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Trends in Global Climate Changes Inferred from Geological Data

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Abstract—Recent paleoclimatic data reveal the following trends in climate changes. (1) During three billion years, characteristic of the Earth was gradual global cooling with the increasing frequency, duration and scale of glaciations. Based on these features, three principal climatic stages can be defined in geological history: (a) non-glacial (Early Archean), (b) with episodic glaciations (Late Archean–Middle Riphean), and (c) with frequent periodical glaciations (Riphean–Recent). (2) Irreversible climate changes were complicated and disguised by numerous superimposed temperature fluctuations of different periodicity and amplitude. In the Phanerozoic, a hierarchy of subordinate climatic fluctuations of 10–12 ranks, from extremely long (few hundreds million years) to short-term (tens years long only) is defined. Signs of climatic fluctuations of two–three highest ranks are recognizable in Proterozoic glacial sections. (3) The hierarchy of climatic fluctuations was stable during the Phanerozoic at least. (4) Amplitudes of climatic fluctuations depended on the cophasing degree of elementary climatic oscillations and character of their feedbacks in the biosphere. (5) The warm non-glacial climate prevailed during the Precambrian and Phanerozoic and was characteristic of 90% of the Phanerozoic and 95% of the post-Archean geological history. (6) Many climatic fluctuations, all those of first rank included, were of a global scale, synchronous, and cophasal. (7) Regional climate changes were caused by paleogeographic factors. (8) Global climate changes resulted in transformation of the latitudinal climatic zonality. The notion "global climate" is introduced to characterize the type of a planetary climatic zonality. Macrogeographic factors transformed latitudinal climatic belts into sublatitudinal ones. (9) Two main types of global climate (non-glacial and glacial) are defined. Transitions from non-glacial to glacial climate and vice versa were accompanied by rapid qualitative zonality reorganizations. (10) Each type of global climate is subdivided into gradations. (11) A peculiar feature of the global climate was an asymmetric position of climatic belts relative the equator. The asymmetry, which was insignificant during the non-glacial periods, and substantially increased at the time of glaciations, particularly of the great ones.

Key words: irreversible climate changes, quasi-periodic fluctuations, their hierarchy, global, synchronous, and cophasal patterns, climatic zonality, its reorganizations, asymmetry.

INTRODUCTION

General features of the Phanerozoic climate evolution became understood in the second half of the last century (Schwartzbach, 1974; Monin and Shishkov, 1979; and others) when several hypotheses were proposed to explain the climate changes (Monin and Shishkov, 1979; Budyko, 1980; Frakes *et al.*, 1992; and others). During the last decade, factual and methodological bases of paleoclimatology widened significantly and underwent qualitative changes. It was a consequence of research in difficultly accessible regions, deep-sea drilling, and of wide application of new approaches and methods (paleogeographic reconstructions based on plate-tectonic concept, geochemistry of stable isotopes, bed-by-bed study of sections using different methods, computer simulation of climate models, multivariant mathematic analysis of paleobotanic and lithological data, and so on). The quantity and quality of paleoclimatic data substantially increased. Accordingly, we got an opportunity to outline some general trends in climatic history of the Earth. The trends of empirical character are best detectable in the Phanerozoic history, although some of them are recognizable also in the Proterozoic, particularly in its late part. Some trends are distinct, while the others need in a further confirmation and should be considered as working hypotheses.

Particularly plentiful are data on the history and dynamics of glaciations, salt accumulation, stable isotopes, sedimentation settings, and paleobiogeography. These materials allow qualitative, semiquantitative, and, less commonly, quantitative estimates of climate temperature and humidity. The main attention is paid in this article to temperatures, because this leading factor is responsible for the regime of the Earth climate system and determines other climatic parameters, scale and intensity of global heat and moisture transfer inclusive.

1. DYNAMICS OF CLIMATE CHANGES

Well recognizable in the geological retrospective are irreversible climate changes and superimposed quasi-periodic climate fluctuations.

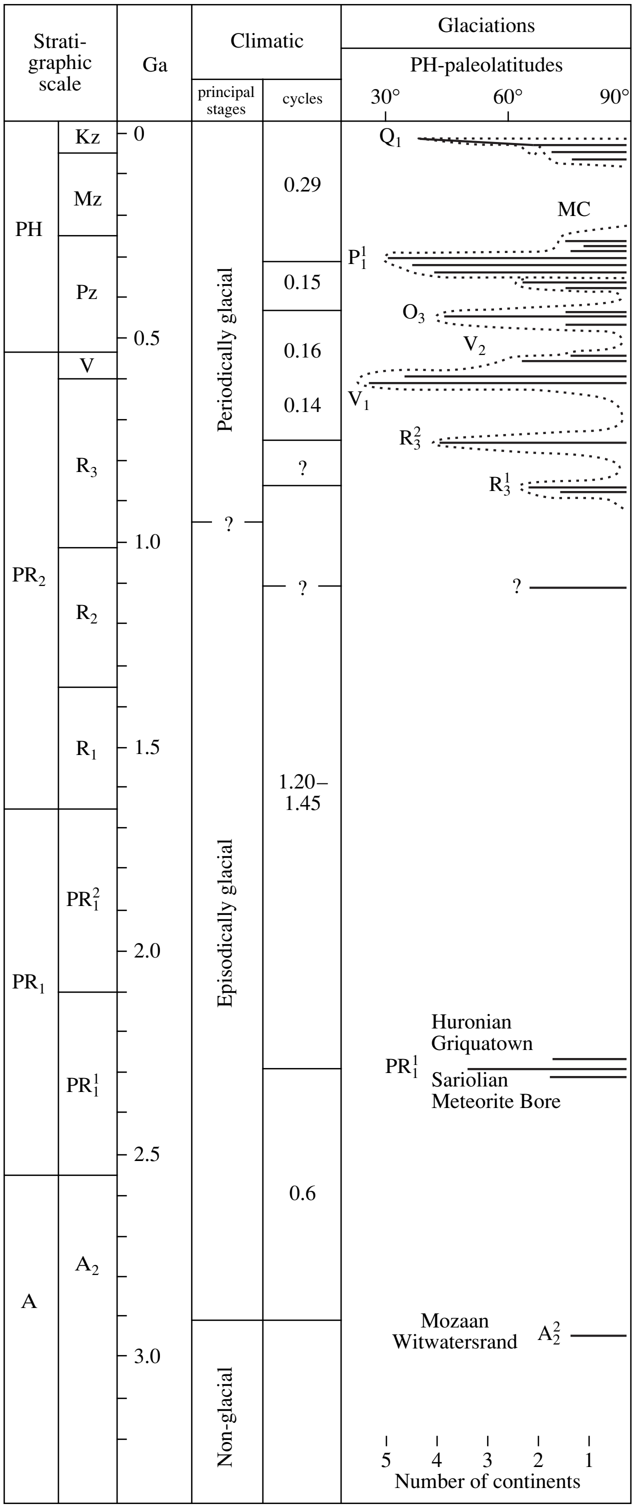

Fig. 1. Distribution of glaciations throughout the geological history; length of thin solid lines designating the glacial epochs is proportional to maximum distribution of glaciations (to paleolatitudes of the Phanerozoic and to number of continents in the Precambrian) during the given time, and dotted line envelopes glacial epochs of the Archean (A) and Mesozoic (MC) glacial periods.

1.1. Irreversible climate changes. The general trend in climate changes on the Earth can be inferred from glaciations. Their frequency and scale increased generally with time (Fig. 1), and this general trend cannot be a result of insufficient knowledge of older rocks only. During the last two–three decades, detailed geological mapping and prospecting covered almost the entire land areas composed of old rocks, those in developing countries included. Numerous and different new mineral deposits were discovered. The detailed studies could not miss glacial sediments, which form usually large bodies, are characterized by regional distribution, and attract attention of geologists by their unordinary appearance and origin. They showed that glaciations increased in scale during the geological history and their time structure was complicating. In addition, the studies of the last 30–40 years substantially specified ages and distribution areas of glacial sediments, although their new stratigraphic levels have not been discovered. An assumption based on astronomical hypothesis that the Earth glaciations started as long ago as in the Early Archean and repeated, beginning from 3460 Ma ago, every 400–370 m.y. (Steiner, 1978) has not been confirmed.[1] There is no also geological evidence of durable global glaciation on the early Earth, which was assumed by many researches in connection with the idea of a "weak" early sun. The planet surface was likely so "cool" in the Early Archean (Valley *et al.*, 2002) that its temperature was below 100°C, i.e., suitable for existence of hydrosphere at the surface.

The aforesaid allows a conclusion that available data reflect, with an appropriate approximation, a real distribution of glaciations in geological history.

The reliable signs of Early Archean glaciations are unknown so far. Their first signs, though scarce and spatially restricted, are established in the Upper Archean Witwatersrand Supergroup (Hambrey *et al.*, 1981) and Mozaan Group (Young *et al.*, 1998) of the relatively small Kaapvaal craton of South Africa. These glaciations are estimated to be approximately 2.9 Ga old (Nelson *et al.*, 1999). The Witwatersrand glaciation was probably of the piedmont or mountainous types while the Mozaan glaciation was of the ice sheet type because the last group encloses abundant dropstones (signs of ice rafting).

The next glaciations occurred only in the initial Early Proterozoic, i.e., 600–700 m.y. later than the Late Archean glaciations. The Early Proterozoic glaciations were of a substantially larger scale and mostly of the ice sheet type. Sediments they left behind are discovered on four remote continental blocks: on the Canadian Shield (Young, 1970), in South Africa (Visser, 1981), on the Baltic shield (Marmo and Ojakangas, 1984), and

[1] I use in this article the Russian stratigraphic terminology and Russian Precambrian geochronometric scale (Semikhatov *et al.*, 1991; Semikhatov, 2000; *Reshenie III Vserossiiskogo…*, 2001). For the Phanerozoic, the geochronometric scale by Harland *et al.* (1990) with some specifications for the Cambrian (Semikhatov, 2000) is used.

in western Australia (Martin, 1999; Eriksson *et al.*, 1999). Previously, I have considered these and the Witwatersrand glaciations as characterizing one into the Canadian Glacioera of the Early Proterozoic age (Chumakov, 1978). Now, it is established that the Witwatersrand (the Kaapvaalian, in general) glacial sediments are significantly older, Archean in age, whereas the Lower Proterozoic glacial sediments proper are estimated, based on generalized radioisotope data, to be within the narrow interval of 2.3 to 2.2 Ga (Crowell, 1999). Therefore, the Kaapvaalian and Early Proterozoic glacial events should be regarded separately as corresponding to the Kaapvaalian and Canadian glaciation periods, but not to one Canadian Glacioera. Similar C-isotope anomalies in carbonate rocks overlying sediments of the Canadian glacial period confirm their approximate synchronism (Semikhatov *et al.*, 1999; Bekker *et al.*, 2001). In North America, the Canadian glacial sediments are known from several remote regions. In the Great Lakes area, the Huron Supergroup encloses sediments of three large glacial epochs separated by interglacial deposits. In turn, the Late Gowgandan glacial epoch comprised two glacial events (Young, 1970).

