuate from 1.6 to 1.9‰, which correspond to paleotemperatures of 7.1–8.4°C; their $\delta^{13}\text{C}$ values fluctuate from 2.8 to –1.1‰. From the upper horizon, only one aragonitic (98±2%) *Pachydiscus* (*Pachydiscus*) cf. *gollevillensis* (d'Orbigny) shell with no $\alpha\text{-SiO}_2$ was collected. Its lateral wall, sample 110-952-45, is also characterized by a high $\delta^{18}\text{O}$ value (1.6‰), corresponding to a paleotemperature of 8.4°C. Its $\delta^{13}\text{C}$ value is –2.0‰.

Analyzed middle Maastrichtian fossils were collected from four levels of the lower part of the *Pachydiscus - Pleurogrammatodon bykovensis* Beds (beds 5-9, 5-7, 5-6 and 5-4). From the first horizon, only one gigantic aragonitic (98-99±2%) *P. (Neodesmoceras) japonicus* Matsumoto shell with no $\alpha\text{-SiO}_2$ was investigated. Its $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values (1.8 and –2.0‰, respectively) from the lateral wall, sample 114-952-50, are similar with previous ones, therefore its calculated paleotemperature is also low (7.6°C). From the second level, three gigantic aragonitic (97±3%) shells of

P. (Pachydiscus) sp. with no α -SiO₂ were investigated. Their $\delta^{18}\text{O}$ values from the lateral walls (samples 105-952-32, 106-952-35,42) range from 1.4 to 1.8‰, which correspond to paleotemperature of 7.6–9.3°C, their $\delta^{13}\text{C}$ values range from –2.8 to –1.1‰. From the third level, a few small silvery-white brachiopods shells of *Orbirhynchia* sp. with well-preserved fibrous structure, without any visual diagenetic alteration (sample 141-952-65) were used. $\delta^{18}\text{O}$ value of the sample is 9.3‰, which corresponds to paleotemperature of 9.3°C; its $\delta^{13}\text{C}$ value is –0.6‰. From the fourth level, three silvery-white rhynchonellid brachiopods shells with well-preserved fibrous structure (samples KL11-1, 2,3) were investigated. Their $\delta^{18}\text{O}$ values are somewhat lower than those of invertebrate shells from underlying strata (0.2 to 0.3‰, which correspond to paleotemperatures of 10.8–11.2°C). Their $\delta^{13}\text{C}$ values change from 1.1 to 1.4‰.

3.3.2. Krasnoyarka and Sary rivers

In the Krasnoyarka River section, the Upper Maastrichtian portion of the Krasnoyarka Formation is represented by gray medium-grained sandstone and tuffaceous siltstone. They contain large calcareous boulders, containing remains of foraminifera, corals, brachiopods, bivalves, and ammonoids (*Gaudryceras* sp., *Diplomoceras* sp., *Pachydiscus* (*Pachydiscus*) cf. *neubergicus* Hauer, *P.(P.)* aff. *P. gollevillensis* (d'Orbigny) and *P.(P.) subcompressus* Matsumoto) and palynological assemblage (Vereschagin, 1977; Kalishevich *et al.*, 1981; Zakharov *et al.*, 1984a). This level corresponds, apparently, to the member 7 of the upper Maastrichtian *Pachydiscus* - *Pleurogrammatodon bykovensis* Beds of the Naiba River section or some lower beds.

Some well-preserved brachiopod shells (Rhynchonellacea (no. KL10-6-2,3) and *Orbirhynchia* sp. (no. KL6) were collected by T.G. Kalishevich within the middle Maastrichtian interval of the *Pachydiscus*-*Pleurogrammatodon bykovensis* Beds (Krasnoyarka River, just below the Krasnoyarka Coal Mine, 47°26'26"N, 142°31'36"E) (Zakharov *et al.*, 1999). They were used by us for isotopic analyses. One well preserved brachiopod shell of *Orbirhynchia* sp. (no. KL6) from the same level collected by T.G. Kalishevich in the Sary River (47°19'38"N, 142°32'02"E) (Zakharov *et al.*, 1999) was used also for our analysis. $\delta^{18}\text{O}$ values of the three silvery-white brachiopod shells from the Krasnoyarka River retaining well-preserved fibrous structure range from 0.2 to 0.5‰, which corresponds to paleotemperatures of 10–11.8°C; their $\delta^{13}\text{C}$ values vary from 0.6 to 1.8‰. $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of a single well-preserved brachiopod *Orbirhynchia* sp. from the Sary River section are 1.0‰ (corresponds to paleotemperature of 8.1°C) and 0.0‰, respectively.

3.5. Hole 305, Shatsky Rise

The Hole 305 located at the southern part of the Shatsky

Rise, Central Pacific, at a depth of 2903 m, is the most representative Cretaceous section of the mentioned area (Fig. 1) (Basov and Vishnevskaya, 1991). Santonian-Maastrichtian sediments, 170 m thick, consist of foraminifera-nannofossil-dominant chalk, silt with boulders and interlayers of flint. The uppermost part of the Maastrichtian interval is represented by the Upper Maastrichtian Zone of *Abathomphalus mayaroensis* (cores 15 and 16) (Basov and Vishnevskaya, 1991; Caron, 1985; Sokolova, 1998). The middle part of the Maastrichtian interval consists of two middle Maastrichtian Zones: *Globotruncana gansseri* (above, core 17) and *G. aegyptica* (below, core 18). The lowermost part of the Maastrichtian interval is represented by the lower Maastrichtian Zone of *Globotruncana tricarinata* (= *Globotruncanella havanensis* Zone, cores 19 and 20) (Caron, 1985; Sokolova, 1998) (Fig. 4).

The lower part of the member (cores 23–28) is characterized by Campanian and Santonian foraminifera and Santon-Campanian radiolarians (Larson and Moberly, 1975; Basov and Vishnevskaya, 1991). The Maastrichtian is unconformity overlain by the Upper Paleocene Zone of *Globorotalia pusilla* (Basov and Vishnevskaya, 1991).

The three samples for isotope analyses were taken from planktic foraminifera tests from the Lower Maastrichtian *Globotruncana tricarinata* Zone (E.A. Sokolova's collection) (Table 2). The sample 32-305-18-3-25-28(A) is represented exclusively by well-preserved, hollow tests of *Heterohelix globulosa* (fraction 0.05-0.020 mm); in the sample 32-305-18-4-99-12 in addition to well-preserved tests of *H. globulosa* (Ehrenberg), two other species are present (*Hedbergella holmdelensis* (Olsson) and *Globotruncana arca* (Cushman) (fraction 0.05-0.315 mm and larger); sample 32-305-18-3-25-28 consists mainly of well-preserved tests of *Globotruncanita* (*G. subspinosa* (Pessagno), *G. stuartiformis* (Dalbiez)) and also *Globotruncana arca* (Cushman) (fraction 0.05-0.315 mm and larger). In spite of very different species composition of investigated samples, all of them are characterized by very similar oxygen- and carbon-isotopic composition: relatively low $\delta^{18}\text{O}$ values ranging from –1.3 to –0.9‰ and positive $\delta^{13}\text{C}$ values from 1.3 to 1.4‰. The oxygen isotopic values indicate that many tests of *Heterohelix globulosa* were formed at a paleotemperature of 17.1°C, but some other early Maastrichtian planktic foraminifera at 15.5–16.0°C.

The obtained data are in good agreement with isotopic results on early Maastrichtian nannofossils of the same Site (Douglas and Savin, 1975): their $\delta^{18}\text{O}$ values fluctuate from –1.1 to –0.66‰, which correspond to paleotemperatures of 14.5–16.5°C; $\delta^{13}\text{C}$ values are very high (about 3‰). Early Maastrichtian benthic foraminiferal tests of the Site 305, on contrary, are characterized by very low $\delta^{18}\text{O}$ values (about 0.1‰), indicating a paleotemperature of 11.5°C, and positive $\delta^{13}\text{C}$ values (about 0.3‰) (Douglas and Savin, 1975). Similar results were obtained for the middle and upper Maas-

