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Climatic Variations in the Arctic Region in the Cretaceous and Cenozoic

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Abstract—As a result of an analysis of published works, a database on paleotemperatures in the Arctic and Subarctic regions based on marine invertebrate skeletons, marine palynomorphs, dinosaur teeth, analysis of the ability of reptiles to lay eggs at low temperatures, terrestrial floral assemblages (CLAMP-analysis), the presence of coal interbeds in continental deposits within the Arctic region, and on membrane lipids of glycerol and dialkylglycerol tetraether in marine sediments and glendonite was assembled. Based on the obtained data, a paleotemperature curve for the Cretaceous–Cenozoic stage of geological history of the Arctic region was constructed. The general trends of this curve are in consistent with the global paleotemperature curve (Scotese, 2015) (with the exception of the cooling period in the Tortonian due to local factors). In total, 16 climatic cycles in the climatic history of the Arctic, including 16 climatic minima (including the glaciation in the Northern Hemisphere) and 15 climatic maxima have been established.

Keywords: Cretaceous, Paleogene, Neogene, Quaternary, climate, paleotemperatures, Arctic

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INTRODUCTION

The Arctic Region, including the sea economic zone and the adjacent continental shelf zone occupies approximately 30% of the territory of the Russian Federation (Nekipelov and Makosko, 2011). The exploration and prospecting of mineral deposits in the Arctic is one of crucial tasks in terms of the study of the development of mineral resources. In order to solve this problem, it is necessary to make a detailed analysis of the geological history of the Arctic and, in particular, to study the dynamics of climatic variations.

The object of the present work is a bibliographical review of the data on paleotemperatures estimated for Cretaceous–Neogene deposits in the Arctic region. A database on paleotemperatures in the Arctic and Subarctic regions was collected.

MATERIALS AND RESEARCH METHODS

The Arctic Region (Arctic, or Trans-Arctic), which is located to the north of the Arctic Circle (66° N) was selected as a subject of investigation. In addition, during the data collection the paleotemperature data from geological sections were obtained and analyzed. These sections are located to the south of the Arctic Circle in a range of 54–66° N within the Sub-Arctic belt or region (50–70° N, in dependence on the local landscape–climatic conditions). In total, more than 500 works on the geological structure of the studied area were analyzed. This work continues and develops the investigation of climatic variations in the Arctic in the Cretaceous (Badulina and Gabdullin, 2018). A comparative analysis of climatic variations in the Arctic with a global temperature trend was performed (Scotese, 2015).

The paleotemperature values were determined on the basis of studying skeletons of marine invertebrates (Zakharov et al., 2011), marine palynomorphs (Shreck et al., 2011), dinosaur teeth (Suarez et al., 2013), terrestrial floral assemblages (including CLAMP-analysis) (Akhmetiev, 2004; Budantsev, 1983; Budantsev, 1999; Budantsev and Golovneva, 2009; Golovneva, 1994; Golovneva, 2000; Greenwood and Wing, 1995; Herman, 2016; Herman and Spicer, 1996; Spicer et al., 2016; Wolfe and Dilcher, 2001), membrane lipids of glycerol and dialkylglycerol tetraether in marine sediments (Crampton-Flood, 2018) and glendonite (Herrle et al., 2015; Rogov et al., 2017). The paleotemperature values obtained with the CLAMP analysis were also analyzed from the standpoint of the ability of dinosaurs to lay eggs at low temperature, for example, when it was insufficiently warm or the summer season was insufficiently long (Herman, 2016). In addition, the paleotemperature vales were estimated on the basis of the presence of coal interbeds in continental deposits of the Arctic region (Galloway et al., 2013). The procedures for determining temperature values based on different organic remains, minerals, and chemical compounds have been described in detail in the above-listed works.

RESULTS AND DISCUSSION

Despite the variety of methods and approaches to determining paleotemperatures, in general, they show good correlation (Fig. 1). In most cases, the average annual temperature was determined, which makes it possible to use these data (with the exception of isotope paleothermometry data on invertebrates obtained for the Cretaceous and Eocene) together with the surface water temperature estimated on the basis of marine palynomorphs for the Neogene. When analyzing the paleotemperature values, we assume that there is no difference between the temperature values obtained for the water column based on the isotope data (belemnite rostra and ammonite shells, etc.) and the temperature for the near-surface part of the atmosphere (CLAMP-analysis) are equal, although they are not similar. In particular, ocean water cannot be colder than −2.5°C, while the average annual air temperature may be −10°C.