Glaciation indications were never discovered at the higher levels of the Lower Proterozoic and in the Lower and most Middle Riphean sections (Fig. 1). The lack of large C-isotope anomalies within the stratigraphic interval of 1000–1200 m.y. long that is termed sometimes as the "glacial pause" indicates likely the absence of glaciations (Bekker *et al.*, 2001). According to brief descriptions published (Salop, 1973; Akhmedov, 2001; and others), scarce scattered boulders in the upper Lower Proterozoic sections are most likely of the landslide and volcanogenic origin or dispersed by seasonal ice, being indicative of the moderately cold though non-glacial climate.[2] Some researchers assumed that the "glacial pause" was caused by the elevated content of methane in the atmosphere, whereas glaciations, which bounded the "pause," were to the contrary related to episodes of atmospheric methane oxygenation (Pavlov *et al.*, 2003). Nevertheless, a wide distribution of red beds (Eriksson *et al.*, 1992; and others) and even tropical laterites (Beukes *et al.*, 2002) in this stratigraphic interval, which indicate the existence of oxidizing atmosphere at corresponding time, is inconsistent with this hypothesis.

It is unclear so far weather glaciations existed in the Middle Riphean or not. In the Baikal–Patom Highland and Brazil, there are glacial sediments, which can be as old as the Middle or Late Riphean (Chumakov, 1993a; Khomentovskiĭ and Postnikov, 2001). Beginning from the Late Riphean until Recent, glaciations occurred on the Erath regularly, and their scale increased with time. During their maximums, the glaciations spread over large territories (4–5 continents at once) to the latitude of 40°–30° and, sometimes, even to lower latitudes.[3] The glacial periods became complicated, consisting of multiply repeated glacial epochs and subordinate smaller-scale glacial events.

Thus, during the last three billion of years, the increasing role of glaciations characterized the main trend in climate changes on the Earth. This indicates a gradual cooling of the planetary surface. During the first two billion of years, glaciations were rare. The total duration of glacial periods corresponds to 7%, not more, of this time interval. Approximately 1 Ga ago, cooling was substantially intensified, and glacial periods in total lasted longer, approximately 25% of one billion of years long interval and 30% of the Phanerozoic history, i.e., during the last 535 Ma (table).

Judging from the facts and considerations mentioned above, three principal stages are recognizable in the Earth history: (1) *non-glacial* (Early Archean); (2) *with rare episodic glaciations* (Late Archean–Middle Riphean); and (3) *with frequent periodic glaciations* (Late or Middle Riphean to present time).

The main factors responsible for global cooling were probably the changes in atmosphere composition, particularly the CO_2 content decrease. This could be related to the lowering intensity of volcanism (Abbote and Isley, 2002a) and endogenic degassing of the Earth, on the one hand, and to intensified burial of organogenic carbon and carbonates in the sedimentary shell, lithosphere, and mantle in the course of plate tectonics development, on the other. This is evident from correlation between the principal climatic stages mentioned above and main stages in evolution of plate tectonics (Chumakov, 2001b). The intensity of carbon burial increased with development of weathering processes and photosynthesis. Global cooling could also be stimulated by the planetary albedo increase with the growth of continents. Reduction of endogenic heat flow could also contribute to cooling, although this process was subordinate, because the endogenic heat proportion in the Earth's heat balance was by two–three orders lower than the heat contribution produced by insolation.

1.2. Periodical climate changes. The general trend of the Earth surface cooling was not unidirectional. It was complicated and disguised by numerous climate changes of different signs and scales. Most of these changes occurred within certain time limits, and the state close to the initial one returned some time later. Therefore, all or almost all climate changes can be con-

[2] If it appears that some shales with boulders occurring in the upper part of the Lower Proterozoic section are of glacial origin, this will not change the conclusion that the frequency and scale of glaciations gradually increased during the post-Archean interval of geological history.

[3] Nevertheless, the hypothesis of global glaciations in the Vendian and Late Riphean (Harland, 1964; Hoffman *et al.*, 1998; and others) cannot be considered as proven because it is still based on scarce (Evans, 2000) and doubtful (Meert and Van der Voo, 1995) paleomagnetic determinations and is poorly consistent with several geological facts (Chumakov, 1992, 2003) and with results of climate mathematical modeling (Poulsen, 2003; and others). Data in favor of the Early Proterozoic global glaciation (Evans *et al.*, 1997) are even less reliable.

Glacial and non-glacial climatic periods and associated events

Age, Ma	Glacial (g[1]) and non-glacial (t)	Duration, m.y.	Polar continents[3]	Super continents[2]	CO_2 content, n[4]	Maximums		
						tectogenesis[5]	volcanism	
							subduction-related	mantle
0								
$Kz_{2–1}$	ggggg	40	++	+	1	+	+	
40			+					
$Kz_{2–1}$–Mz	t	205	++	++	5–7			++
245						++		
Pz_3^2	ggggg	78	++	++	1	+		
323								
C_1	t	30	++	+	2–4			
353								
C_1–D_3	ggggg	14	++	+	4–7		+	
367								
D_3–S_3	t	59	+	+	7–12	++		+
426								
S_2–O_3	ggggg	18	+	+	12–17	+	+	
444								
O_3–Cm_1	t	86	+	+				
530								
Cm_1–V_2	gg	10 ?	+	+				
540								
V_2	t	45	+					+
585								
V_1	ggggg	15 ?	+	+				
600								
$R_3^{2–3}$	t	140						+
740								
R_3^2	gggg	20 ?		+				
760								
$R_3^{1–2}$	t	130		++				+
890								
R_3^1	ggg	20 ?						
910								
$R_{3–2}^1$–PR_1^2	t	1290						+
2200								
PR_1^2	gggg	100						
2300								
PR_1^1–AR_3	t	600						+
2900								
Ar_3^2	ggg	50 ?						
2950								

[1] the number of letters designates approximate glaciation scale; [2] Gondwana-type (+) and Pangea-type (++) supercontinents; [3] glaciations on one (+) and two (++) poles; [4]n present-day CO_2 content in the atmosphere; [5] significant (+) and maximal final (++) phases.

sidered at first approximation as climatic fluctuations of different periodicity. The periodicity notion is used here *sensu lato* as it is usually understood in geology. In reality, these fluctuations were not strictly periodical, harmonic, and, consequently, linear. This is evident from the patterns of climatic secular successions and from bifurcations of oscillation periods inferred from their spectrum–time analysis (Chumakov and Oleinik, 2002). Strictly speaking, the fluctuations should be termed as self-similar or similar. Periodic fluctuations are sufficiently well established in the Phanerozoic and, to a lesser extent, for the Vendian and Late Riphean; sometimes, they are recognizable in older intervals as well. The spectrum of periodic climatic fluctuations was very wide: from several tens years to hundreds million years. Different-period fluctuations were superimposed on each other (Fig. 2) and, as a rule, non-harmonic. These features and approximate character of many geochronological dates hamper the exact determination of the fluctuation periods duration. Nevertheless, oscillations of several ranks and their groups are usually well distinguishable in paleoclimatic curves (Douglas and Woodruff, 1981; Zakharov, 1992; Chumakov, 1993b; Fot'janova and Serova, 1994; and many others) as well as in spectrum (Imbrie *et al.*, 1984; and others) and spectrum–time (Chumakov and Oleinik, 2002) diagrams. This implies that many fluctuations substantially differed from each other in duration. They are divisible into groups differing from each other in fluctuations average duration that varied several-fold and, sometimes, by an order of magnitude. Let us consider periodical climatic fluctuations, which are recognizable by geological methods, beginning with the large-scale ones. It is convenient to distinguish fluctuations of the following five groups.

Superlong climatic fluctuations. The idea of a large-scale climatic periodicity was formulated long ago. Holmes (1937), Umbgrove (1947), and Lungershausen (1956) were among the first to propose such an idea based on distribution of glaciations through geological history. Because of poor knowledge of ancient glaciations and insufficient isotope dating, the suggested periodicity was largely intuitive. These authors estimated the duration of climatic (glacial) cycles as ranging from 250 to 190–200 m.y. The other estimates published afterward are 300 (Keller, 1972), 300–1200 (Avdeev, 1973), 280–400 (Steiner, 1978), 217 (Zakoldaev, 1991), and 215 m.y. (Yasamanov, 1993). Assessments of Yasamanov and some other researchers are based on incomplete and partly outdated paleoclimatic and geochronological data. In addition, these data were selected and interpreted rather *ad arbitrum* leaning upon the duration of the current galactic year or deductively calculated duration of past galactic years. It should be noted that astronomers estimate very approximately (from 180 to 300 m.y.) even the duration of the current galactic year. It is more reasonable therefore to define the large-scale climatic periodicity using an inductive approach based on trustful paleoclimaic data. Estimating time intervals between dates most frequently mentioned for the Late Precambrian and Phanerozoic glaciations, Williams (1975) arrived at the conclusion that climatic cycles were approximately 150 m.y. long. Afterward, Frakes *et al.* (1992) estimated that time spans between the middles of Phanerozoic and Late Precambrian cold and warm periods range from 138 to 181 m.y. Based on the Phanerozoic curve of secular $\delta^{18}O$ variations, some researchers suggested that prevalent climatic cycles lasted approximately 135 ± 9 m.y. (Shaviv and Veizer, 2003).