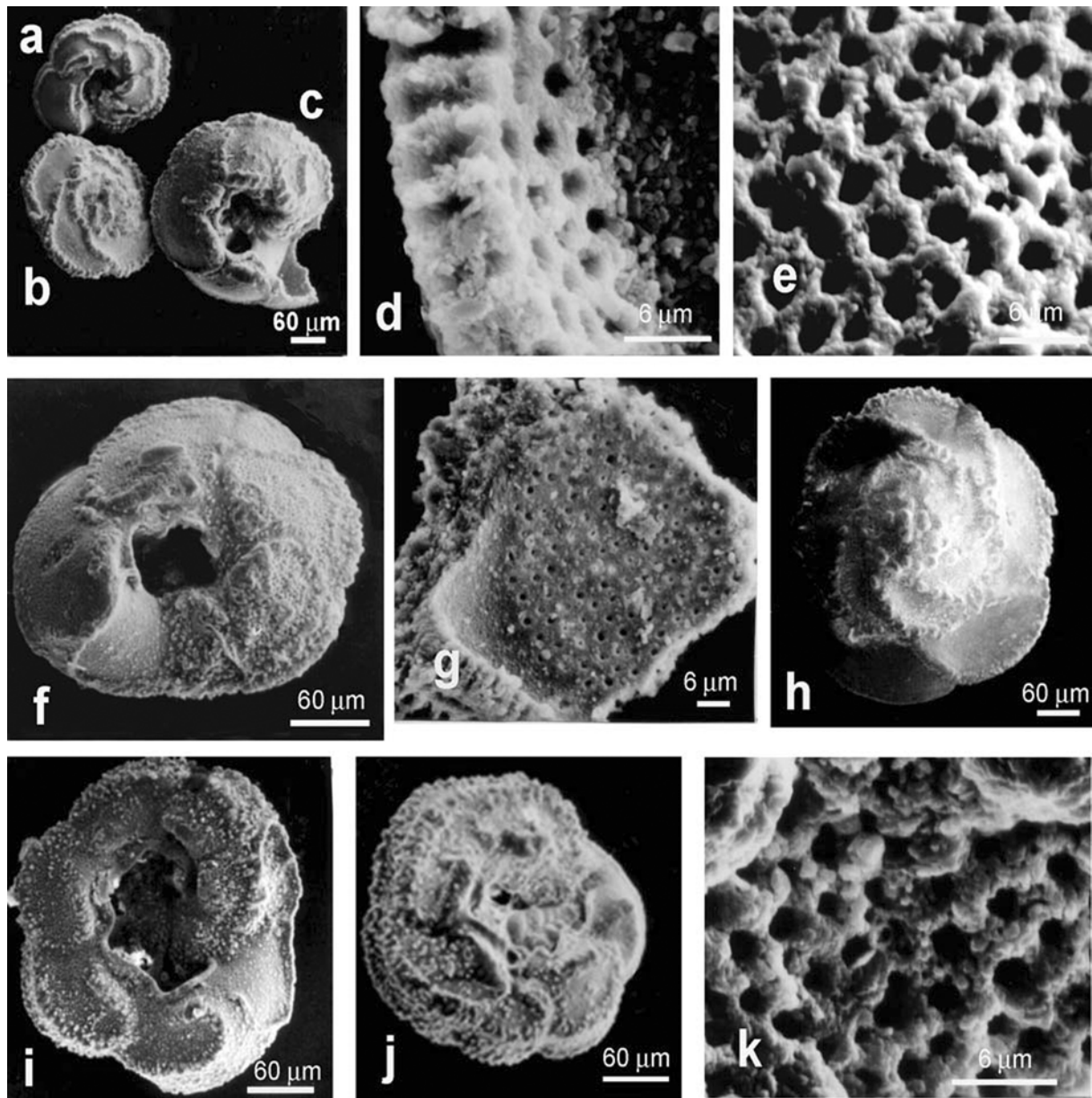


Fig. 4. Early Maastrichtian planktic foraminifera from the Pacific: a–b—*Globotruncana linneana* (de Orbigny) shells (Hole 305, Shatsky Rise, Pacific), a – sample DVGI 32-305-18-4(a); a – sample DVGI 32-305-18-4(b); c - *Globotruncantina stuarti* (de Lapparent) shell (Hole 305, Shatsky Rise, Pacific), sample DVGI 32-305-18-4(c); d–e – fractured *Globotruncana linneana* (de Orbigny) shell: d - sample DVGI 44-390A-142 (Hole 390A, Shatsky Rise, Pacific), e – sample DVGI 32-305-18-4 (Hole 305, Shatsky Rise, Pacific), f – *Globotruncantina stuarti* (de Lapparent) shell, foramen side, sample DVGI-32-305-18-1 (Hole 305, Shatsky Rise, Pacific), g – fractured *Globotruncantina stuarti* shell, sample DVGI 32-305-15-3 (Hole 305, Shatsky Rise, Pacific), h – *Globotruncana arca* (Cushman) shell, spiral side (Hole 390A, North Atlantic), i–k – *Globotruncana linneana* (de Orbigny) (Hole 390A, North Atlantic): i- shell, foramen side, sample DVGI 44-390A-142; j- shell, spiral side, sample DVGI 44-390A-14, k- surface of the sample DVGI 44-390A-142.

trichtian of the Site 305 on the basis of data from benthic foraminifera and nannofossils (Douglas and Savin, 1975).

3.6. Hole 289, Ontong Java Plateau

The Cretaceous unit in Hole 289 (Fig. 1) consists of foraminifera-nannofossil-dominant chalk with boulders of flint, nannofossil limestone with interlayers of tuff and flint

boulders. Cretaceous sediments lay on fresh tholeiitic basalt and are overlain by Paleocene carbonates. Aptian, lower-middle Campanian, and late Campanian - Maastrichtian assemblages of foraminifera and nannofossils were recognized in the section (Basov and Vishnevskaya, 1991). The absence of the assemblages from the Albian-Santonian interval and reduced thickness (110 m) of the Cretaceous are explained by significant erosion during Albian to Santonian time interval.

Table 2. Carbon and oxygen isotope analyses of well preserved planktic foraminiferal shells from the lower and middle Maastrichtian of the Central Pacific and North Atlantic

Sample	Locality, paleolatitude	Stage	Zone	Diagenetic alteration		Species, rock	Fraction size, mm	$\delta^{13}\text{C}$	$\delta^{18}\text{O}$	T, °C
				Original calcite, %	$\alpha\text{-SiO}_2$, %			(VPDB), ‰	(VPDB), ‰	
30-288A-9-4-100-102(A)	Hole 288A, Pacific, paleolatitude of 35°S	Upper middle Maastrichtian	Globotruncana contusa	-	-	<i>Globotruncanita stuarti</i> (de Lapparent)	>0.315	2.8	-1.3	17.0
30-288A-9-4-100-102(B)	Hole 288A, Pacific, paleolatitude of 35°S	Upper middle Maastrichtian	Globotruncana contusa	-	-	<i>Pseudotextularia elegans</i> (Rzehak), <i>Globotruncanita stuarti</i> (de Lapparent)	0.2-0.315	2.3	-1.8	19.4
30-288A-9-4-100-102(C)	Hole 288A, Pacific, paleolatitude of 35°S	Upper middle Maastrichtian	Globotruncana contusa	-	-	<i>Pseudotextularia elegans</i> (Rzehak), <i>Globotruncana arca</i> (Cushman)	0.05-0.2	2.5	-1.4	17.7
30-289-124-1-133-135(bulk)	Hole 289, Pacific, paleolatitude of 35°S	Upper middle Maastrichtian	Globotruncana gansseri	100	0	<i>Pseudotextularia elegans</i> (Rzehak), <i>Globotruncanita stuarti</i> (de Lapparent), <i>G. stuartiformis</i> (Dalbiez), <i>Globotruncana arca</i> (Cushman)	-	2.6	-1.4	17.7
30-289-124-2-146-148(A)	Hole 289, Pacific, paleolatitude of 35°S	Upper middle Maastrichtian	Globotruncana gansseri	-	-	<i>Globotruncanita stuartiformis</i> (Dalbiez), <i>Pseudoguembelina costulata</i> (Cushman)	>0.315	2.7	-1.4	17.7
30-289-124-2-146-148(B)	Hole 289, Pacific, paleolatitude of 35°S	Upper middle Maastrichtian	Globotruncana gansseri	-	-	<i>Globotruncanita stuartiformis</i> (Dalbiez), <i>G. stuarti</i> (de Lapparent), <i>Globotruncana arca</i> (Cushman)	0.2-0.315	2.8	-1.1	16.5
30-289-124-2-146-148(C)	Hole 289, Pacific, paleolatitude of 35°S	Upper middle Maastrichtian	Globotruncana gansseri	-	-	<i>Pseudoguembelina costulata</i> (Cushman), <i>Pseudotextularia elegans</i> (Rzehak)	0.05-0.2	2.6	-1.3	17.0
35D206-4B	Pacific, IOAN	Maastrichtian	<i>Zelandites</i> aff. <i>japonicus</i> Beds	-	-	White chalk	-	2.7	0.7	-
32-305-18-3-25-28(A)	Hole 305, Pacific, paleolatitude of 10°N	Lower Maastrichtian	Globotruncana tricarinata	100	0	<i>Heterohelix globulosa</i> (Ehrenberg)		1.4	-1.3	17.1
32-305-18-3-25-28(bulk)	Hole 305, Pacific, paleolatitude of 10°N	Lower Maastrichtian	Globotruncana tricarinata	100	0	<i>Globotruncana arca</i> (Cushman), <i>Globotruncanita subspinoso</i> (Pessagno), <i>G. stuartiformis</i> (Dalbiez)		1.3	-1.0	16.0
32-305-18-4-99-12(bulk)	Hole 305, Pacific, paleolatitude of 10°N	Lower Maastrichtian	Globotruncana tricarinata	100	Weak trace	<i>Heterohelix globulosa</i> (Ehrenberg), <i>Hedbergella holmdelensis</i> (Olsson), <i>Globotruncana arca</i> (Cushman)		1.3	-0.9	15.5
44-390A-14-1-83-86(A)	Hole 390A, Atlantic, paleolatitude of 30°N	Lower Maastrichtian	Globotruncana tricarinata	100	0	<i>Globotruncana linneiana</i> (de Orbigny), <i>G. arca</i> (Cushman)		2.0	-0.8	15.1
44-390A-14-1-83-86(bulk)	Hole 390A, Atlantic, paleolatitude of 30°N	Lower Maastrichtian	Globotruncana tricarinata	100	0	<i>Globotruncana linneata</i> (de Orbigny), <i>G. arca</i> (Cushman), <i>Planoglobulina multicamerata</i> de Klasz, <i>Globigerinelloides volutus</i> (White), <i>Heterohelix globulosa</i> (Ehrenberg)		2.0	-0.5	14.1
44-390A-14-2-86-90(bulk)	Hole 390A, Atlantic, paleolatitude of 30°N	Lower Maastrichtian	Globotruncana tricarinata	100	0	<i>Globotruncana linneiana</i> (de Orbigny), <i>G. arca</i> (Cushman), <i>G. falsostuarti</i> (Sigal), <i>Globotruncanita stuartiformis</i> (Dalbiez), <i>Rosita fornicata</i> Plummer, <i>Planoglobulina multicamerata</i> de Klasz, <i>Heterohelix globulosa</i> (Ehrenberg)		2.0	-0.6	14.4