The flora-based paleotemperature values correlate well with the isotope paleothermometry data estimated on the basis of invertebrate shells (for example, for the Cretaceous and Eocene). The data obtained from plant macroremains are comparable with those from the terrestrial fauna (for example, for the Cretaceous) or palynological data (for example, for the Paleogene). The temperature values obtained with the palynological analysis correlate well with those for membrane lipids of glycerol and dialkylglycerol tetraether in marine sediments.

When correlating the paleotemperature values, one should take the paleolatitude into account, since the temperature rises when moving towards the equator. Therefore, in addition to the geographical reference, the degrees of north latitude (in the modern system of coordinates) are indicated for temperature values at a point. It would be methodologically more correct to use paleolatitude values. However, since it is difficult to accurately determine the location of modern sections in the geological past in the modern system of coordinates, modern coordinates should be used.

When shifting by approximately 10° in the latitudinal direction from the pole to the equator, the temperature increased in the Cretaceous, on average, by 1°C at high latitudes, by 1.5°C in the middle latitudes, and by 2°C and more in low latitudes (Herman and Spicer, 2012). These regularities are also characteristic of the Cenozoic. The paleotemperature curve constructed for Western Siberia (Volkova, 2011) (55° N, on average) is characterized (Fig. 1), on the one hand, by relatively higher temperature values, relative to points located within 70–80° N, and lower temperature values relative to the Global Temperature Curve, on the other (Scotese, 2015). The last curve serves as a test curve from the point of view that the temperature values at high latitudes cannot be equal to or be higher than the planetary temperature values. The general trends must coincide and variations in trends can be explained by local factors that determine the climate (for example, the opening or the closure of straits, along which the water exchange takes place: warm waters penetrated into the high latitudes or cold waters reached the low latitudes).

It is likely that the sharp decrease in temperature that was recorded on the global paleotemperature curve (by 11°C (Scotese, 2015)) and the curve, constructed for the Arctic and Subarctic (by 7.5°C (Herman et al., 2016)) shows the correlation with the Cretaceous–Paleogene boundary and corresponds to an impact event (the impact of an asteroid in the Gulf of Mexico). A number of climatic maxima (phases of climate warming) can be distinguished in the geological history of the Arctic: at the end of the Berriasian (Zakharov et al., 2011), in the Hauterivian, at the end of Barremian (Zakharov et al., 2011), in the mid-Aptian (Zakharov et al., 2011), at the end of Albian (Schroeder-Adams, 2014), in the Turonian and Campanian (separate phases at the beginning and at the end) (Zakharov et al., 2011), in the beginning of the Ypresian time, in the beginning of the Lutetian time, in the mid-Chattian time, at the end of the Burdigalian time, in the mid-Tortonian, and at the end of the Messinian, in Zanclian time. The latest climatic optimum was followed by the onset of glaciation in the Northern Hemisphere (Crampton-Flood, 2018).

The available data show a good correlation of the paleotemperature data within the paleoclimatic zones of the Arctic Region from the North Atlantic (the North Sea in the North German Depression, the Hank well), Greenland (the Icelandic (well ODP 907A) and Norwegian seas) and Spitsbergen (the Barents Sea) via the Polar Urals, Yamal Peninsula, Tazovsky Peninsula, northern Siberia (the Kara Sea basin), the Far East (Laptev Sea–Lena River, Nordvik Bay, Anabar River; East Siberian Sea–New Siberia Island, Ayon Island), Alaska (Beaufort Sea area–Brooks Range, Kenning River, Martins Crick and the Bering Sea, the Kamchatka Peninsula, Koryak Highland, Pacific Ocean, and the Gulf of Alaska–Kupriyanov Island), Canadian Arctic Archipelago (Baffin Sea: Sverdrup, Amund Ringes, Axel Heiberg, and Ellesmere Islands).