Because of significant scatter of estimated duration of climatic cycles, some authoritative researchers cast doubts on existence of any large-scale climatic periodicity at all (Harland, 1981; House, 1995; Crowell, 1999). Their skepticism is not quite sound. It was already mentioned why the deduced climatic cycles are of discordant duration. As for the empirical approach, discrepancies are related, largely, to the fact that transitions from warm to cold intervals and vice versa were rather gradual and lasted tens of million years. In addition, transitional periods were often poorly studied and duration of cold and warm epochs was estimated very arbitrarily, i.e., subjectively. When periodicity is established based on most prominent and accurately dated climatic events, par example, on glacial maximums, the discordance can be reduced substantially. Let us consider this problem in more detail. There are known five glacial maximums of the Late Riphean–Paleozoic time termed sometimes as "great glaciations": (1) first Late Riphean event of this kind was 850–890 Ma ago, (2) second Late Riphean 740–750 Ma ago, (3) the Early Vendian 600 Ma ago, (4) Late Ordovician, 440 Ma and (5) Late Paleozoic 290 Ma ago. The last four "great glaciations" are separated by time intervals of 140, 160, and 150 m.y., respectively. The method substantially decreases the scatter of cycles duration. In fact, it appears to be within accuracy limits of stratigraphic methods. Accordingly, the existence of a regular climatic periodicity within the Late Riphean–Paleozoic interval quite apparent (Chumakov, 2001a). It is probable that the interval between the first and second Late Riphean glaciations was also as long as 140–150 m.y. The last Paleozoic (Asselian–early Sakmarian) glacial maximum is separated from the next (Pleistocene) one by the time interval of 290 m.y. If we take into consideration the fact that signs of cooling and, locally, even of seasonal ice rafting are recorded in the terminal Jurassic–initial Cretaceous, i.e., in the middle of this interval (Hambrey *et al.*, 1981; Frakes *et al.*, 1992), then two additional Mesozoic–Cenozoic climatic cycles each 140–150 m.y. long can be outlined. These five or, probably, six climatic cycles lasting 150 m.y. are close in average duration to those established by Williams and Frakes with colleagues. Periods of similar duration are inferred also from the spectrum and spectrum–time analyses of semiquantitative paleoclimatic data available for the Late Riphean and Phanerozoic (Chumakov and Oleinik, 2002). The duration value

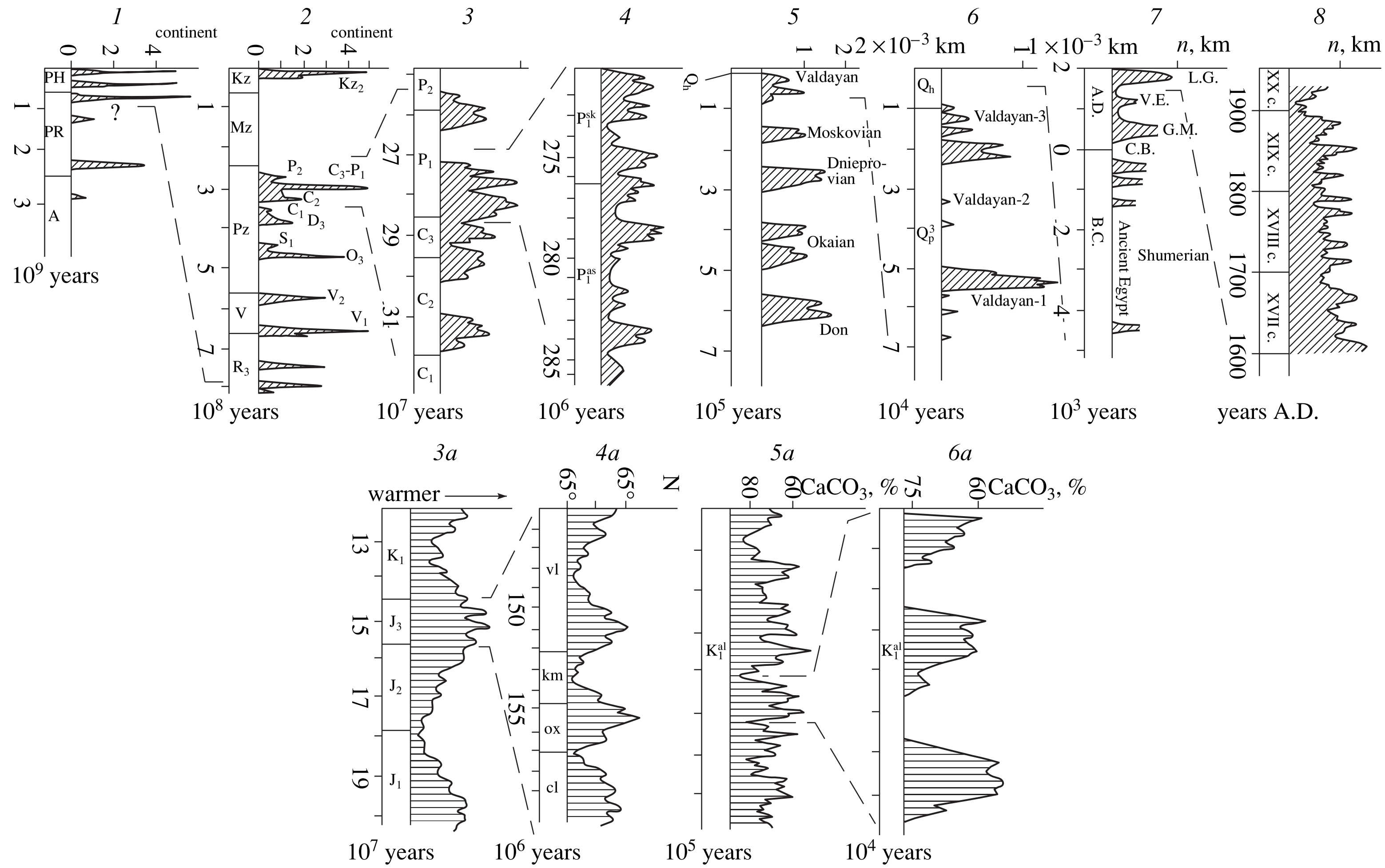

Fig. 2. Upper diagrams exemplify glacial events of different ranks (hatched) and relevant climatic fluctuations (Chumakov, 1993b with additions); lower diagrams illustrate climatic events and fluctuations of similar ranks in non-glacial intervals of the Phanerozoic (remember 10 times difference between scales of adjacent diagrams): (*1*) three principal climatic stages AR_1, AR_2–R_2, and R_3–Recent; (*2*) glacial and non-glacial periods, superlong climatic fluctuations; (*3*) glacial and non-glacial epochs, long climatic fluctuations, C_2–P_2, southeastern Australia; (*3a*) warmer and cooler epochs (thermal and semithermal epochs), J–K_1, Siberia (after Zakharov, 1992); (*4*) glacial and non-glacial ages, middle-ranged climatic fluctuations, P_1^{as}–P_1^{sk}, western Australia; (*4a*) warmer and cooler ages (thermal and semithermal ages), J_3, and Callovian (cl), Oxfordian (ox), Kimmeridgian (km), Volgian (vl) invasions of Tethyan fauna into basins of northeastern Asia (after Velichko *et al.*, 1994); (*5*) glacials and interglacials, short climatic fluctuations, Pleistocene, East European Plain; (*5a*) short (Milankovitch-type) fluctuations reflected in color variations and $CaCO_3$ contents in sediments of the *Ticinella praeticinensis* Subzone, K_1^{al}, Italy (after Larson *et al.*, 1993); (*6*) stadials and interstadials; short climatic fluctuations; Q_p^3, East European Plain; (*6a*) the same as *5a*, but of the lower rank; (*7*) secular fluctuations of present-day glaciers, Q_h, Tirol, (L.G.) "Little Glacial Age," (V.E.) Vikings Epoch, (A) Archean, (G. M.) great migration of people; (C. B.) Christ birth; (*8*) shorter fluctuations of present-day glaciers (17th to 20th centuries), Alps.

inferred from the Phanerozoic $\delta^{18}O$ curve (approximately 135 ± 9 m.y.; Shaviv and Veizer, 2003) is rather close to values mentioned above.

In addition to prominent cycles 140–150 m.y. long, there are also even longer, although less manifested cycles. Every second "great glaciation," beginning from the Vendian one, is of a slightly larger scale. Therefore, the largest glaciations (Pleistocene, Late Paleozoic, and Vendian) form two cycles approximately 300 m.y. long. The climatic cycles 140–150 and 300 m.y. long can be united, for convenience, into a group of superlong fluctuations (Chumakov, 1995a). They determined main climatic changes during the Phanerozoic, Vendian and Late Riphean.

The causes of superlong climatic fluctuations were discussed for a long time by many experts. The total number of published hypotheses about causes of glacial periods is close to a hundred (for critical review of most important hypotheses see Chumakov, 2002). Here, I consider briefly only some of them. As was noted, many researchers attempted before and are trying now to correlate superlong climatic cycles with the galactic year (Umbgrove, 1947; Lunsgerhausen, 1956; Keller, 1972; Zakoldaev, 1991; Yasamanov, 1993; and others) or with the galactic half hear (Williams, 1975; Frakes *et al.*, 1992). The cycles 140–150 m.y. long are however substantially shorter as compared with both synodic and anomalistic galactic years, whereas the comparison of glacial cycles with the galactic half year is illogical, unless additional complex assumptions are introduced (Williams, 1975; House, 1995).

Quite widespread is an idea that alternation of glacial and non-glacial periods was mainly determined by oceanic circulation (the latest example is publication by Smith and Pickering, 2003). This hypothesis does not take into account the fact of very close configurations of continents and oceanic currents at the commencement (Eocene, Antarctica) or termination (Permian, Gondwana) of some glacial periods.

The glacial periodicity correlates significantly better with the tectonic and volcanic activity of the Earth (Chumakov, 2001a). The glacial maximums were synchronous with early phases of tectonic cycles and peaks of subduction-related explosive volcanism. This leads to the conclusion that accumulation of thick sedimentary sequences during the early phases of tectonic cycles and associated intense weathering of silicates determined the burial of large volumes of carbonates and organic matter, reduced the CO_2 content in the atmosphere, and set the stage for glaciations (Lindsay and Brasier, 2002; Schrag *et al.*, 2002). The "volcanic winters" related to intensified explosive volcanism triggered glaciations. Owing to numerous and strong positive feedbacks in the biosphere, glaciations became stable and widened.