Four samples of planktic foraminiferal tests from the lower part of the middle Maastrichtian (*Globotruncana gansseri* Zone) represented by different fractions were analyzed (Table 2). Preserved shells were cleaned by ultrasound, but a small amount of secondary calcite exists inside them. $\delta^{18}\text{O}$ value of the largest fraction, >0.315 mm (sample 30-289-124-2-146-148A: mainly *Globotruncanita stuartiformis* (Dalbiez) and *Pseudoguembelina costulata* (Cushman)), is -1.4‰ , which corresponds to paleotemperature of 17.7°C ; its $\delta^{13}\text{C}$ is also high (2.7‰). $\delta^{18}\text{O}$ value of the middle fraction, 0.2–0.315 mm (sample 30-289-124-2-146-148B: mainly *Globotruncanita stuartiformis* (Dalbiez), *G. stuarti* (de Lapparent), *Globotruncana arca* (Cushman)), is -1.1‰ ; its $\delta^{13}\text{C}$ is 2.8‰ . $\delta^{18}\text{O}$ value of the smallest fraction, 0.05–0.2 mm (sample 30-289-124-2-146-148C: mainly *Pseudoguembelina costulata* (Cushman) and *Pseudotextularia elegans* (Rzehak)) is -1.3‰ , which corresponds to a paleotemperature of 17.0°C ; its $\delta^{13}\text{C}$ is 2.6‰ . Sample 30-289-124-1-133-135 is represented by a mixture of the above-mentioned fractions (0.05–0.315 and larger) and consists mainly of *Pseudotextularia elegans* (Rzehak), *Globotruncanita stuartiformis* (Dalbiez), *G. stuarti* (de Lapparent), and *Globotruncana arca* (Cushman). Its $\delta^{18}\text{O}$ value is -1.4‰ ; which corresponds to a paleotemperature of 17.7°C ; its $\delta^{13}\text{C}$ is 2.6‰ .

3.7. Hole 288A, Ontong Java Plateau

Cretaceous sediments of the southeastern part of the Ontong Java Plateau, Southwest Pacific (Fig. 1), are represented by foraminifera-nannofossil-dominant chalk and limestone with flint, mudstone and siltstone interlayers (more than 500 m) (Packham and Andrews, 1975).

Three well-preserved planktic foraminiferal test samples (E.A. Sokolova's collection) (Table 2) were investigated from the upper part of the middle Maastrichtian *Globotruncana contusa* Zone, after cleaning by ultrasound. Only hollow specimens without secondary calcite inside were used for isotopic analysis.

The samples are represented by three different shell fractions. Sample 30-288A-9-4-100-102(A), the largest fraction, >0.315 mm, consists of exclusively by *Globotruncanita stuarti* (de Lapparent). Sample 30-288A-9-4-100-102(B), the middle fraction, 0.2–0.315 mm, is represented by a combination of mainly *Pseudotextularia elegans* (Rzehak) and *Globotruncanita stuarti* (de Lapparent). Sample 30-288A-9-4-100-102(C), the smallest fraction, 0.05–0.2 mm, represented mainly by *Pseudotextularia elegans* (Rzehak) and *Globotruncana arca* (Cushman). Samples of different fractions show very similar results: their $\delta^{18}\text{O}$ values fluctuate from -1.8 to -1.4‰ , which correspond to paleotemperatures of 17.0 – 19.4°C ; $\delta^{13}\text{C}$ values are high (2.3 to 2.8‰) (Table 2).

3.8. Fedorov (IOAN) guyot, Magellan Rise

The Cretaceous biostratigraphy of the Magellan Rise and

neighboring areas have been investigated by a number of researchers on the basis of deep-sea drilling and dredging data (Douglas and Savin, 1971; 1973; Pletnev and Biryulina, 1989; Bogdanov et al., 1990; Basov and Vishnevskaya, 1991; Ablav et al., 1992; Melnikov et al., 1995a, b; Shkolnik et al., 1996). The total thickness of Mesozoic and Cenozoic sediments at the western part of the Fedorov guyot (Fig. 1) is about 800 m. Maastrichtian ammonites *Zelandites* aff. *japonicus* Matsumoto and Tetragonitidae gen. et sp. indet. have been given reported from this area ($14^\circ25.6'\text{N}$, $155^\circ40'\text{E}$) (Zakharov et al., 2003a).

Because the ammonoid shells of the sample 35D-206-4C are recrystallized, we only investigated the isotopic composition of the white chalk, containing ammonoid (bed of white chalk). Sample 35D-206-4C has a positive $\delta^{13}\text{C}$ value (2.7‰) (Table 2). Based on this fact, we correlated the bed of white chalk, containing ammonoids, with the *Globotruncana gansseri* Zone (Middle Maastrichtian) of the Pacific Holes 288A and 289, which is characterized by positive $\delta^{13}\text{C}$ values (2.3 – 2.8‰) of planktic foraminiferal tests. Recently, we discovered 54 belemnite rostra in the Maastrichtian limestone from the Magellan Rise; their isotopic composition will be investigated later.

3.9. South Dakota, USA

The Fox Hills Formation is represented by three members: Trail City (*Hoploscaphites nicolletii* Range Zone), Timber Lake (*Jeletzkytes nebrascensis* Range Zone) and Iron Lightning. Well-preserved ammonoids are known from the lower part (middle Maastrichtian Trail City Member) of the Fox Hills Formation west of the Missouri River in parts of Corson, Deway, and Ziebach counties, Southern Dakota. The Fox Hills Formation constitutes the marginal marine phase of a progradational sequence representing the final episode of withdrawal of the WIS, and contains the youngest Cretaceous marine faunas known from the WIS region (Landman and Waage, 1993). The formation includes 76–107 m of dominantly silty to sandy strata gradational downward into the Pierre Shale and upward into the nonmarine beds of the Hell Creek Formation. The Trail City Member consists of thinly interbedded clay and silt with numerous layers of calcareous boulders.

Three well-preserved silvery-cream and grayish-pink coloured bivalve *Protocardia* sp. shells (73–100% of aragonite) and five silvery-cream coloured aragonitic ammonoid shells (83–97% of aragonite) from the middle Maastrichtian Trail Member of the Fox Hills Formation of Deway and Corson counties (Y. Shigeta's collection) were taken for isotopic analysis (Fig. 1). Most well-preserved shells of *Protocardia* sp. (FH-2, PM14784-2 and PM14781-2: 97–100% of aragonite) are characterized by somewhat lower $\delta^{18}\text{O}$ values (-2.3 to -1.7‰) and higher $\delta^{13}\text{C}$ values (0.9 – 3.8‰) in comparison with ammonoid samples from the Trail City

Table 3. Carbon and oxygen isotope analyses of both aragonite-preserved and diagenetically altered bivalve and ammonoid shells from the middle Maastrichtian of the South Dakota, USA (Y. Shigeta's collection).

Sample	Species	County (formation, member)	Location (H, in mm)	Diagenetic alteration			$\delta^{13}\text{C}$ (VPDB), ‰	$\delta^{18}\text{O}$ (VPDB), ‰	T, °C
				Original aragonite, %	Admixture, %	Colour			
FH-2	Bivalve mollusk	Dewey (Fox Hills, Trail City)	10	100	0	Grayish-pink	3.8	-2.2	“25.1” (fresh-water influence)
PM14784-2	Bivalve mollusk	Dewey (Fox Hills, Trail City)	12	97±3	0	Silvery-cream	-1.8	-0.5	17.6
PM14781-2	Bivalve mollusk (shell fragment)	Dewey (Fox Hills, Trail City)	>1.8	97±3	0	Silvery-cream	3.7	-2.3	“25.2” (fresh-water influence)
PM14784-1	<i>Hoploscaphites nicolleti</i> (Morton), microconch	Dewey (Fox Hills, Trail City)	Two small shells (7 and 8 mm)	93±3	0	Silvery-cream	-1.8	-0.5	17.6
PM14781-1	<i>Jeletzkytes spedeni</i> Landman and Waage	Dewey (Fox Hills, Trail City)	26	97±3	0	Silvery-cream	0.0	-1.4	21.3
PM14787	<i>Discoscaphites gulosus</i> (Morton), microconch	Dewey (Fox Hills, Trail City)	21	95±3	0	Silvery-cream	-5.2	-1.2	20.7
DI-5	Bivalve mollusk	Corson (Fox Hills, Trail City)	40	73±3	0	White	-3.9	-3.3	[29.7] (diagenetically altered)
DK-3	<i>Discoscaphites gulosus</i> (Morton), microconch	Corson (Fox Hills, Trail City)	17	83±3	0	White-cream	0.6	-4.9	[36.7] (diagenetically altered)

Table 4. Carbon and oxygen isotope analyses of ankerite-containing planktic foraminiferal shells from the Lower and Upper Maastrichtian of the South Atlantic (Hole 516F).

Sample	Locality	Stage	Zone	Diagenetic alteration			$\delta^{13}\text{C}$ (VPDB), ‰	$\delta^{18}\text{O}$ (VPDB), ‰	“T”, °C
				Original calcite, %	Ankerite, %	$\alpha\text{-SiO}_2$, %			
72-516F-100-2-34-38	Atlantic, paleolatitude of 30°S	Lower Maastrichtian	<i>Globotruncana tricarinata</i>	~50	~50	Trace	1.5	-4.0	29.6
72-516F-100-3-35-39	Atlantic, paleolatitude of 30°S	Lower Maastrichtian	<i>Globotruncana tricarinata</i>	~50	~50	Trace	1.5	-4.0	29.6
72-516F-92-4-109-113	Atlantic, paleolatitude of 30°S	Upper Maastrichtian	<i>Abathomphalus mayaroensis</i>	~60	~40	Weak trace	1.9	-4.1	30.2

Member of Deway County (Table 3). Paleotemperature calculated from bivalve is 22.8°C.