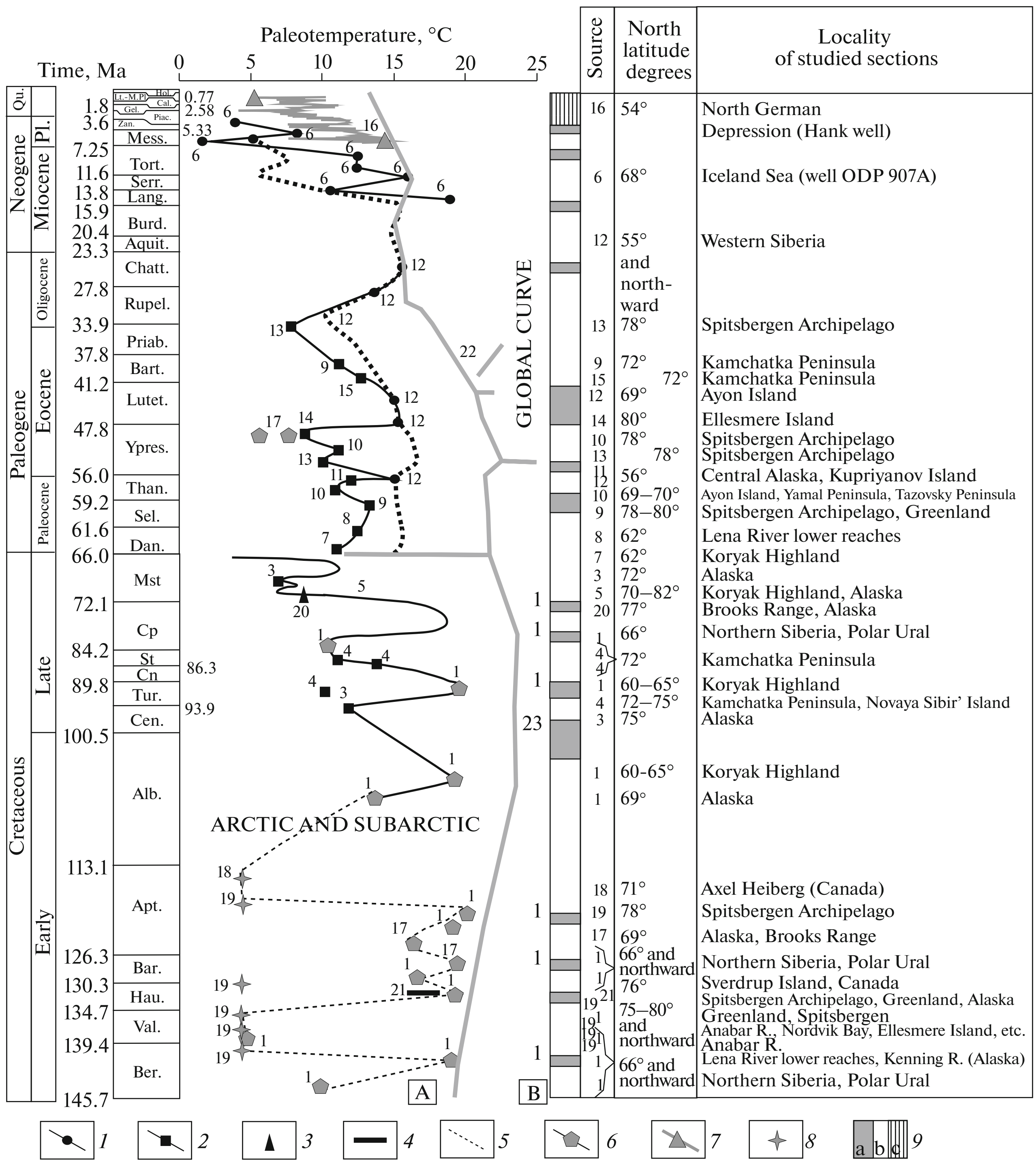

Fig. 1. Paleotemperature curves for the Cretaceous–Cenozoic: A, Arctic and Subarctic regions; B, global after (Scotese, 2015): (*1*) marine palynomorphs; (*2*) continental flora (CLAMP-analysis); (*3*) dinosaur teeth; (*4*) coal; (*5*) isotopy based on skeletons of marine invertebrates; (*6*) based on palynology, (*7*) based on membrane lipids of glycerol and dialkylglycerol tetraether in marine sediments; (*8*) glendonite; (*9*) type of climate: (*9*a) climatic maximum (warming period), (*9*b) climatic minimum (cooling period), (*9*c) the onset of glaciation in the Northern Hemisphere. Source, a number of a source of literature: (1) Zakharov et al., 2011; (2) Eberle et al., 2010; (3) Spicer et al., 2016; (4) Herman and Spicer, 1996; (5) Herman et al., 2016; (6) Shreck et al., 2011; (7) Golovneva, 1994; (8) Budantsev, 1983; (9) Akhmetiev, 2004; (10) Golovneva, 2000; (11) Wolfe and Dilcher, 2001; (12) Volkova, 2011; (13) Budantsev and Golovneva, 2009; (14) Greenwood and Wing, 1995; (15) Budantsev, 1999; (16) Crampton-Flood et al., 2018; (17) archive data after D.P. Naidin (MSU); (18) Herrle et al., 2015; (19) Rogov et al., 2017; (20) Suarez et al., 2013; (21) Galloway et al., 2013; (22) Scotese, 2015; (23) Schroeder-Adams, 2014.

THE HISTORY OF CLIMATIC VARIATIONS IN THE ARCTIC AND SUBARCTIC

The cyclicity of climatic fluctuations is presented by the following maximums and minimums.

(1) The *Early Berriasian climatic minimum* (Zakharov et al., 2011) in sections of Northern Siberia and Polar Urals was established on the basis of isotope paleothermometry data (+11.8–14.9°C).

(2) The *Late Berriasian climatic maximum* is characterized by isotope paleothermometry data (Zakharov et al., 2011) obtained for deposits in Northern Siberia and the Polar Urals (+18.0–23.6°C).