Glacial maximums predated the main final phases of tectonic cycles. During maximal phases of tectogenesis and orogeny, glaciations ceased or rapidly degraded. This can be explained by reduced subduction-related volcanism, increased atmosphere transparency, and elevated heat balance at the Earth surface. The processes that accompanied main orogenic phases, such as intense erosion, granite formation, and regional metamorphism in orogenic belts, resulted in oxidation of organic and carbonaceous matter in sedimentary sequences and in decomposition of carbonates in mixed terrigenous–carbonate deposits ("dedolomitization" and "decarbonatization"). Some periods were marked by the intensified magmatism of mantle plumes. The released carbon dioxide entered the atmosphere and stimulated additional warming. Being combined, these processes resulted in substantial warming and termination of glaciations.

It cannot be entirely ruled out that periodicity of tectonic and volcanic activity, and to a certain extent, the periodicity of global climate were controlled by cosmic (Abbot and Isley, 2002b) or galactic processes, for instance, by the solar system pass through sleeves of spiral star accumulations (Shaviv and Veizer, 2003).

"Great glaciations" corresponded to peaks of glacial epochs, duration of which varied in the Late Riphean–Phanerozoic from approximately 10–15 to 78 m.y. averaging approximately 27 m.y. The glacial epochs were separated by durable non-glacial periods (interglacials) that lasted in the Late Riphean–Phanerozoic from 30 to 205 m.y. averaging approximately 95 m.y. Climatic cycles corresponding to alternating glacial periods (Middle Mesozoic cooling included) and interglacial periods show no regularity and range from 55 to 180 m.y. Taking into consideration a rather strict periodicity of great glaciation maximums, irregularity of related climatic cycles of lower ranks represents a problem, which requires special discussion. Here, it should be noted only that this is probably caused by different combinations of paleogeographic, geochemical, and geodynamic factors that accompanied glaciations and influenced, to some extent, the latter.

Long climatic fluctuations. The abovementioned glacial periods consisted of separate shorter events: glacial and interglacial epochs, which lasted from a few to 10–15 m.y. (Figs. 1, 2). Biostratigraphic and geochronometric methods, which can be used to measure duration of glacial and non-glacial periods, are of insufficient resolution in many cases for determination of climatic epochs duration. The average duration of these epochs can approximately be estimated, when several of them are recognizable in a continuous sedimentary succession of determined time-range. For the Late Paleozoic, such a succession can be exemplified by the Hunter River valley section in eastern Australia, where glacial sediments repeatedly alternate with marine varieties (Chumakov, 1993b). Some researchers believe that basal layers of this succession are of the Namurian Stage. According to the other standpoint, the base of this glacial section is the Westphalian in age. The glacial section is crowned by the uppermost Kungurian or

Kazanian sediments. There are indications that iceberg-rafted coarse debris accumulated in southeastern Australia until the Tatarian Age (Veevers *et al.*, 1994). Thus, the integral duration of the Carboniferous–Permian glacial period in this region is estimated as ranging from 55 to 67–70 m.y. according to the most widely accepted scale (Harland *et al.*, 1990). In the Hunter River section, recognizable are three glacial units (tillites, iceberg-related, fluvioglacial, and glacial-marine sediments) and two intervening interglacial units (terrigenous marine and coaliferous sequences), i.e., five climatic epochs in total. The average duration of these epochs ranged from 11 to 14 m.y. Thus, the duration of climatic cycles of two, glacial and interglacial epochs can be estimated as 20–30 m.y. long. When recent data on probable Namurian age of basal layers in the Carboniferous–Permian Dwyka Formation (Streel and Theron, 1999) are taken into consideration, the similar average duration of 9 m.y. appears to be characteristic of four glacial and three intervening non-glacial epochs in South Africa (Visser, 1997). The cycles formed by these epochs are as long as approximately 18 m.y. Climatic fluctuations with periods of a few tens million years long, although of a lesser amplitude, are also recorded in ice-free realms. During the Late Paleozoic glaciations, such fluctuations were recorded in central Siberia, where their duration is estimated to be about 35 m.y. long (Chumakov, 1995a). As for Cenozoic glaciations, recognizable are climatic fluctuations with periods of 15 to 20 m.y. in northeastern Asia (Velichko *et al.*, 1994), of about 11 to 14 m.y. in Kamchatka and Sakhalin (Fot'janova and Serova, 1994), of 15 m.y. in Sakhalin and the Primor'e region (Krassilov, 1994), and in southern West Siberia and southern East Europe as well (Velichko *et al.*, 1994). Differences between the estimated durations can be explained by insufficient accuracy of available stratigraphic data and correlation of continental sediments, particularly of the Permian deposits.

Climatic fluctuations a few tens million years long were first united into a group of long-wave or long oscillations within glacial sections (Chumakov, 1993b), although they are characteristic also of non-glacial Phanerozoic periods. These fluctuations were manifested as alternation of more and relatively less warm climatic epochs. Taking into consideration the fact that differences between these epochs are relative only, they can be termed as thermal and semithermal epochs. Thermal and semithermal epochs are distinct in the Mesozoic of West Siberia (Gol'bert, 1987; Zakharov, 1992), in the Late Cretaceous of mountainous (Krasheninnikov *et al.*, 1990) and coastal (Herman, 1993) areas of northeastern Asia, in the Mesozoic–Cenozoic of low latitudes (Douglas and Woodruf, 1981), in the Falkland Plateau area of the Southern Hemisphere (Krasheninnikov *et al.*, 1990), in middle latitudes of the Indian Ocean (Clarke and Jenkyns, 1999), and in the Paleogene of Antarctica (Dingle and Levelle, 1998). For Siberia, their typical periods are estimated to be 11–35 m.y. long (Chumakov, 1995a) and their amplitudes as corresponding to several degrees (Gol'bert, 1987).

Like superlong climatic periods, the long-wave fluctuations were probably related to variations in intensity of tectonic and magmatic processes on the Earth (Chumakov, 2001a).

Middle climatic fluctuations. During the last decades, it was established that climatic epochs were also non-uniform and consisted of relatively cold and warm time intervals. Inasmuch as these intervals are subordinate to climatic epochs, it is logical to term them, according to accepted system, climatic ages: glacial and interglacial ages of glacial epochs, and semithermal and thermal ages of non-glacial epochs. Climatic ages were usually many hundreds of thousand to several million years long. Cycles of alternating cold and warm ages, which lasted from about a million to tens million years, were defined as middle-wave climatic fluctuations (Chumakov, 1993b). At present, there are numerous examples of such fluctuations characteristic of both the glacial and non-glacial climatic epochs (Fig. 2). In addition to the Late Ordovician, Silurian, Early Permian, Jurassic, and Late Cenozoic middle climatic fluctuations mentioned in previous publications (Chumakov, 1993b; Velichko *et al.*, 1994) they are established in many other regions. These fluctuations are 1.2 m.y. long in the Ordovician–Silurian of Arabia (Vaslet, 1990), about 1.3 m.y. long in the Silurian of South America (Grahn and Caputo, 1992), 2 to 5 m.y. long in the Aptian–Eocene deposits of Indian Ocean (Clarke and Jenkins, 1999; Chumakov and Oleinik, 2001), 4 to 10 m.y. long in the Late Cretaceous of the Far East (Zakharov *et al.*, 1999), 1.7 to 3.6 m.y. long in the southern Atlantic (Herbert *et al.*, 1999), 1.5 to 8 m.y. long in the Oligocene–Miocene worldwide (Zachos *et al.*, 2001), and 1 to 2.7 m.y. long in the Late Cenozoic of the Arctic (Velichko and Nechaev, 1999).

For a long time, researchers ignored the middle-wave climatic fluctuations flattening them in secular climatic curves and leaving aside by paleoclimatic reconstructions and climate mathematical modeling. Nonetheless, these fluctuations larger in scale than those of Milankovitch type prevail (over 50%) over the others (Chumakov, 1995a). They could be responsible for anomalous temperature oscillations of 10–12°C, as estimated for the Late Cretaceous of the Far East (Zakharov *et al.*, 1999), and of 3–6°C, as established for the Oligocene–Early Miocene deep oceanic sediments (Zachos *et al.*, 2001). Ice-rafting episodes in the Greenland and Norwegian seas were also related to the middle-wave climatic maximums (Velichko and Nechaev, 1999).

Causes of middle climatic fluctuations are unclear being under active debates. Some researchers argue that they are related to long-lasting variations in eccentricity of the Earth orbit (Herbert *et al.*, 1999). Remarkable is the proximity between periods of middle climatic fluctuation and periods of high-frequency eustatic fluctuations.

Short climatic fluctuations. Short climatic fluctuations from a few tens to hundreds thousand years long (Milankovitch fluctuations, after Imbrie *et al.*, 1984) are known for a long time as characterizing alternating glacials and interglacials of the Pleistocene and their phases in high and middle latitudes. These fluctuations were of a global scale and manifested, although to a lesser extent, in low latitudes and in all subsystems of the biosphere as well. On land, short climatic fluctuation are readily recognizable in loess–soil (Zubakov, 1986; Dodonov, 2001) and lacustrine (Karabanov *et al.*, 2000, Lowenstein *et al.*, 1999) sequences being also inferable from alternation of humid and more or less arid environments and from respective vegetative communities (Dupont *et al.*, 2000; Van der Kaars and Dam, 1995). In seas and oceans, these fluctuations resulted in changes of surface (Velichko and Nechaev, 1999) and deep-water (Chapman and Shackleton, 1999) temperatures, sea-level oscillations, and substantial changes in sedimentation patterns (Lisitsyn, 1988). They are also responsible for changes in faunal assemblages (Barash *et al.*, 1989; and others), primary bioproductivity (Beaufort *et al.*, 1997), variations of oxygen (Imbrie *et al.*, 1984) and carbon isotope compositions, and for other sedimentological, biotic, and geochemical events. In the atmosphere, short climatic fluctuations were accompanied by changes in the carbon dioxide, hydrogen, and dust content, and in the oxygen isotope composition, as it is inferred from investigation of gas inclusions in ice core samples (Kotlyakov and Lorius, 2000).