Sample PM14784-1 consists of material from two small cream coloured ammonoid shells (*Hoploscaphites nicolleti* (Morton), microconch). Its $\delta^{18}\text{O}$ value is -0.5‰ , which corresponds to a paleotemperature of 17.6°C. Its $\delta^{13}\text{C}$ value is negative (-1.8‰). The $\delta^{18}\text{O}$ value in the shell of *Jeletzkytes spedeni* Landman and Waage (PM14781-1) is -1.4‰ , which corresponds to paleotemperature of 21.3°C, its $\delta^{13}\text{C}$ value is 0.0‰. $\delta^{18}\text{O}$ value of the *Discoscaphites gulosus* (Morton), microconch shell (PM14787) is -1.2‰ , which corresponds to paleotemperature of 20.7°C, but its $\delta^{13}\text{C}$ value is very low (-5.2‰). One investigated shell of an ammonoid (*Discoscaphites gulosus* (Morton) (DK-3) from the Trail City Member of the Corson County and one bivalve shell (D1-5) from the same locality shows unusual low $\delta^{18}\text{O}$ values ranging from -4.9 to -3.3‰ (Table 3), suggesting diagenetic alteration.

We add the paleotemperature data by Cochran et al. (2003) for ten well-preserved ammonoid shells, two belemnite rostra and one gastropod shell from the lower Upper Maastrichtian *Jeletzkytes nebrascensis* Zone (fluctuated from 13.8° to 22.4°C, from 10.7° to 17.3°C and 17.6°C, respectively).

3.10. Hole 390A, North Atlantic

Hole 390A is located in the northwestern part of the Atlantic. The core interval 83-86 mm of Hole 390A corresponds to the lower Maastrichtian *Globotruncana tricarinata* Zone (Fig. 1 and 4). $\delta^{18}\text{O}$ value of the bulk sample of planktic foraminiferal tests of the mentioned interval (44-390A-14-1-83-86) is -0.5‰ , which corresponds to paleotemperature of 14.1°C; $\delta^{13}\text{C}$ value in it is positive (2.0‰). Another sample from the same interval (44-390A-14-1-83-86(A)) consists of very well preserved, hollow tests of *Globotruncana*, whose isotopic composition -0.8‰ for $\delta^{18}\text{O}$, 2.0‰ for $\delta^{13}\text{C}$ are similar to those of the bulk sample (Fig. 4; Table 2).

The core interval 86-90 mm of Hole 390A (lower Maastrichtian *Globotruncana tricarinata* Zone) contains a planktic foraminiferal assemblage consisting of *Globotruncana linneiana* (de Orbigny), *G. arca* (Cushman), *G. falsostuarti* Siga, *Rosita fornicata* Plummer, *Globotruncanita stuartiformis* (Dalbiez), *Planoglobulina multicamerata* de Klasz. $\delta^{18}\text{O}$ value in the foraminiferal bulk sample of the mentioned interval (44-390A-14-2-86-90) is -0.6‰ , which corresponds to a paleotemperature of 14.4°C; its $\delta^{13}\text{C}$ is 2.0‰.

3.11. Hole 516F, South Atlantic

Two bulk-samples of planktic foraminiferal tests from the early Maastrichtian *Globotruncana tricarinata* Zone (72-516F-100-2-34-38 and 72-516F-100-3-35-39) and one bulk-sample from the late Maastrichtian *Abathomphalus mayaroensis* Zone (72-516F-92-4-109-113) were analyzed (Fig. 1).

All foraminiferal tests are represented by original calcite, but significant amounts of secondary ankerite are present inside the tests. The tests were not cleaned before isotopic analysis. Ankerite saturation provoked very low $\delta^{18}\text{O}$ values in the samples: -4.0‰ for the samples from the early Maastrichtian *Globotruncana tricarinata* Zone ($\delta^{13}\text{C}=1.5\text{‰}$) and -4.1‰ for the sample from the late Maastrichtian *Abathomphalus mayaroensis* Zone ($\delta^{13}\text{C}=1.9\text{‰}$) (Table 4). Calculated paleotemperatures seem to be unrealistically high ($T=29.6\text{--}30.2^\circ\text{C}$).

3.12. Ankerpoort quarry, Maastricht area (Netherlands)

At the type area of the Maastrichtian, the Uppermost Cretaceous member is represented by the Meerssen Chalk exposed at the Ankerpoort/Curfs quarry (50°52'08"N, 5°45'60"E) and characterized by ammonoids *Menuites*, *Baculites*, *Hoploscaphites*, and *Sphenodiscus* (Fig. 1). Overlying Danian sediments consist of chalk of the Geulhem Member (Jagt, 1995).

$\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of the well-preserved taxodont bivalve shell (NET-3-3) from the uppermost Maastrichtian (Fig. 5) are -1.9‰ ($T=19.8^\circ\text{C}$) and 0.8‰, respectively. Isotopic data on carbonate rocks from the Maastrichtian-Danian transition are shown in Table 5.

3.13. South India

The upper Ariyalur Group of the Tiruchirapalli area is represented by the Kallankurichchi Limestone (40 m thick), Ottakovil (60 m thick) and Kallamedu Sandstone (100 m thick) Formations toward the upward sequence and is characterized by Maastrichtian fossils (foraminifera, brachiopods *Carneithyris*, *Chatwinothyris*, bivalves *Lopha*, *Griphea*, *Exogyra*, *Inoceamus*, and ammonoid *Eubaculites*) (Archara Tewari et al., 1996). We analysed isotopic composition of well-preserved bivalve shells of *Lopha* sp. (no. In-2a) from the lower Maastrichtian Kallankurichchi Formation of the Tiruchirapalli area, Southern India (11°07'9.1"N, 79°07'51.5"E) (Fig. 1) in R. Nagendra's collection. $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values in 35 samples collected from its different ontogenetic stages fluctuate between -5.8 and -2.2‰ and between -1.4 and 0.8‰, respectively. Another shell of *Lopha* sp. (no. In-2a-2), as well as those of *Gryphaea* sp. (no. In-1-1 and In-3-1) and *Exogyra* sp. (no. In-4-1) from the same horizon show very light $\delta^{18}\text{O}$ values (-5.6 to -2.0‰). Their $\delta^{13}\text{C}$ values range from -2.2 to 2.6‰.

4. DISCUSSION

4.1. Paleotemperatures calculated from latest Campanian-Maastrichtian mollusks of the Western Interior Seaway

New data on late Campanian trigoniid bivalves of the

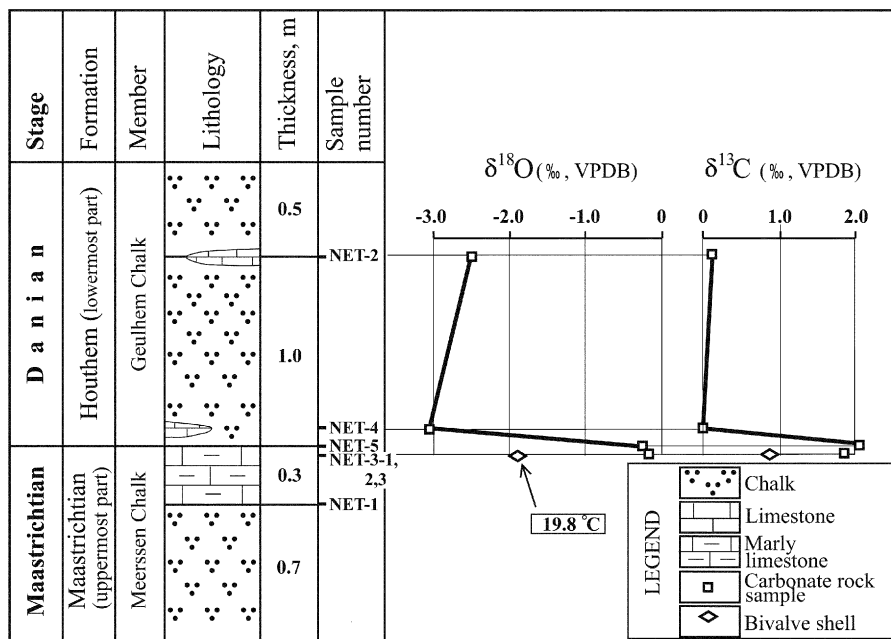


Fig. 5. The $\delta^{18}\text{O}$ and $\delta^{13}\text{C}$ values of latest Maastrichtian and earliest Danian biogenic carbonates from the Maastrichtian and Houthem Formations.

Table 5. Carbon and oxygen isotope analyses of calcitic bivalve shell and carbonate rocks from the Maastrichtian-Danian boundary Transition of the Ankerpoor-Carfs quarry, Maastrichtian type area, Netherlands

Sample	Mollusk group, rock	Formation (member)	Location (H & D, in mm)	Diagenetic alteration			$\delta^{13}\text{C}$ (VPDB), ‰	$\delta^{18}\text{O}$ (VPDB), ‰	T, °C
				Original calcite, %	Admixture, %	Colour			
NET-3-3	Taxodont mollusk	Maastricht (Meerssen chalk, 5 cm below the K/T boundary)	H=15	100	0	White	0.8	-1.9	19.8
NET-4	Limestone	Houthem (Geulhem chalk, 10 cm above the K/T boundary)	-	-	-	Light-gray	-0.1	-3.0	-
NET-2	Limestone	Houthem (Geulhem chalk, 1 m above the K/T boundary)	-	-	-	Light-gray	0.1	-2.5	-

Tennessee State provide evidence on the existence of sufficiently high bottom-water paleotemperatures (16.7–21.2°C) for the WIS shelf (southern extremity). For the determination of sea-surface paleotemperatures in the main Late Cretaceous stages on the basis of data from bottom paleotemperatures of shallow water basins, we offer to use a small correction (about 2.0–2.5°C). As a result, we expect that sea-surface temperatures in the late Campanian WIS shelf fluctuated from 19.2 to 23.7°C.