(3) the *Late Berriasian–Valanginian climatic minimum* (Zakharov et al., 2011) is based on the isotope paleothermometry data obtained for the Early Valanginian (+5.3–10.4°C, Spitsbergen, Greenland), Late Valanginian (+15.6–17.8°C) and the Hauterivian–Valanginian boundary (+2.0–14.0°C, Polar Urals, Northern Siberia), as well as on finds of glendonite (+7°C) in the Upper Berriasian sections in the Lena River Kenning River (Alaska), as well as in the Valanginian sections in the Anabar River basin, in Nordvik Bay, on Amund Ringnes Island (Canadian Arctic Archipelago), Lower Valanginian, on the Axel Heiberg and Ellesmere islands; and the Upper Valanginian in Western Siberia (Rogov et al., 2017).

(4) The *Early Hauterivian climatic maximum* was recognized on the basis of isotope paleothermometry data (Zakharov et al., 2011) obtained for Lower Hauterivian deposits in the Northern Siberia and Polar Urals (14.8–21.2°C) and in the Russian Far East (21.0°C), as well as on the basis of coals in the high latitudes: on Sverdrup Island in the Canadian Arctic Archipelago (Galloway, 2013).

(5) The *Late Hauterivian–Early Barremian climatic minimum* was recognized on the basis of isotope paleothermometry data (Zakharov et al., 2011) obtained for Lower Barremian deposits in the Russian Far East (+18.4–24.5°C), as well as finds of glendonite (the formation temperature is usually approximately +7°C) in the Upper Hauterivian sections on the Spitsbergen Archipelago, Kilen River (Greenland), and Kenning River (Alaska) (Rogov et al., 2017).

(6) The *Late Barremian climatic maximum* was recognized on the basis of isotope paleothermometry data obtained for the Upper Barremian deposits in the Russian Far East (+17–18°C, archive data).

(7) The *Early Aptian climatic minimum* was recognized on the basis of the isotope paleothermometry data obtained for the Lower Aptian deposits in Alaska (+17–18°C, archive data).