It was believed for a long time that short climatic fluctuations are characteristic of glacial periods only (Woldstedt, 1954; Velichko, 1987). At present, it is shown that these fluctuations prevailed, in various forms, during the entire Phanerozoic at least, in its glacial and non-glacial periods. Listed below are only some of numerous publications, where short climatic fluctuations of various pre-Pleistocene subdivisions are mentioned: Pliocene (Zubakov and Borzenkova, 1983; Raymo, 1992; and many others); Miocene and Paleogene (Zachos *et al.*, 2001); Cretaceous (Larson *et al.*, 1993; Mutterlose and Ruffell, 1999); Jurassic (Waterhouse, 1999); Permian (Anderson and Dean, 1995), Carboniferous (Weedon and Read, 1995; Miller and Eriksson, 1999), Devonian (Wu *et al.*, 2001), Silurian–Ordovician (Williams, 1991); and Ordovician (Sutcliffe *et al.*, 2000).

After works by J. Adhemar, J. Croll, and, particularly, by M. Milankovitch, many geologists began to relate Pleistocene glacial events to variations in orbital parameters of the Earth, the inclination of rotation axis included. The study of oceanic sediments in the 1970s showed that the oxygen isotope composition variations in carbonates, which reflect the ice volume on the planet, are well correlative with astronomical periods of 19, 23, 41, and, particularly, 100 thousands years (Imbrie *et al.*, 1984). Variations in paleotemperatures and several other climatically important parameters with a close periodicity were revealed by geochemical study of ice cores from Greenland and Antarctica (Kotlyakov and Lorius, 2000). In addition, a cycle 400–410 thousand years long is recorded in most complete sections of both the glacial and non-glacial sediments (Herbert *et al.*, 1999). These data, as well as numerous age estimates obtained by isotopic and other methods confirm earnestly the astronomical control of short variations. The fact that short climatic fluctuations occurred through the entire Phanerozoic regardless of repeated changes in geological, geographic, climatic, and biotic situations on the Earth is an additional argument in favor of the astronomical control.

Ultrashort climatic fluctuations (millennial, centennial, and shorter). These fluctuations are registered by instrumental observations and historical, archeological, dendrometric, glaciological, and geological data on the Holocene and Late Pleistocene (Broecker and Denton, 1989; Bianchi and McCave, 1999; Lee and Slowey, 1999; Kotlyakov and Lorius, 2000; and many others). Some of these fluctuations show certain correlation with the solar activity. The radiocarbon analysis of wood growth rings established, in particular, the cycles of relatively elevated solar activity, which are approximately 2400, 299, and 90 years long. By duration and phases, the first two cycles are almost identical to the Holocene climatic cycles (Dergachev, 1994; Vasil'ev *et al.*, 2002; Raspopov *et al.*, 2000). In older sediments, ultrashort fluctuations, for instance those, which are 200 years long, are recognized only in some cases (Anderson and Dean, 1995).

Weather fluctuations. The climate is characterized by average perennial meteorological parameters. Their changes with periods below 40 years are usually referred to weather fluctuations. As is known, dominant among the latter are annual (seasonal) and daily fluctuations. In addition, there are perennial cycles, some of which are also related to variations in the solar activity. Sometimes, seasonal (Chambers *et al.*, 2000; and many others) and perennial climatic cycles are recognizable, with a certain confidence, in older sediments as well.

Summing all the mentioned data on climatic periodicity, we can state that the Phanerozoic was characterized by the hierarchy of climatic fluctuations of 10–12 ranks at least (Fig. 3).

2. CLIMATIC FLUCTUATIONS IN THE PRECAMBRIAN

Currently, the problem of climatic fluctuations in the Precambrian is of particular interest. Many geochemical, stratigraphic, and paleogeographic interpretations, as well as an objective estimate of the hypothesis of global glaciations are connected with this problem.

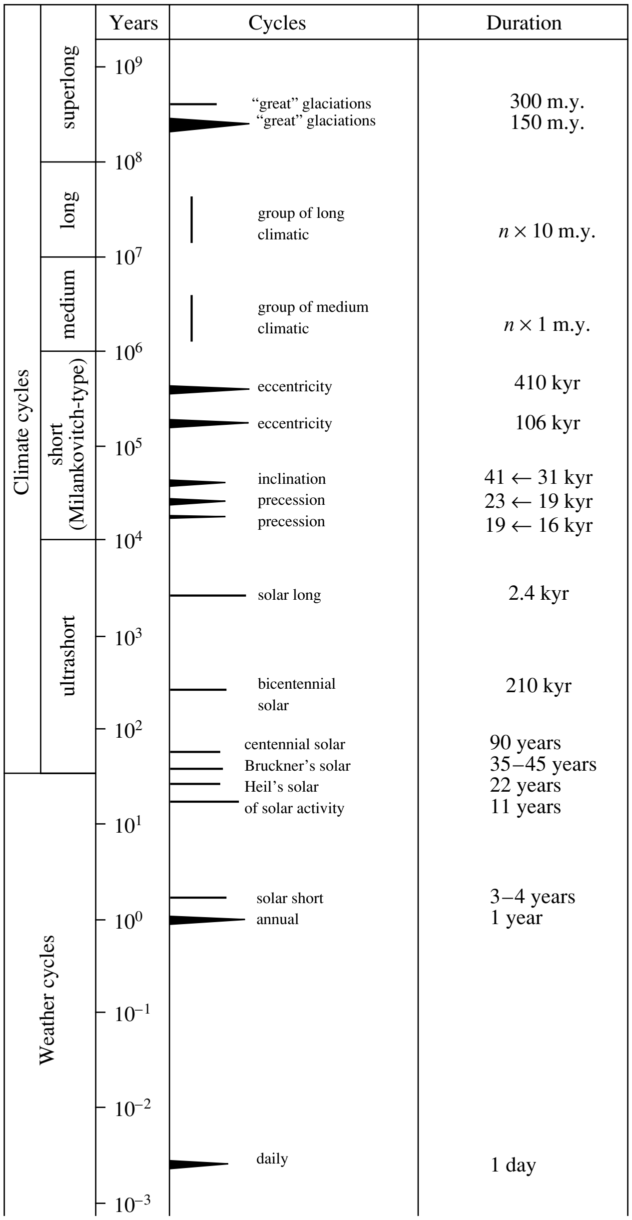

Fig. 3. Hierarchy of climate and weather cycles (thick solid lines designate the most important cycles).

The low resolution of biostratigraphic methods for this time interval and scarcity of reliable radiometric ages hamper usually a direct estimate of periodicity in Precambrian climatic fluctuations. Nevertheless, analogues of largest Phanerozoic climatic fluctuations can be recognized, with certain confidence degree, in the Proterozoic as well. Let us briefly consider some typical Precambrian glacial sections from the viewpoint of climatic fluctuations.

The Vendian and Riphean. The occurrence of two superlong climatic cycles approximately 140–150 m.y. long in the Vendian–Late Riphean is quite obvious (Fig. 1). Glacial and interglacial sediments alternating in most complete sections correspond to separate intervals of superlong Precambrian cycles and suggest existence of subordinate shorter cycles. By facies peculiarities, their combinations and thicknesses, such sections are sometimes indistinguishable from Phanerozoic glacial sections and were first erroneously referred to the Phanerozoic Eon (for instance, the Vendian glacial sequences of Middle and Central Asia were first considered as Carbonaceous–Permian in age).

The Late Vendian–Early Cambrian (Baikonur or West African) glaciation was complicated by climatic oscillations. Good examples are glacial sections in the Mauritanian Ardar composed of two main continental glacial formations separated by fluvial, lagoonal, and marine sediments and by unconformities (Deynoux and Trompette, 1981), which imply two glacial epochs at least. Even more intricate scenario is inferred for that glaciation from sections of western Mali by Proust and Deynoux (1994). They distinguished glacial epochs (a few million years long) and Milankovitch cycles approximately 100 ka long in the West African "glacial period" (20–30 m.y. long).

Similar climatic fluctuations are also recognizable in the Early Vendian Laplandian (Varangian) Glacial period. The Vil'chitsy Group of Belarus of that period consists of two units: the lower, Blon' Formation and unconformably overlying Glussk Formation (Fig. 4c). The first unit is composed of glacial sediments in its lower part and of interglacial sandstones and sandy dolomites in the upper part. In its most complete sections, the Glussk Formation encloses three till members separated by varved clays with dropstones and by well-sorted, probably fluvial sands with thin clay intercalations, rare ripple marks, and desiccation cracks (Fig. 4b). The uppermost layers of these members reveal glacial dislocations. Each member is of an intricate structure and consists of several layers different in composition and facies affinity, which are usually separated from each other by thin-bedded clay with glacial dislocations (Fig. 4a). It may be assumed that the Glussk and lower Blon' formations correspond to glacial epochs, whereas intervening strata characterize an interglacial epoch of the Laplandian Glacial period. In the considered case, combinations of different epochs represent long-period climatic fluctuations. A similar intricate three-unit structure in general is characteristic of the stratotype section of the Laplandian Glacial period in northern Norway, sections of Spitsbergen (Chumakov, 1978), eastern Greenland (Hambrey and Spencer, 1987), one of the most complete Laplandian sections in the Middle Urals (Chumakov, 1998), and Lower Vendian Nantuo glacial formation in southern China (Chumakov and Sergeev, 2004). Even more intricate are the Lower Vendian Fiq Formation in Arabia (Leather *et al.*, 2002), Blaini Formation in India (Kumar *et al.*, 2000), Yerelina Subgroup in Australia (Preiss, 2000), Puga and Bauxi subgroups in Brazil (Alvarenga and Trompette, 1988), which consist of three and more subformations or till members separated by non-glacial sediments.

The alternating tillites and members of lacustrine and fluvial sediments within the Glussk Formation suggest three events of glacier advance and retreat. These events may correspond to glacial and non-glacial ages, and their combinations characterize medium climatic fluctuations. Similar members and climatic events can be outlined, with a variable confidence degree, in other abovementioned Laplandian sections as well.