Available data on Maastrichtian isotopic paleotemperatures, including some original data on WIS (South Dakota), Koryak Upland, Sakhalin, Shatsky Rise, Ontong Java Plateau and South Atlantic are shown in Figs. 6–8. Judging from new data on Dakota ammonoids, the middle Maastrichtian bottom-water paleotemperatures of the southern part of the WIS fluctuated from 17.6 to 22.8°C (Fig. 7); sea-surface paleotemperatures there seem to be about 20.1–25.3°C.

According to Cochran *et al.* (2003), late Maastrichtian paleotemperatures of near bottom waters of the WIS (with

normal salinity) varied between 10.7 to 22.4°C; sea-surface paleotemperatures were about 13.2–24.9°C. As noted by Naidin (2001), the WIS and similar meridional straits in the Northern Hemisphere were the main poleward conveyors during the Late Cretaceous. These types of basins are absent in the hydrodynamic system of the Recent World Ocean.

4.2 Prospective ocean surface temperatures for equatorial latitudes from latest Campanian, middle and late Maastrichtian mollusks data

Isotopic composition of both Late Campanian aragonitic trigonid bivalve shells and middle-late Maastrichtian aragonitic ammonoid shells of the WIS show that its sea-surface temperatures were sufficiently high in the Latest Campanian (to 23.7°C), middle (to 25.3°C) and late (to 26.6°C) Maastrichtian time. This is the case of inhabitation by mollusks in conditions of normal salinity. Because the southern part of the WIS might have had a direct and active thermal connection with equatorial sea-surface waters, we

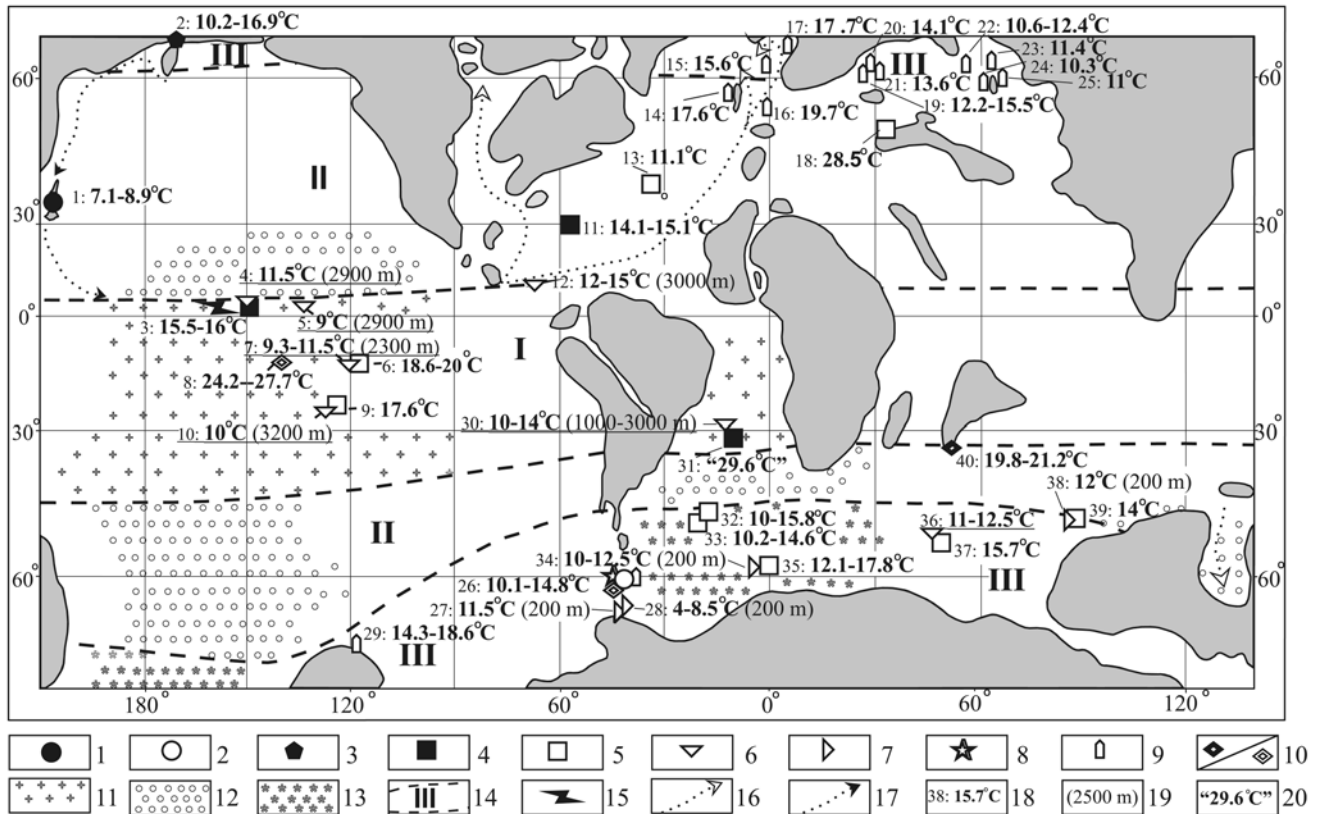


Fig. 6. Map showing isotopic paleotemperatures for the early Early Maastrichtian. 1-2 - from ammonoid shells: 1 - original data; 2 - from literature; 3 - from brachiopod shells (original data); 4-5 - from planktic foraminiferal shells: 4 - original data; 5 - from literature; 6-7 - from benthic foraminifera (from literature); 6 - oceanic depths; 7 - shallow water basin; 8 - from nautiloid shells (from literature); 9 - from belemnite rostra (from literature); 10 - from bivalve shells (a - original data, b - and from literature); 11-13 - tanatocoenosis types of planktic foraminifera (Sokolova, 1998, 1999): 11 Tropical-Subtropical; 12 - Subtropical, 13 - Warm-Temperate, 14 - climatic zone and their boundaries; 15 - upwelling; 16 - proposed warm current; 17 - proposed cold current; 18 - locality number and paleotemperatures; 19 - depth; 20 - paleotemperatures calculated from diagenetically altered material. Localities: 1 - Naiba River; Sakhalin (Zakharov et al., 1999); 2 - Tundrovaya River, Koryak Upland (Zakharov et al., 2002); 3-4 - DSDP Hole 305, Shatsky Rise: 3 - original data from planktic foraminifera; 4 - from benthic foraminifera (Douglas and Savin, 1975); 5 - Hole 465, Central Pacific (Boersma and Shackleton, 1981); 6-7 - Hole 171, Horizon Guyot (Douglas and Savin, 1973); 6 - from planktic foraminifera; 7 - from benthic foraminifera; 8 - Wodejebato guyot (Wilson and Opdyke, 1996); 9 - DSDP Hole 167, Magellan Rise (Douglas and Savin, 1973); 10 - Tennessee, USA (original data); 11 - Hole 390A, North Atlantic (original data); 12 - Hole 151/152, Caribbean Sea (Burma, 1986); 13 - Hole 384, North Atlantic (Burma, 1986); 14 - England (Lowenstam and Epstein, 1954; 1959); 15 - Denmark (Lowenstam and Epstein, 1954; 1959); 16 - Netherlands (Lowenstam and Epstein, 1954; 1959); 17 - Sweden (Lowenstam and Epstein, 1954; 1959); 18 - Hekimhan, north-eastern Turkey (Yildiz and Özdemir, 1999); 19 - Ukraine (Teiss and Naidin, 1973); 20 - Chernigov region, northern Ukraine (Lowenstam and Epstein, 1959); 21 - Sumy region, northern Ukraine (Lowenstam and Epstein, 1959); 22 - South Urals (Teiss and Naidin, 1973); 23 - Trans-Urals (Teiss and Naidin, 1973); 24 - Emba River (Golbert, 1987); 25 - Ayat River, Turgai area (Golbert, 1987); 26 - James Ross and Vega Islands, Antarctic (Pirrie and Marshall, 1990); 27 - Seimur Island, Antarctic (Barrera et al., 1987a); 28 - Seimur Island, Antarctic (Pirrie and Marshall, 1990); 29 - New Zealand (Early? Maastrichtian) (Stevens and Clayton, 1971); 30 - Hole - 357, South Atlantic (Barrera et al., 1987a); 31 - Hole 516F, South Atlantic (ankerite saturation - original data); 32 - Hole 327A, Falkland Plateau (Huber et al., 1995); 33 - Hole 511, Falkland Plateau (Huber et al., 1995); 34-35 Hole 690, South Atlantic (Barrera et al., 1987a); 34 - from benthic foraminifera; 35 - from planktic foraminifera; 36 - Hole 747, South Pacific (Barrera, 1994); 37 - Hole 750, South Pacific (Barrera, 1994); 38-39 Hole 261, South Pacific (Barrera, 1994); 38 - from benthic foraminifera; 39 - from planktic foraminifera; 40 - Ariyalur, Trichinopoly area, Southern India (original data).

suggest that calculated temperatures might be closer to tropical deep-sea surface temperatures. Consequently, latest Campanian and middle Maastrichtian equatorial ocean surface temperatures calculated from mollusk isotopic composition were about 3.7–4.3°C higher than those calculated from equatorial planktic foraminifera of the Pacific (Dou-

glas and Savin, 1971, 1973; D'Hondt and Arthur, 1996; MacLeod et al., 2000; Price and Hart, 2002). According to our results, equatorial sea-surface paleotemperatures could have been in excess of 23.7° and 25.3°C during the Latest Campanian and middle Maastrichtian, respectively; according to Yiliz and Özdemir's (1999) data, sea-surface pale-