(8) The *Middle Aptian climatic maximum* (Zakharov et al., 2011) is characterized by isotope paleothermometry data obtained for the deposits of the Russian Far East (+16.4–24.5°C) (Zakharov et al., 2011).

(9) The *Late Aptian–Early Albian climatic minimum* is based on glendonite finds in the section in the Spitsbergen Archipelago (Rogov et al., 2017) and on Axel Heiberg Island (Canadian Arctic Archipelago) (Herrle et al., 2015).

(10) The *Late Albian–Early Cenomanian climatic maximum* (Schroeder-Adams, 2014) was recognized on the basis of isotope paleothermometry data obtained for the Russian Far East (12.5–18.6°C) and in Alaska (12.9–19.0°C) (Zakharov et al., 2011), as well as the relative thermophilic character of the Albian flora on Kotel'ny Island (Herman and Spicer, 2010).

(11) The *Late Cenomanian–Early Turonian climatic minimum*. The following flora-based paleotemperature values were obtained for the Late Cenomanian: 12.5°C (Alaska) and 12.9°C (the Russian Far East). The average winter temperature for Alaska was estimated at +5.7°C, while the average summer temperature was +20.0°C; the average winter temperature for the Grebenka River section (northeastern Russia, the Chukotka Autonomous Region) was 5.9 ± 3.8°C, while the average summer temperature was 20.8 ± 2.8°C (Spicer et al., 2016).

(12) The *Turonian climatic maximum* (Zakharov et al., 2011). The paleotemperature values obtained for the Turonian time based on the CLAMP-analysis of floral assemblages are in a temperature range of 6.9–9.0°C (Herman, 2012), while those obtained for the late Turonian were 15.5–23.3°C (Spicer and Herman, 2013), which is confirmed by the isotope paleothermometry data (+14.1–16.3°C, Russian Far East) (Zakharov et al., 2011).

(13) The *Coniacian–Santonian climatic minimum*. The Coniacian time (the Russian Far East) is characterized by paleotemperature values in the range of 14.1–16.3°C (Zakharov et al., 2011), while there are CLAMP-based data in a range of 9.0–12.5°C (Herman, 2012). Based on CLAMP-analysis of flora, the paleotemperature at the Coniacian–Santonian boundary was 17.0°C (Spicer and Herman, 2013) (Kamchatka Peninsula). Based on the CLAMP-analysis of the floral assemblages, the Santonian time is characterized by paleotemperature values that vary from 9.1 to 15.6°C (Spicer and Herman, 2013) (Northern Siberia), while on the basis of the isotope paleothermometry data they vary from 10.9 to 22.4°C (Russian Far East) (Zakharov et al., 2011).

(14) The *Campanian climatic maximum* (Zakharov et al., 2011). The time period from the Early Campanian to the first half of the Maastrichtian inclusive is characterized by paleotemperature values that vary from 19.4 to 25.5°C (Zakharov et al., 2011) (Alaska) and from 20.6 to 25.5°C or a temperature interval of 12–18°C for the Campanian time (Herman and Spicer, 2010) (Russian Far East).

(15) The *Early Maastrichtian climatic minimum* was recognized based on the CLAMP-analysis data on floral assemblages (Herman et al., 2016).

(16) The *Late Maastrichtian climatic maximum* is confirmed by the data of CLAMP-analysis of floral assemblages (Herman et al., 2016) and by the isotope paleothermometry data (10.2–16.9°C, Koryak Highland; 18.1°C (Alaska) (Zakharov et al., 2011).

(17) The *Late Maastrichtian climatic minimum*. The second half and the end of Maastrichtian is the period of relative cooling; the paleotemperature value is 6.7°C (Herman et al., 2016). The sharp decrease in temperature at the end of the Maastrichtian is correlated with an impact event, which is recorded on the global paleotemperature curve (+11°C (Scotese, 2015)) and the curve constructed for the Arctic and Subarctic (+7.5°C (Herman et al., 2016)).