The most complete successions of the first and second Late Riphean glacial periods are usually composed also of alternating glacial and interglacial sequences, which suggest different climatic epochs. For instance, sediments of the last Late Riphean glaciation in southern Australia (Sturtian after Preiss, 1987) and in the western United States (Pocatello and Perry Canyon groups and their analogues; Link *et al.*, 1994) show indications of two large glacial events, which can be considered as glacial epochs consisting of shorter glacial episodes. The same is typical of two glacial subunits of the Grand Conglomerate in Katanga (Hambrey *et al.*, 1981), which corresponds probably to the first Late Riphean glacial period.

The Early Proterozoic and Late Archean. Signs of large climatic fluctuations are distinguishable also in the Early Proterozoic. Sections of the Canadian Glacial period in North America are up to 8 km thick in total, composed of two lower groups and of the lower part of the third group of the Lower Proterozoic Huronian Supergroup (Fig. 5b). Three glacial formations, the basal units in these groups, are separated by interglacial sequences, which are 1.5 to 3.5 km thick, consisting each of two formations. As was noted, the Canadian Glacial period lasted approximately 100 m.y. Therefore, subordinate glacial events, which resulted in accumulation of the abovementioned glacial formations, should logically be considered as glacial epochs. Each of glacial formations is of an intricate structure: par example, the Gowganda Formation encloses three glacial members from 80 to 150 m thick, which are composed of several different tillite beds separated by unconformities, shales, and sandstones (Fig. 5a). Events corresponding to these members can be identified with analogues of glacial ages and their alternation with interglacial sediments as the middle-rank climatic fluctuations. Sufficiently intricate is the section of the Canadian glacial period in South Africa, where boreholes recovered six glacial members of the Griquatown glacial unit, which are from 8 to 95 m thick and separated by interglacial members of sandstones, ferruginous carbonates, and limestones from 4 to 16 m thick (Visser, 1981). Similar intricate structure is characteristic of this unit in the Transvaal basin as well (Reitfontein Diamictite Member).

The alternation of several glacial and interglacial members is also observed in Upper Archean glacial sections of the South African Republic. For instance,

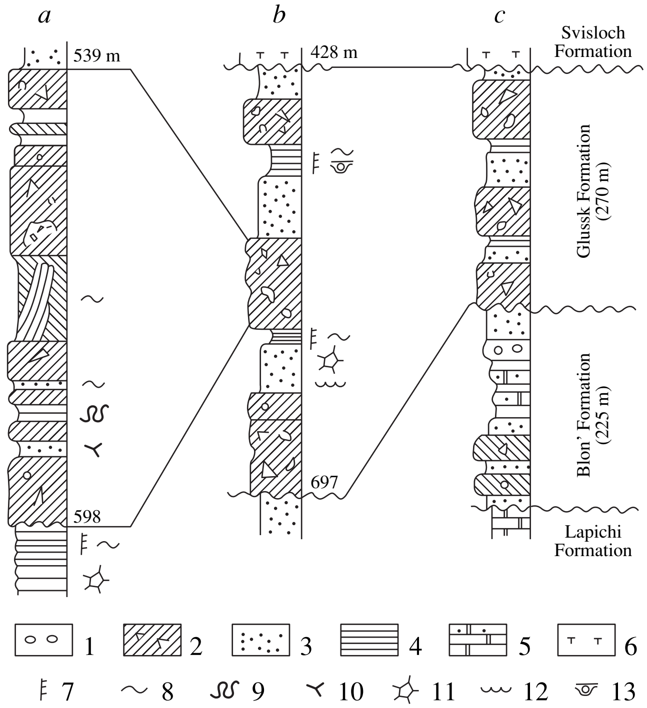

Fig. 4. Cyclic structure of the glacial Vil'chitsy Group, Lower Vendian, Belarus: (*a*) middle member of the Glussk Formation; (*b*) Glussk Formation; (*c*) Vil'chitsy Group; (1) conglomerate; (2) till; (3) sand; (4) clay and siltstone; (5) dolomite and sandy dolomite; (6) tuffite; (7) varved lamination; (8) glacial dislocation; (9) cryoturbation; (10) bioturbation (?); (11) desiccation cracks; (12) erosion signs; (13) dropstones.

the glacial part of the Mozaan Supergroup encloses four glacial beds from several to 20–30 m thick, which are separated by sandstone and shale members tens of meters thick (Young *et al.*, 1998).

Thus, it can be assumed, with a great confidence, that climatic fluctuations of highest ranks (superlong, long, and, probably, middle) were characteristic not only of the Phanerozoic, but of the Precambrian glacial periods as well. There are grounds to believe that there were also shorter climatic oscillations in the Precambrian.

3. GENERAL FEATURES OF CLIMATIC FLUCTUATIONS

The following peculiarities of climatic fluctuations can be distinguished within several large intervals of geological history.

Stability of climatic fluctuations hierarchy. The aforementioned data indicate that the multilevel hierarchy of climatic fluctuations maintained its stability during the entire Phanerozoic, i.e., over 500 m.y. Within glacial intervals of the Vendian and Late Riphean, recognizable are superlong, long, and middle fluctuations; short oscillations are also not inconceivable. Similar

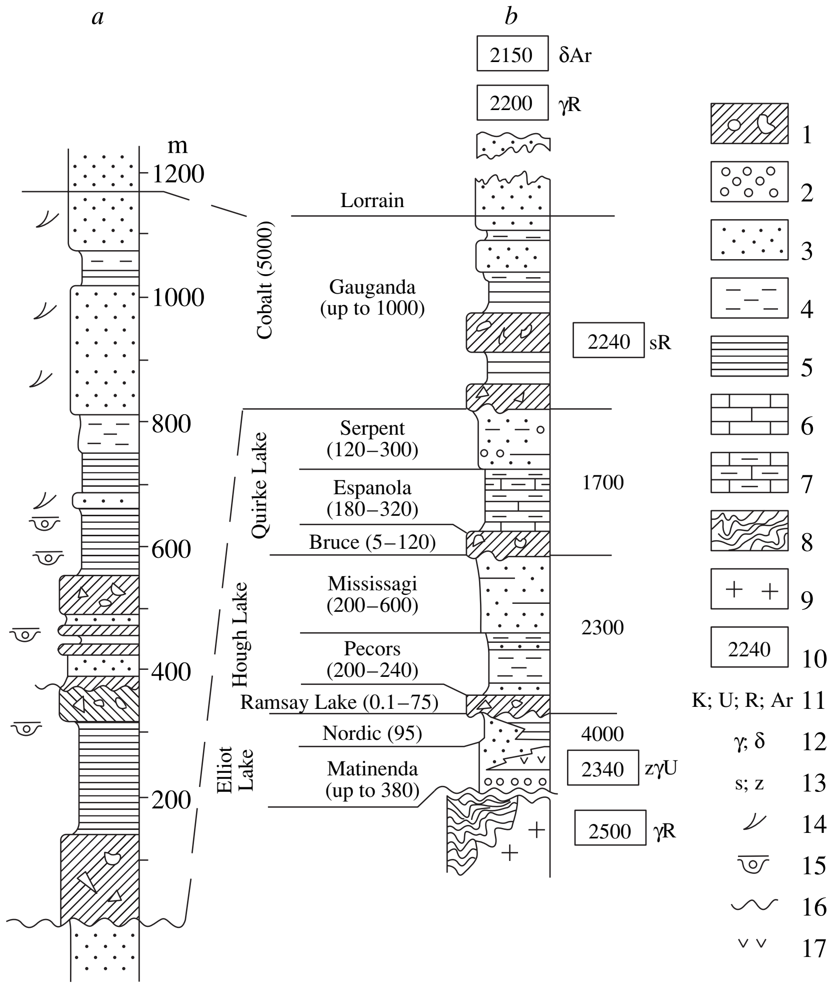

Fig. 5. Cyclic structure of the glacial Gauganda Formation (*a*) and Huron Supergroup (*b*), Lower Proterozoic, Canada (after Young, 1970): (1) tillite; (2) conglomerate; (3) sandstone; (4) siltstone; (5) argillite and shale; (6) limestone; (7) clayey limestone and marl; (8) crystalline rocks of the basement; (9) granite; (10) radioisotope age; (11) K–Ar (K), U–Pb (U), R–Sr (R), and Ar–Ar (Ar) radioisotope dating methods; (12) granite-based (γ) and diabase-based (δ) dating; (13) mica-based (s) and zircon-based (z) dating; (14) cross bedding; (15) dropstones; (16) erosional surface; (17) volcanics.

climatic fluctuations were characteristic as well of the Early Proterozoic and Late Archean glacial intervals, although is difficult to estimate their rank so far.

The existence of a stable multilevel hierarchy of climatic fluctuations throughout the Phanerozoic at least implies numerous and stable factors governing the above fluctuations. Accordingly, there is a principal opportunity to define individual oscillations in integral paleoclimatic curves and to interpret relevant processes using characteristic parameters of fluctuations (frequency, amplitude, configuration).

Amplitudes of climatic fluctuations. It is logical to consider the amplitudes of climatic fluctuations as a sum of superimposed elementary oscillations. For most small–scale climate changes, such an assumption is likely correct. A great deal in climatic fluctuations was however determined by feedbacks induced by climate changes in the biosphere. Negative feedbacks could weaken these changes and positive ones, to the contrary, strengthen them. Therefore, everything was complicated. For instance, short (Milankovitch-type) fluctuations in the Mesozoic resulted in relatively insignificant climate changes. To the contrary, similar fluctuations in the Pleistocene and Late Paleozoic could be accompanied by immense glaciations. This striking difference can be explained in the following way. Because of prolonged cooling periods related to superlong and long climatic fluctuations, the axis of short-period oscillations approached the temperature threshold, after which there were the onset and rapid formation of ice shields (Chumakov, 1995b). Thus, during short (Milankovitch-type) cooling periods, this threshold could be crossed that triggered formation of ice shields. The subsequent rapid growth of albedo and other processes connected by a strong positive feedback with glaciations (Fig. 6) stimulated further the temperature fall and substantial increase in short-period cooling amplitudes.