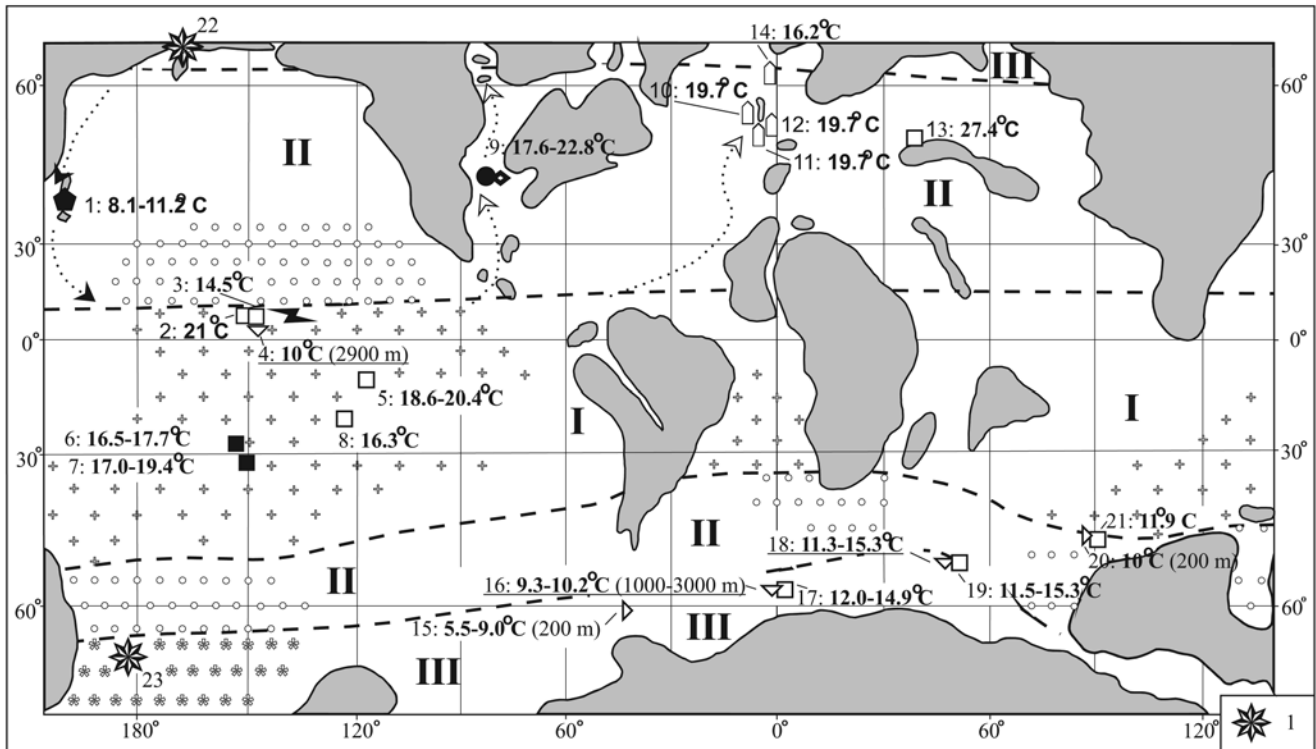


Fig. 7. Map showing paleotemperatures for the middle Maastrichtian 1 – Boreal type radiolarians (Basov and Vishnevskaya, 1991). Other designations as in Fig. 6. Localities: 1 – Naiba River, Sakhalin (Zakharov *et al.*, 1999); 2 – Hole 47, Shatsky Rise (Douglas and Savin, 1971); 3–4 – Hole 305, Shatsky Rise (Douglas and Savin, 1975); 3 – from planktic foraminifera, 4 – from benthic foraminifera; 5 – Hole 171, Horizon Guyot (Douglas and Savin, 1973); 6 – Hole 289, Ontong Jawa; 7 – Hole 288A, Ontong Jawa Plateau; 8 – Hole 167, Magellan Rise (Douglas and Savin, 1973); 9 – South Dakota (original data); 10 – England and Netherlands information on middle Maastrichtian paleotemperatures of West Europe here and in other places has been done by Lowenstam and Epstein (1954, 1959) taking into account some data shown in Fig. 58 (Teiss and Naidin, 1973); 11 – France (Lowenstam and Epstein, 1954; 1959); 12 – Belgium and Netherlands (Lowenstam and Epstein, 1954; 1959); 13 – Hekimhan, north eastern- Turkey (Yildiz and Özdemir, 1999); 14 – Denmark, Sweden and Poland (Lowenstam and Epstein, 1954; 1959); 15 – Seimur Island, Antarctic (Barrera *et al.*, 1987a); 16–17 Hole 690, South Atlantic (Barrera, 1994); 16 – from benthic foraminifera, 17 – from planktic foraminifera; 18–19 – Hole 750, South Pacific (Barrera, 1994); 18 – from benthic foraminifera, 19 – from planktic foraminifera, 20–21 – Hole 761, South Pacific (Barrera, 1994); 20 – from benthic foraminifera, 21 – from planktic foraminifera; 22 – Koryak Upland (Basov and Vishnevskaya, 1991); 23 – Lord How Rise (Basov and Vishnevskaya, 1991).

otemperature in the northern Subtropical Zone during the middle Maastrichtian was about 27.4°C. (Fig. 6).

Equatorial deep-sea surface paleotemperature estimates from Maastrichtian mollusks were first made by Wilson and Opdyke (1996), who obtained high paleotemperatures from early Maastrichtian aragonite-bearing rudist bivalve shells from Hole 877A. They inferred that bottom-water tropical paleotemperatures for an atoll shelf area fluctuated from 24.2° to 27.7°C (expected sea-surface paleotemperatures are 26.9 to 30.2°C) (Fig. 6). However, we must take into account the vital effect of the rudist shell oxygen isotope values of the predisposition of some rudist mollusks to photosymbiotic adaptations (Skelton and Wright, 1987; Kauffman and Johnson, 1988; Lewy, 1995; Steuberg, 1996; Moro *et al.*, 2002). Somewhat lower temperatures (to 28.5°C) were calculated from Early Maastrichtian planktic foraminifera of the Subtropical Zone (Yiliz and Özdemir, 1999).

Paleotemperatures obtained from Early Maastrichtian cephalopods of the WIS (Cochran *et al.*, 2003) are consistent with those from late Maastrichtian planktic foraminifera in the low latitudes (26.6°C) (Fig. 8) (D'Hondt and Arthur, 1996). The average tropical deep-sea surface paleotemperature values in the middle and late Maastrichtian might be about 0.8–3.7°C lower than those in present-day tropics (Stepanov, 1974), between 5° and 10°N.

4.3. Paleotemperatures calculated from Cretaceous mollusks of India

Recently, we have calculated paleotemperatures from well-preserved belemnite *Actinocamax (Praeactinocamax)* sp. rostra recovered from the Turonian Karai Shale Formation of the Trichopoly area, Southern India (observation from 65 samples) (Zakharov *et al.*, 2006). Those for presumed sum-