(18) The *Danian–Selandian climatic maximum*. The floral assemblages of the Danian Stage are known from the localities in the lower reaches of the Anadyr River, on the Rarytkin Ridge (Golovneva, 1994), in the lower reaches of the Lena River (Budantsev, 1983), in Western and Northern Greenland and in the Spitsbergen Archipelago (Budantsev, 1983); Budantsev and Golovneva, 2009). Golovneva [1994] presented the following climatic conditions for the lower reaches of the Anadyr River: an average annual temperature of 11.5°C, an average winter temperature of 4.6°C, an average summer temperature of 19.4°C, and an average annual precipitation of 1722 mm. The Spitsbergen territory is characterized by an average annual temperature of 12.6°C (an average winter temperature of +6.5°C, an average summer temperature of +19.8°C, and an average annual precipitation of 1826 mm). The Kharaulakh flora from the lower reaches of the Lena River (Akhmetiev, 2004; Budantsev, 1983) is characterized by the following paleotemperature values: an average annual temperature of +10–12°C (up to +14°C) at an average winter temperature of up to –4–6°C and an average annual precipitation of 1400 mm. Based on the palynological data obtained for the West Siberian Plain (Volkova, 2011), the average annual temperature is +15°C, the average winter temperature is +4–5°C, the average summer temperature is +28–30°C, and the average annual precipitation is up to 1826 mm

(19) The *Thanetian climatic minimum*. The Arctic flora in the Late Paleocene became moderately warm, transitioning to subtropical (Akhmetiev, 2004). The flora from the central and southern (Kupriyanov Island) parts of Alaska is characterized by the presence of representatives of a warm temperate climate, as well as palms and cycads. The average annual temperature for existence of these floras (Wolfe and Dilcher, 2001) is +10–12°C; the average annual precipitation is up to 2000 mm. In addition, the Thanetian flora is known in Scotland (Isle of Mull): the average annual temperature is +10.3°C at the average winter temperature of +2.8°C and an average summer temperature of +18.8°C and an average annual precipitation of 1739 mm (Golovneva, 2000).

(20) The *Early Ypresian climatic maximum*. The Early Eocene (Ypresian time) was characterized by a moderately warm humid climate. In the territory of Kamchatka (Late Paleocene–Early Eocene Napanian flora) the average annual temperature is +11.8°C, the average winter temperature is +5°C, the average summer temperature is +19.1°C, at an average annual precipitation of 2048 mm (Golovneva, 2000).

(21) The *Late Ypresian climatic minimum*. The annual average temperature in the territory of the Spitsbergen Archipelago (Sturvol flora, Eocene) is 9.5°C (an average winter temperature of 1.5°C, an average summer temperature of 18.4°C, and an average annual precipitation of 1716 mm (Budantsev and Golovneva, 2009). For the floral assemblages from Axel Heiberg and Ellesmere islands (Northern Canada) the following temperature parameters were calculated: the average annual paleotemperatures are +8.2°C; +9.3°C, the average cold temperatures are 0.8 and –2.0°C (Greenwood and Wing, 1995).

(22) The *Lutetian climatic maximum*. Based on the palynological data, the temperature parameters in Ayon Island during the Eocene were as follows: an average annual temperature of 13°C, an average winter temperature of 5 and 7°C, while the average summer temperature is 21–23°C. In the Middle Eocene, in the lower reaches of the Pur River (Yamal Peninsula) the average annual temperature reached 14–15°C, the coldest month temperature is 6–8°C, and the warmest month is 21–23°C (Volkova, 2011).

(23) The *Bartonian–Rupelian climatic minimum*. In the lower reaches of the Lena River (Tastakh flora, Middle–Late Eocene: the average mean temperature was up to 12°C, while the average winter temperature was +6°C (Budantsev, 1999). According to (Akhmetiev,2004), the average annual temperature for the Middle Eocene floral assemblages from the northernmost areas was 10–12°C, (an average annual temperature of approximately or just below 0°C and an average annual precipitation of 1000–1500 mm). The territory of the Spitsbergen Archipelago is characterized by gradual decrease in temperature up to 8°C to the end of Eocene (Budantsev and Golovneva, 2009). Based on palynological data, it was established that in the Western Siberia in the first half of Oligocene the average annual temperature varied from 10 to 15°C (the average warm temperature was 15°C, while the average cold temperature is +10°C) (Volkova, 2011).