From the standpoint of above assumptions, the following working hypothesis can be proposed to explain different scales of climate changes. The low-amplitude fluctuations dominate during the non-glacial climate and represent a sum of changes produced elementary climatic oscillations at a given moment. In this case, the integral effect of changes is determined by cophasing of elementary climatic oscillations. When integral climate changes are significant, their final amplitudes can be strongly influenced via positive feedbacks by processes in biosphere. The formation of perennial glaciosphere was an extreme case of positive feedbacks. In such a case, amplitudes of climatic fluctuations increased stepwise.

Prevalence of non-glacial climate. Non-glacial (thermal) periods spanned approximately 70% of the Phanerozoic time, 78% of the last 1000 m.y. with periodic glaciations, and almost 90% of the preceding 2000 m.y. with episodic glaciations. The real role of non-glacial climate in geological history was even greater, because glacial episodes were discrete and alternated with interglacial episodes at all levels, from glacial periods to glaciations s.str. (or "glacial stages" in terminology of Quaternary geologists). Therefore, considering glacial and non-glacial episodes as equal in duration at first approximation, we can calculate that the integral duration of glacial epochs was approximately 50 to 60% of glacial periods. Only 50 to 60% of glacial epochs themselves were represented by glacial ages, which consisted, in turn, of glaciations s. str. and interglacial events. Judging from temperature curves compiled for the last 420 ka based on deuterium- and oxygen-isotope data for the Antarctic ice cores (Kotlyakov and Lorius, 2000), glacial conditions in high latitudes lasted approximately 85% of the second half of the Pleistocene glacial age. This value is likely the highest one characterizing the polar regions during the glacial maximum. Thus, it can be easily calculated that glaciations s. str. spanned not more than 30% of glacial periods and, consequently, not more than 10% of the Phanerozoic, 7% of the last 1000 m.y., and 3% of the preceding 2000 m.y. These values specified based on recent stratigraphic data define more exactly previous slightly higher estimates (Chumakov, 1995b).

The prevalence of non-glacial climate and subordinate role of glacial climate throughout the geological history suggest that global temperature on the Earth fluctuated mainly around positive values. In other words, the axial line of these fluctuations was within the interval of positive global temperatures. This can be illustrated by the following example, which allows simultaneously an approximate estimate of the global temperature on the glacial and ice-free Earth. We live in the glacial period. This is unambiguously indicated by the existence of significant polar ice caps and thick planetary psychrosphere. As is known, the average present-day temperature of the Earth surface is close to +15°C. During the last glacial maximum, this parameter was by several degrees lower and during the Late

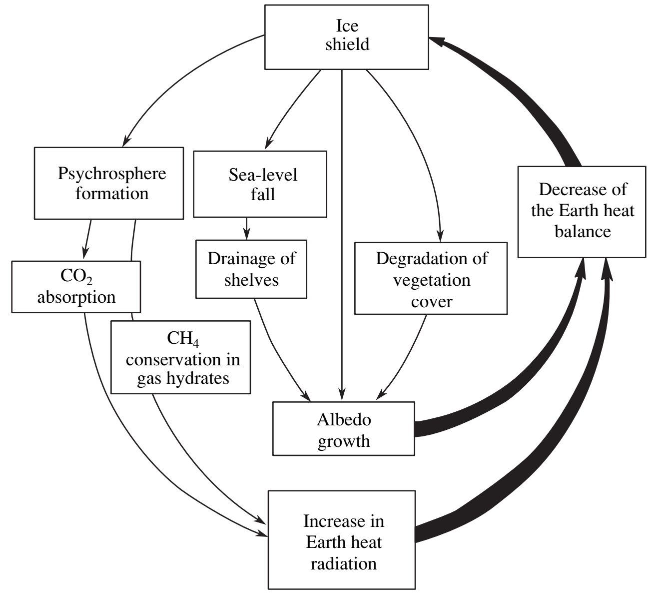

Fig. 6. Positive feedbacks in the ice shield–biosphere system.

Cretaceous non-glacial epoch, by several degrees higher as compared with the present-day value, i.e. in both cases it was well within the interval of positive temperatures. Mathematical modeling shows that the greatest part of the World Ocean was ice-free during the Late Precambrian glaciations and consequently, the average temperature of the Earth surface remained above zero, even though the land glaciers could form in low latitudes (Poulsen *et al.*, 2002; Poulsen, 2003).

The temperatures, around of which climatic fluctuations were scattered, changed with time. The main factors, which could change them, were planetary albedo, composition of the atmosphere, and orbital parameters. It is logical to assume that the temperature threshold, after which the perennial glaciosphere formed and non-glacial climate gave way to the glacial one, was not constant as well. It depended, to a certain extent, on paleogeographic situation in high and middle latitudes, as well as on circulation systems in the hydrosphere and atmosphere.

Global, synchronous, and cophasal climatic fluctuations. The approximate synchronism of glacial periods on different continents and, hence, their global scale is undoubted, confirmed by paleontological and radiometric dating. It is however more difficult to correlate climatic fluctuations of lower ranks in remote regions. This is possible only in case of stratigraphically well-studied intervals of the Phanerozoic. The correlation within these intervals shows that global and regional events are distinguishable among shorter climatic fluctuations.

Global fluctuations were synchronous and cophasal. This is distinctly exemplified by modern glaciers (Zakharov *et al.*, 1999; Solomina, 1999), by short and ultrashort fluctuations during the historical period (Broecker and Denton, 1989), Holocene, and Pleistocene (Broecker and Denton, 1989; Thompson *et al.*, 1997; and others), by long-period fluctuations, and by general climatic trends during the Cenozoic (Fot'janova and Serova, 1994; Dingle and Lavelle, 1998), Cretaceous (Krasheninnnikov *et al.*, 1990; Herman, 1993; Frakes, 1999), and Permian (Chumakov, 1995a). It is noteworthy that cophasing of these fluctuations is traceable in various latitudes (low ones included) of both hemispheres. For instance, the medieval Little Glacial "Period" and preceding warm Vikings "Period" were manifested as respective cooling and warming near the North Polar Circle and in the Sargasso Sea (Keigwin, 1996). The Holocene ultrashort climatic fluctuations are registered in polar regions and Equatorial Africa (Stager and Mayewski, 1997). The last glacial maximum of the Pleistocene is recorded as cooling in low latitudes. In equatorial South America, this event is inferred from lowering of the snow line and feeding areas of mountainous glaciers by 700–1200 m (Broecker and Denton, 1989) and from oxygen-isotope characteristics of their ice. The cooling is evident from geochemistry of noble gases in underground water of equatorial Brazil and lowered snow lines and vegetative belts in Equatorial Africa, Sumatra, and New Guinea (Ninglan *et al.*, 1999). The penultimate interglacial and last glacial episodes are distinguishable in high and middle latitudes, and corresponding warming and cooling are established in Java (van der Kaars and Dam, 1995). During the Late Cretaceous and Cenozoic, the general cooling trend and long-period climatic fluctuations were similar, although differing in amplitude, in high and middle latitudes of Northeastern Asia (Krasheninnikov *et al.*, 1990; Herman, 1993; Velichko *et al.*, 1994; Fot'janova and Serova, 1994; Zakharov *et al.*, 1999), in surface and deep oceanic waters of low latitudes (Douglas and Woodruff, 1981), in middle latitudes of the southern Indian Ocean (Clarke and Jenkins, 1999), and in Antarctica (Dingle and Lavelle, 1998). Almost everywhere in these regions, warming is noted for the Cenomanian, Santonian, Campanian, early–late Eocene, and middle Miocene intervals, whereas cooling is registered in the Maastrichtian, terminal Eocene–Oligocene, and in the second half of the Miocene to Pliocene (Fig. 7).

The mentioned data imply that many climatic fluctuations (ultrashort, short, long, and superlong) were of global scale, synchronous and cophasal. This leads to the important conclusion that global climatic fluctuations represented a response to changes in the heat balance of the Earth (Chumakov, 1995a), rather than a result of changes in the mechanism of heat redistribution in the biosphere as its is commonly assumed (*CLIMAP…*, 1976; Nikolaev, 2000; and others). It should be emphasized that a thorough study of global fluctuations is very important for paleoclimatic interpretations. Their identification narrows the possible sphere of factors responsible for global climate changes.

Regional climatic fluctuations. Fluctuations of such kind are observable now and, undoubtedly, were characteristic of many regions in the past. As is known from meteorology, geography, and historical geology, they are induced by macrogeographic factors (changes in position, size, and configuration of continents, seas, and oceans, in the topography and landscapes), which redistribute heat in the biosphere. Past regional climatic fluctuations can be exemplified by the Turonian–Coniacian warming in Alaska that occurred in response to opening and widening of the fore-Cordillera strait ("Western Seaway") and by a relative cooling in Northeastern Asia caused by closure of a seaway between Asia and Alaska (Spicer and Herman, 1998). The other examples of regional climatic changes related to seaways opening and closing are climatic fluctuations in West Siberia and the Arctic region during the Paleogene and Neogene (Akhmetiev, 1996).

4. TRANSFORMATION OF CLIMATIC ZONALITY

4.1. Global, latitudinal, and regional climates. Global climatic fluctuations resulted in transformations of latitudinal zonality, which were different in scale and duration. Sufficiently well studied are changes in climatic zonality of the Quaternary (Zubakov, 1986; Barash *et al.*, 1989; Velichko and Nechaev, 1999; and others), Cretaceous (Ronov and Balukhovskiĭ, 1981; Krassilov, 1985; Naidin *et al.*, 1986; Vakhrameev, 1988; Chumakov *et al.*, 1995), Permian, and initial Triassic periods (Ziegler *et al.*, 1997; Zharkov and Chumakov, 2001a; Chumakov and Zharkov, 2002, 2003), which are good examples of zonality transformations. The transformations were mostly rather gradual, of a relatively small-scale. They changed somewhat the width, latitudinal position and climatic parameters of some belts. On the other hand, particular transformations were rapid, qualitatively significant, and they can be termed as reorganizations of the climatic belts systems. Under their influence, new belts appeared, some old belts disappeared or became reduced, and main belts changed their latitudinal position and width. These parametric changes were of prime significance for climate, because the latitudinal position of belts determines not only the intensity of solar radiation and average temperatures, but also the seasonality (thermal and light), baric parameters, wind direction, precipitation, evaporation, cyclone trajectories, and circulation (Chumakov, 1995b). The reorganizations gave rise to new types of climatic zonality. Therefore, in addition to different climates in latitudinal climatic belts (*latitudinal climates*), one should distinguish the *global climate*, i.e., the type of planetary climatic zonality, first of all, by a set, width, latitudinal position, and contrasting parameters of climatic belts.