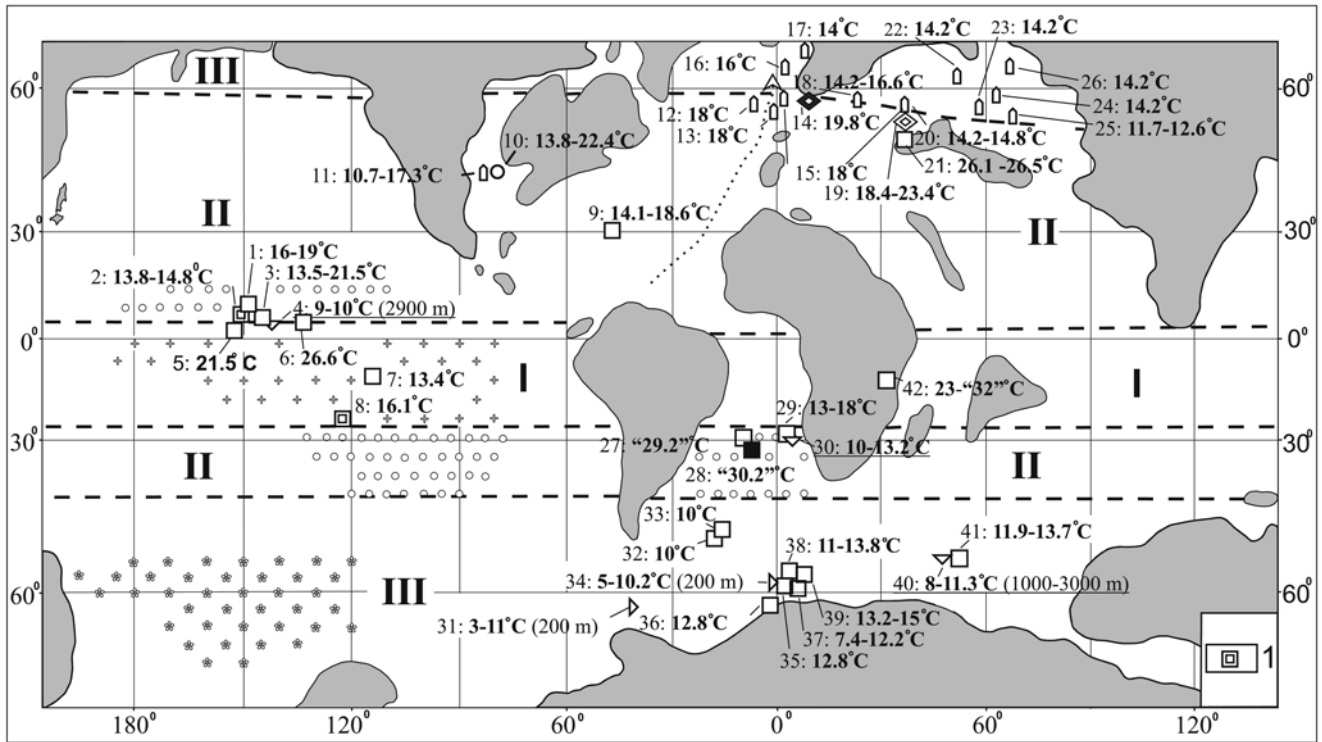


Fig. 8. Map showing paleotemperatures for the Late Maastrichtian. 1 - nanoplanktons Other designations as in Fig. 6. Localities: 1 - Hole 47, Shatsky Rise (Boersma and Shackleton, 1981); 2 - Hole 472, Shatsky Rise (Douglas and Savin, 1971, 1975); 3-4 - Site 305, Shatsky Rise (Douglas and Savin, 1975); 3 - from planktic foraminifera, 4 - from benthic foraminifera; 5 - Hole 577A, Shatsky Rise (D'Hondt and Arthur, 1996); 6 - Hole 465A, Shatsky Rise (D'Hondt and Arthur, 1996); 7 - Hole 171, Horizon Guyot (Douglas and Savin, 1973); 8 - Hole 167, Magellan Rise (Douglas and Savin, 1973); 9 - from belemnite rostra; 10 - from ammonoids shells; 11 - Hole 390A, North Atlantic (Houston et al., 1999); 12 - England (Lowenstam and Epstein, 1954; 1959); 13 - France (Lowenstam and Epstein, 1954; 1959); 14-15 - Netherlands: 14 - from a bivalve shell (original data), 15 - from belemnite rostra (Lowenstam and Epstein, 1954; 1959); 16 - Denmark (Lowenstam and Epstein, 1954; 1959), 17 - Sweden (Lowenstam and Epstein, 1954; 1959); 18 - Lvov area, western Ukraine (Teiss and Naidin, 1973); 19-20 - Crimea (Teiss and Naidin, 1973); 19 - from bivalve shells, 20 - from belemnite rostra; 21 - Hakimhan, North-eastern Turkey (Yildiz and Özdemir, 1999); 22 - Sura River, Russian Platform (Teiss and Naidin, 1973); 23 - Mangyshlak (Teiss and Naidin, 1973); 24 - watershed of the Ural and Emba River (Teiss and Naidin, 1973); 25 - lower reaches of the Amudarya River and northern Aral area (Teiss and Naidin, 1973); 26 - Ayat River, Trans-Urals (Teiss and Naidin, 1973); 27 - Hole 357, South Atlantic (D'Hondt and Arthur, 1996); 28 - Hole 516F, South Atlantic (ankerite including - original data), 29-30 - Hole 515A, South Atlantic (Abramovich and Keller, 2003): 29 - from planktic foraminifera, 30 - from benthic foraminifera, 31 - Seimur Island, Antarctic (Barrera et al., 1987a); 32 - Hole 511, Falkland Plateau (Huber et al., 1995); 33 - Site 327, Falkland Plateau (Huber et al., 1995); 34-35 - Hole 690, Antarctic (Barrera, 1994): 34 - from benthic foraminifera, 35 - from planktic foraminifera; 36 - Hole 692, Antarctic (Stott and Kennett, 1990); 37 - Site 690, Antarctic (Huber et al., 1995); 38 - Hole 689B, Antarctic (Stott and Kennett, 1990); 39 - Hole 689C, Antarctic (Stott and Kennett, 1990); 40-41 - Hole 750, Antarctic (Barrera, 1994): 40 - from benthic foraminifera, 41 - from planktic foraminifera; 42 - Tanzania (Pearson et al., 2001).

mer season for the early Turonian fluctuated from 16.3 to 18.5°; probable isotopic paleotemperatures fluctuate from 14.3 to 16.1°C (near-bottom condition for the epipelagic zone). Paleotemperatures calculated from belemnites which lived in the early Turonian mesopelagic zone varied between 14.3 to 15.9°C. Our revision of Bowen's (1961) isotopic data indicates that the Late Albian paleotemperatures for the Trichopoly are appear to be similar to those of the Turonian time.

In contrast to the isotopic compositions of late Albian and early Turonian belemnites of the Trichopoly area, early

Maastrichtian well-preserved bivalves, as shown above are characterized mainly by very light $\delta^{18}\text{O}$ values and normal $\delta^{13}\text{C}$ values. This might be the result of refreshing of the marine environment, reflecting local fresh-water input. In view of maximal $\delta^{18}\text{O}$ values (fluctuating from -2.2 to -1.9‰ in some portion of investigated bivalve shells), the paleotemperatures in the Southern India shallow-water basin was about 19.8-21.2° in the Early Maastrichtian, during which time India shifted from Warm-Temperature climate zone into the nearby Subtropical zone with the warm and perhaps humid climate.

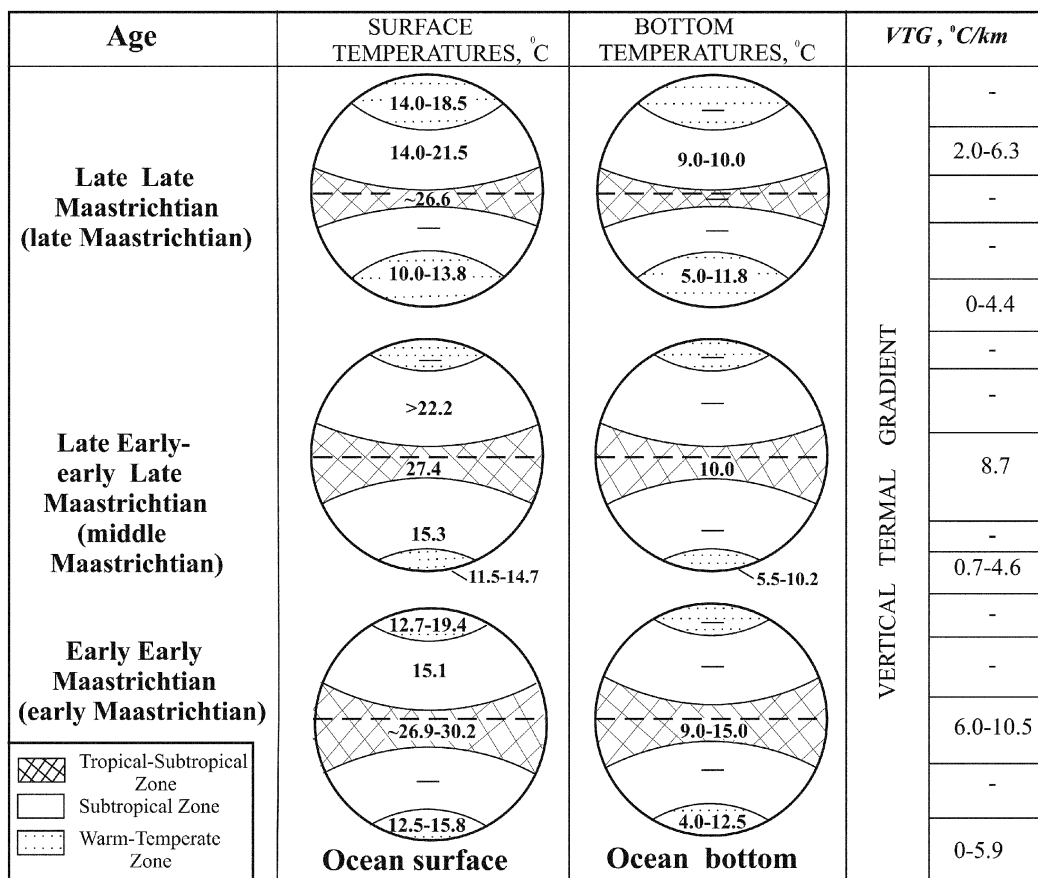


Fig. 9. Reconstruction of ocean surface and deep ocean temperatures for the early, middle and late Maastrichtian.

4.4 Early Maastrichtian planktic foraminiferal thanatocenos and interpretation of ocean surface temperatures in tropics from foraminifera

Three types of planktic foraminiferal thanatocenos were proposed for the Maastrichtian: a Tropical-Subtropical, a Subtropical and a Warm-Temperate (Fig. 9-11) (Sokolova, 1998). According to Sokolova (1998; 1999), their location reflects to a great extent, the Maastrichtian climatic zonation.