(24) The *Chattian climatic maximum* was traced in sections of the Western Siberia. Based on the palynological data obtained for the Late Oligocene the average annual temperature was 14–15°C, while the average cold temperature was 4–6°C (Volkova, 2011).

(25) The *Aquitanian–Burdigalian climatic minimum*, recognized on the basis of the palynological data in the sections of Western Siberia, is characterized by

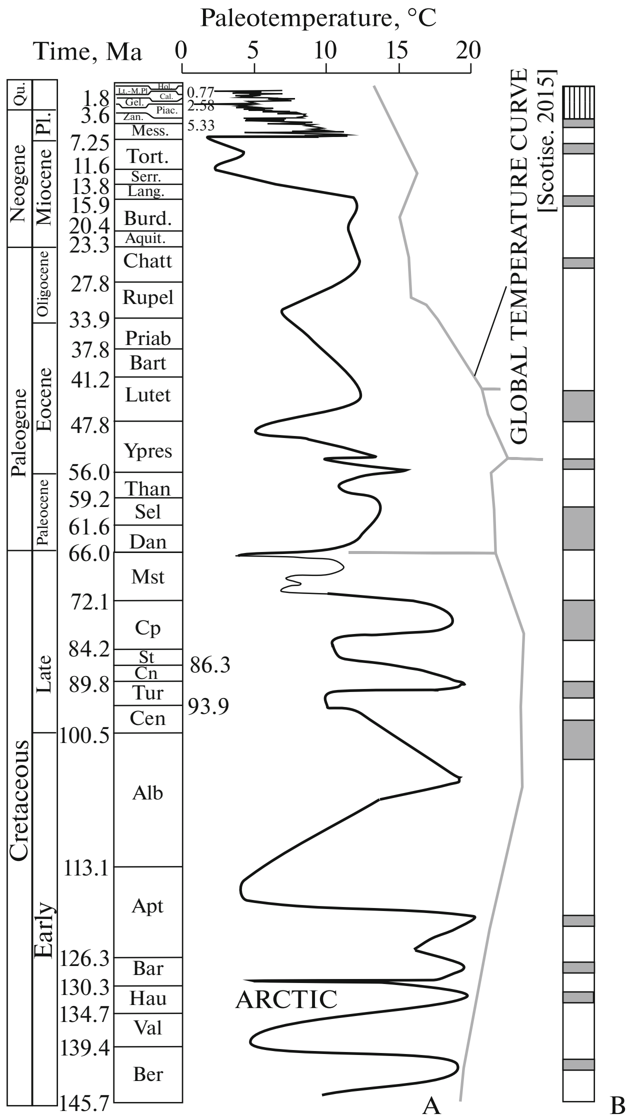

Fig. 2. Paleotemperature data for the Cretaceous–Cenozoic: A, the proposed combined curve for the Arctic Region; B, the global curve after (Scotese, 2015).

a slight decrease in temperature approximately by 1°C (up to +14°C) on both the regional paleotemperature curve (Volkova, 2011) and the global curve (Scotese, 2015).

(26) The *Langhian climatic maximum* was determined based on marine palynomorphs in the ODP Hole 907A section (North Atlantic) with a temperature of approximately 17°C (Shreck et al., 2011) and on the palynological data from the sections of Western Siberia of +11–15°C (Volkova, 2011).

(27) The *Serravalian–Early Tortonian climatic minimum* (a temperature of approximately 12–16°C) was established based on finds of marine palynomorphs (Shreck et al., 2011) in ODP Hole 907A section (North Atlantic). The cooling period was in a temperature range from 12.3 to 11.6 Ma, and the global temperature minimum is referred to a temperature range of 11.8–11.4 Ma; the surface water temperature varied in a range of + 9–15°C. The following temperature parameters were established for the Middle Miocene (13.4 Ma): annual average temperature values at the water surface equal to +20°C, the average summer temperature was +15–25°C. Based on the palynological data, the average annual temperature in the Western Siberia was equal to +10–12°C (Volkova, 2011).