Our analysis of early-middle exceptionally well-preserved Maastrichtian, hollow planktic foraminiferal tests from low latitudes of the Pacific (Holes 289, 305 and 288Å) and Atlantic (Holes 390A), confirm that Maastrichtian planktic foraminiferal tests have unusually high $\delta^{18}\text{O}$ values. We have shown that the $\delta^{18}\text{O}$ values in Early Maastrichtian equatorial foraminifera of the Pacific and subtropic foraminifera of Atlantic (at a paleolatitude of 30° N) fluctuate from -1.3 to -0.9‰ (corresponding to paleotemperatures of 15.5–17.1°C) and from -0.8 to -0.5‰ (14.1–15.1°C), respectively. A very similar pattern is also recognized for the middle Maastrichtian subtropical zone: $\delta^{18}\text{O}$ values in investigated planktic foraminifera of the Pacific, at a paleolatitude of 35°S, fluctuate from -1.8 to -1.3‰ (corresponding to paleotemperatures of 17.7–19.4°C). Data obtained

are in good agreement with some published results (Douglas and Savin, 1971, 1973; D'Hondt and Arthur, 1996; MacLeod *et al.*, 2000; Price and Hart, 2002) but are significantly lighter than those from foraminifera of Albian and Turonian ages (Wilson and Norris, 2001; Wilson *et al.*, 2002). Maximum paleotemperatures obtained from early and middle Maastrichtian equatorial foraminifera of the equatorial Pacific do not exceed those by previous authors (20°C; Douglas and Savin, 1973 and 21°C; Douglas and Savin, 1971, respectively). For the tropics of the Late Maastrichtian open ocean, Douglas and Savin (1973) and Boersma and Sackleton (1981) calculated relatively low paleotemperatures ($>19^\circ\text{C}$) and only recently, as shown above, higher paleotemperatures have been reported for some Maastrichtian tropical and subtropical foraminiferal species inhabitation (21.5–28.5°C) (D'Hondt and Arthur, 1996; Yiliz and Özdemir, 1999). Isotopic and compositions of late Maastrichtian foraminifera from isolated epicontinental seas are now available (Pearson *et al.*, 2001).

Price and Hart (2002) noted that although Barron *et al.* (1995) have modelled relatively cool equatorial sea surface temperatures as a result of an increased oceanic poleward heat flux, the isotopically-derived temperatures, followed by the classic Cretaceous record from the foraminifera, are

nevertheless possibly too cold. Investigators have given different explanations for unusually heavy $\delta^{18}\text{O}$ values in some Cretaceous and Paleogene equatorial planktic foraminiferal shells (mainly from the Pacific):

(1) Diagenesis. As pointed out by Schrag (1999) and Pearson et al. (2001) and others (Clarke and Jenkyns, 1999), a strong diagenetic recrystallization in cold bottom waters increased the $\delta^{18}\text{O}$ values in oceanic sediments. This sort of diagenesis is the reverse of the more known trend towards negative values in meteoric and high-temperature diagenesis. (2) Thermocline dwelling. Wilson and Opdyke (1996), allowed the possible influence of recrystallization of planktic foraminiferal shells in cool waters of shallow burial depths on changing their isotopic composition. They assume that similar effects might be provoked by $\delta^{18}\text{O}$ analyses of foraminifera that were either thermocline or subthermocline dwelling. D'Hondt and Arthur (1996) considered that a general tendency toward weak ^{18}O depletion may reflect kinetic discrimination against ^{18}O and ^{13}C during rapid calcification (McConnaughey, 1989), and/or ontogenetic migration of planktic foraminifera downward through water mass of decreasing calculated temperature. Most micropaleontologists regarded that well selected materials represented by well-preserved foraminiferal tests without having secondary calcite or other minerals in their cavities and pores are entirely suitable for successful investigation on isotopic thermometry (MacLeod et al., 2002; D'Hondt and Arthur, 1996; Price and Hart, 2000; Houston et al., 1999; Huber et al., 2002), but it is very important to realize at what depth their formation took place (information on foraminiferal shell inflation and a strong positive shift in $\delta^{13}\text{C}$ usually used for the identification of surface conditions) (Houston et al., 1999). (3) Hypersalinity. Price and Hart (2002) expect that high $\delta^{18}\text{O}$ values in Cretaceous equatorial planktic foraminifera might be partly caused by increased water salinity. (4) Seawater pH. Zeebe (2001) considered that mid-Cretaceous might be $\sim 2\text{--}3.5^\circ\text{C}$ higher than previous estimates because the pH effect on foraminiferal $\delta^{18}\text{O}$ has not been taken into account. (5) Upwelling. According to Zachos et al. (1994) and Price and Hart (2002), the oxygen isotopic data giving relatively low tropical sea surface temperatures are robust. For data coming from sites located in past upwelling zones, a relatively low thermal gradient within some ages of Cretaceous may reflect a well-mixed ocean, resulting from mixing through upwelling of waters. (6) Unusual hydrography and active vertical migrations. According to Kobashi et al. (2001, 2004), in a less stratified ocean, suggested by early and middle Eocene temperatures and density profiles, planktic foraminiferal habitat depth might be wider, thus biasing isotopic sea surface temperature estimates for planktic foraminifera to cooler values.

Each of these factors seems to explain the strange isotopic composition of some Maastrichtian equatorial planktic foraminiferal tests. However, we infer that for early-middle

Maastrichtian time, the most important factors are (1) local influence of past equatorial upwelling zones and cool currents and also (2) ontogenetic vertical migration of planktic foraminifera. Recognition of diagenetic alterations in calcitic tests is difficult, the role of diagenetic alteration in the explanation of cool tropic paradox seem to be overestimated, in view of significant oxygen and carbon isotope differentials among the different foraminifera species analyzed (Houston and Huber, 1998; Houston et al., 1999; MacLeod and Huber, 2001; Price and Hart, 2002; Huber et al., 2002). In the course of strong diagenesis, these distinctions might have been graded. Besides, diagenetically-altered foraminiferal tests are often excluded from the process of thermometry.

It must be admitted that knowledge on diagenetic process, which took place at the oceanic bottom, is not yet fully understood. Zachos et al. (1994) have shown their data on recrystallization of foraminiferal tests, leading to increase in their $\delta^{18}\text{O}$ values, which is not typical for shallow marine sediments. But we came across another type of the secondary change, studying Early Maastrichtian planktic foraminifera of the South Atlantic (Hole 516F): the relatively high "paleotemperatures" (30.2°C), obtained from these uncleaned shells are explained by the infilling of their cavities with secondary ankerite. These data challenge the information concerning high paleotemperatures calculated from foraminifera of the nearest Hole 357 (29.2°C) (D'Hondt, Arthur, 1996) as it may be explained by the same reason. Overestimated paleotemperature results from the Hole 390A, unconfirmed by our investigation, seem to be also need to be more accurately defined.

4.4. $\delta^{13}\text{C}$ Negative Excursions

Two negative $\delta^{13}\text{C}$ excursions are known for the Maastrichtian. The first $\delta^{13}\text{C}$ negative excursion fixed in both the benthic and the planktic foraminifera took place just before the inoceramid and rudist bivalve extinction at the end of the early Maastrichtian (the heaviest $\delta^{13}\text{C}$ values occur within the Lower-Upper Maastrichtian boundary sequence) (Barrera 1994; McLeod and Huber, 1996, 2001; McLeod et al., 2000). The negative values were discovered in the brachiopod and ammonoids shells from the Lower Maastrichtian of Sakhalin; brachiopod shells from the Upper Maastrichtian in the same region are characterized by the positive $\delta^{13}\text{C}$ values (to 1.8 ‰). In some bivalve shells from the Lower-Upper Maastrichtian boundary beds of South Dakota, the heaviest $\delta^{13}\text{C}$ (3.8‰) were discovered. In the Pacific (288A and 289 Holes), planktic foraminiferal tests from the same level are also characterized by relatively heavy $\delta^{13}\text{C}$ values (to 2.8‰), which distinguish them from the Early Maastrichtian planktic foraminifera of the Hole 390A in North Atlantic (2.0‰).

The second $\delta^{13}\text{C}$ negative excursion was discovered by us

just in the Maastrichtian-Danian boundary level in the stratotype area of the Maastrichtian (Netherlands, Ankerpoort). $\delta^{13}\text{C}$ value in the taxodont bivalve shell collected at 5 cm below the K/T boundary is 0.8‰, limestone at 10 cm above the mentioned boundary shows significantly lighter value (–0.1‰). The light $\delta^{13}\text{C}$ value in the foraminiferal shells from the same level has been discovered in many other regions (Boersma and Shackleton, 1981; Stott and Kennett, 1990).

This suggest that the $\delta^{13}\text{C}$ negative excursion in the end of the Early Maastrichtian preceeded the extinction of some large mollusk groups (inoceramid and rudist bivalves), but those in the Cretaceous-Tertiary boundary time coincided with the mass extinction of marine and non-marine organisms. The mentioned negative carbon-isotopic anomalies may be provoked by different reasons: cooling for the first case and cooling and possible global anoxia attendant to volcanic activity for the second one, taking into account the scale of the effect in the biosphere.

5. CONCLUSIONS

1. New data on oxygen isotopic composition of Cretaceous mollusks does not confirm D'Hondt and Arthur's (1996) idea on Maastrichtian cool tropic paradox relying on planktic foraminifera. Judging from isotopic composition of mollusk shells from the Maastrichtian of the Western Interior Seaway, characterized by the very active oceanic poleward heat transport, and Maastrichtian rudist bivalve shells from Hole 877A (Wilson and Opdyke, 1996) and foraminiferal tests of SE Turkey (Yidliz and Özdemir, 1999), deep-sea surface temperatures in tropic during early, middle, and late Maastrichtian could have been from 26.9 to 30.2°C, 27.4 and 26.6°C, respectively. Apparently, they did not reach the level (32±3°C) denoted for the Late Albain and Turonian (Norris and Wilson, 1998; Wilson and Norris, 2001; Wilson *et al.*, 2002).

2. Both the Maastrichtian (Zakharov *et al.*, 2003a) and the Eocene (Kobashi *et al.*, 2004) seem to be the time when tropical planktic foraminifera lived in a much wider depth range in comparison with planktic foraminifera of some other Mesozoic and Cenozoic stages, biasing the total foraminiferal $\delta^{18}\text{O}$ to heavier values (but calculated temperatures to lower values) in conditions of a less stratified ocean.

3. These data confirm that there were late Early Maastrichtian and latest Maastrichtian-earliest Danian $\delta^{13}\text{C}$ negative excursions at the very end of the Mesozoic reflecting two significant evolutionary events: extinction of inoceramid and rudist mollusks and mass extinction, including extinction of ammonoids and belemnites.

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