(28) The *Middle Tortonian climatic maximum* was recognized on the basis of the marine palynomorphs in the ODP 907A well section (North Atlantic) with a temperature of approximately +17°C (Shreck et al., 2011) and based on the palynological data in the sections of Western Siberia, a temperature of 6–7°C (Volkova, 2011).

(29) The *Late Tortonian–Early Messinian climatic minimum* was recognized on the basis of marine palynomorphs (Shreck et al., 2011) from the ODP 907A well section (North Atlantic) with a temperature of approximately 1.5°C. Based on the palynological data, the average annual temperature in the Western Siberia was 5°C (Volkova, 2011).

(30) The *Late Messianian–Zanclian climatic maximum* (a temperature range from +7 to +13°C) was recognized on the basis of marine palynomorphs in the ODP Hole 907A section (North Atlantic) (Shreck et al., 2011).

(31) The *Piacenzian–Quaternary climatic minimum* (in a temperature range from +4 to +12°C) and the glaciation period in the Northern Hemisphere were established on the basis on membrane lipids of glycerol and dialkylglycerol tetraether in marine sediments of the North German Depression (Hank well) (Crampton-Flood et al., 2018) and on marine palynomorphs (Shreck et al., 2011) in the ODP Hole 907A section (Northern Atlantic).

Finally, 16 climatic cycles, including 16 climatic minimums and 15 climatic maximums, were established in the climatic history of the Arctic. Taking the fact into consideration that some paleotemperature values were obtained for sections in the Subarctic region (with more southern paleolatitudes), the paleotemperature values were corrected at the construction of a combined paleotemperature curve (Fig. 2) for the Arctic. Here, the following climatic variations were taken into consideration: a regular decrease in temperature by 1°C on average in the high latitudes, by 1.5°C in the middle latitudes, by 2°C and more (Herman and Spicer, 2012) in the low latitudes, when shifting laterally preliminarily by 10° from the equator to the pole.

The position of two fragments of the paleotemperature curve for the Rupelian–Piacenzian interval of the geological history of the Arctic was corrected. In particular, the temperature curve (Hank well) (Crampton-Flood et al., 2018) for the Messinian–

Piacenzian interval, as well as a fragment of the temperature curve constructed for Western Siberia (Volkova, 2011) for the Serravalian–Tortonian interval were shifted by 3°C towards the decrease in temperature values. The correction was based on the core data from the ODP 907A well section (Shreck et al., 2011).

The correlation between the paleotemperature curve constructed for the Arctic and the global paleotemperature curve (Scotese, 2015) is satisfactory: the general trends coincide and the cyclicity of climate variations of a lower order was recognized on the local curve constructed for the Arctic. The only exception, a decrease in the temperature in the Tortonian time, is likely due to local factors.

The reference of climatic events to the timescale is still conditional to a great extent and the duration of any event is rather difficult to determine. In the future, when correlating the climatic events with tectonic events and the variations in the World Ocean level it will be possible to tie the climatic variations to the timescale more exactly.

CONCLUSIONS

(1) In order to assemble the paleotemperature database for the Arctic and Subarctic regions we used data on marine invertebrate skeletons, marine palynomorphs, and dinosaur teeth, as well as those based on the results of analyzing the ability of reptiles to lay eggs at low temperature, terrestrial floral assemblages (CLAMP-analysis), the presence of coal interbeds in the Arctic continental deposits, and membrane lipids of glycerol and dialkylglycerol tetraether in marine sediments and glendonite.

(2) Despite the fact that different methods and approaches were applied for the determination of paleotemperatures, in general, they show a good correlation. In most cases, the average annual temperature was determined, which makes it possible to use all these data together (except for the isotope paleothermometry based on marine invertebrates for the Cretaceous and Eocene and surface water temperature vales estimated on marine palynomorphs for the Neogene).

(3) In total, 16 climatic cycles, including 16 climatic minima (including the glaciation period in the Northern Hemisphere) and 15 climatic maxima, were recognized in the climatic history of the Arctic.

(4) The local paleotemperature curve was constructed for the Arctic region. The general trends of this curve are consistent with the global paleotemperature curve (Scotese, 2015) (except for the cooling period in the Tortonian time due to local factors).

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