Latitudinal climatic belts were influenced by macrogeographic factors such as dimensions and position of continents, mountainous areas, seas, and seaways. These factors substantially modified latitudinal climate. Deforming and complicating climatic belts, they transformed latitudinal belts into sublatitudinal ones simultaneously creating a system of *regional, sectorial* (marine and continental to a variable extent), and *altitude* climate varieties within the belts.

4.2. Glacial and non-glacial global climate. Beginning from the Late Archean, two main types of the global climate alternated in geological history: glacial and non-glacial. Despite the prevalence of the non-glacial climate throughout the geological history (see section 3), numerous short glacial episodes were characteristic of long historical periods, particularly of the Riphean–Phanerozoic, and, therefore, two global climates repeatedly changed each other. This was caused mainly by changes in the heat balance of the Earth and led to formation and degradation of perennial glaciosphere (ice shields, perennial sea ice and permafrost) and psychrosphere. The alternating climates changed albedo and concentration of carbon dioxide in the atmosphere, sea-level and oceanic temperature fluctuations, proportion of land and sea areas, integral moisture transfer and latitudinal temperature gradient, circulation system in the atmosphere and hydrosphere, distribution of atmospheric precipitation and vegetative cover. Positive feedbacks intensified these processes (Fig. 6) and led to additional changes in the average Earth surface temperature and in biosphere as a whole, and eventually to principal transformations of climatic zonality (Chumakov, 1995b).

Glacial climate. The present-day climatic zonality of the Earth formed in the second half of the Cenozoic. This zonality and its varieties are sufficiently well known owing to recent geographic, biogeographic, and oceanographic studies, and to the thorough research of Late Cenozoic and Quaternary geology. Although the climatic zonality during the last glacial maximums and interglacials is sufficiently well studied, its development at the beginning of glacial periods is poorly known yet. An important comparative material is inferable from evolution of the glacial climatic zonality during the Permian (Ziegler *et al.*, 1997, 1998; Zharkov and Chumakov, 2001a; Chumakov and Zharkov, 2002, 2003) and in the Late Ordovician (Barnes and Williams, 1991; Frakes *et al.*, 1992; Crowell, 1999). Data on Permian glaciations are of particular interest. In fact, only these data provide an insight into processes in the biosphere during the large-scale warming episodes on the Earth and by transition from the glacial to non-glacial climate.

The glacial climatic zonality is characterized by existence of ice shields and glacial belts. During glacial maximums ("great glaciations") of the Phanerozoic, glacial belts of high to middle latitudes advanced sometimes to the latitude of 30° as, for instance, in South

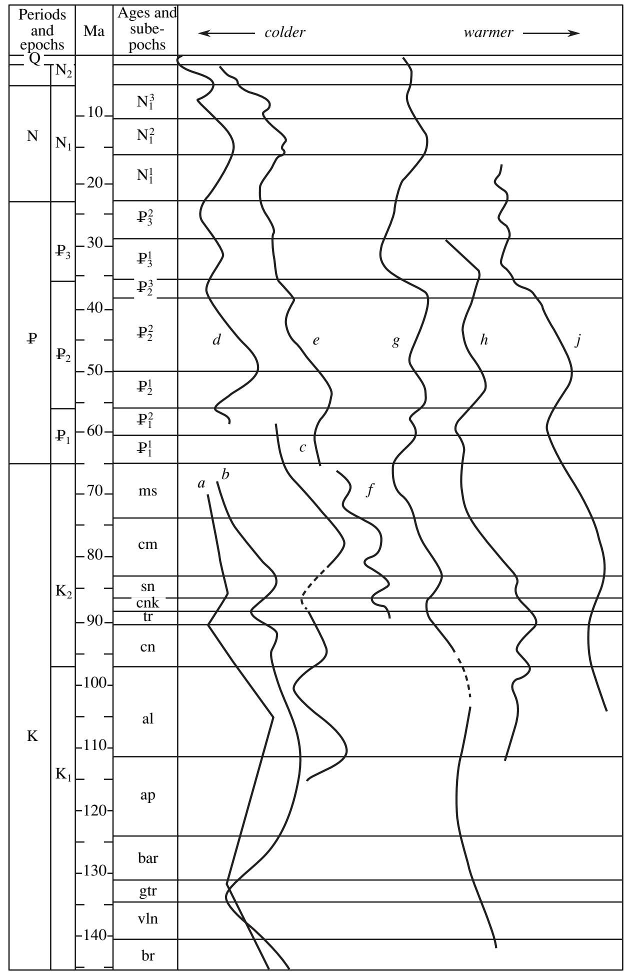

Fig. 7. Correlation of Cretaceous and Cenozoic climatic fluctuations in different latitudes of the Northern and Southern hemispheres: (a) conditional mean annual temperatures in Siberia inferred from oxygen-isotope and Ca/Mg data (Gol'bert, 1987); (b) relative temperature changes in northern Siberia inferred from lithological and paleontological data (Velichko *et al.*, 1994); (c) relative temperature changes in mountainous areas of northeastern Asia inferred from paleobotanic data (Krasheninnikov *et al.*, 1994); (d) relative variations of January temperatures in northeastern Siberia (Velichko *et al.*, 1994); (e) relative temperature changes in the Koryak Highland and Kamchatka inferred from paleobotanic and foraminiferal data (Fot'janova and Serova, 1994); (f) temperatures in the Northwest Pacific inferred from oxygen–isotope and paleobotanic data (Zakharov *et al.*, 1999); (g) surface water temperatures in low latitudes inferred from oxygen-isotope data (Douglas and Woodruff, 1981); (h) surface water temperatures in middle latitudes of the Indian Ocean inferred from oxygen-isotope data (Clarke and Jenkins, 1999); (j) relative temperature changes in the Antarctic Peninsula inferred from geochemical and other data (Dingle and Lavelle, 1998). Ages: (br) Berriasian, (vln) Valanginian, (gtr) Hauterivian, (bar) Barremian, (ap) Atpian, (al) Albian, (cn) Cenomanian, (tr) Turonian, (cnk) Coniacian, (sn) Santonian, (cm) Campanian, (ms) Maastrichtian.

America during the Asselian–early Sakmarian time of the Early Permian (Ziegler *et al.*, 1997; Chumakov and Zharkov, 2002, 2003; and Fig. 8a) or in the Late Ordovician. Temperate humid belts became reduced in these periods and even completely degraded sometimes, giving place to periglacial zones with cold steppe landscapes (tundra–steppes, after Philips, 1986; "periglacial steppes, after Velichko and Nechaev, 1999). In lower latitudes, the cold steppes gave way to the belt of semideserts and deserts with temperate to tropical climate and, sometimes, to cold semi-deserts as, for example, in South America during the Early Permian (Fig. 8a). As is shown in available reconstructions for the last Pleistocene glacial maximum, a strong reduction and transformation involved also subtropical belts, which were almost completely replaced by steppe, semi-desert, and desert landscapes with a temperate climate. Although climatic belts of low latitudes escaped such a drastic transformation, zones of deserts and semi-deserts substantially widened there at the expense of subequatorial belts (savanna), whereas savannas replaced partly the former humid equatorial belts (Phillips, 1986). Widening of arid and semiarid belts at the expense of neighboring belts is also established for the Early Permian glacial maximum. The global climate type under consideration can be termed as the *climate of glacial maximums* or *climate of great glaciations.*

The present-day climatic zonality allows us to imagine the past global climate of interglacial epochs. Data on the Early Permian climate confirm such a possibility. In addition, they suggest that initial and terminal stages of glacial periods were also characterized by the climatic zonality similar to the present-day one. Retreated from middle latitudes during the Late Sakmarian time and contracted to the size of the present-day polar cap, the tremendous Gondwanan ice shield still existed, although it was oscillating, in the Artinskian Age and probably later (Zharkov and Chumakov, 2001b; Chumakov and Zharkov, 2002). The Permian polar cap resembled the present-day one being similarly surrounded by periglacial zones and temperate cold forests, which were comparable with taiga in opinion of some experts in paleobotany. The glacial belt reduction was accompanied by reviving of the temperate cold humid belt in the Southern Hemisphere and by a slight widening of similar belt in the Northern Hemisphere. An insignificant reduction was characteristic of the northern arid belt, while its southern counterpart widened. Thus, the Permian epoch was characterized by a substantial reorganization of the climatic zonality as compared with that of the glacial maximum. The global glacial climate similar to the present-day or Late Sakmarian–Early Artinskian climate can be termed as the *climate of polar caps.* It was likely characteristic of the other Pleistocene interglacials and, probably, of the Late Eocene, Oligocene, and Neogene minimums, when the Antarctic glacial cap existed.

Paleoclimatic reconstructions for the Kazanian–Tatarian ages of the Late Permian illustrate the climatic zonality during the terminal phase of the Carboniferous–Permian glacial period. The southern polar cap degraded by that time, although some active glaciation centers that produced icebergs were preserved in Antarctica and southeastern Australia. In northeastern Asia, glaciation centers appeared again indicating continuation of the glacial period on the Earth. Nevertheless, the former climatic zonality changed, as a continuous glacial belt disappeared and polar areas became moderately cold in general. They were rimmed by belts with a temperate humid climate. The climatic asymmetry of hemispheres substantially decreased and synonymous climatic belts were more symmetrical relative to the equator (Zharkov and Chumakov, 2001a; Chumakov and Zharkov, 2002, 2003). This variety of the glacial climate can be termed as the *climate of cold polar areas.* The next climatic reorganization induced by a sharp global warming near the Permian–Triassic boundary resulted in replacement of the glacial climate by the non-glacial one.

Non-glacial climate. In distinction from the glacial climate, this type of the global climate is distinguishable based on the past geological examples, i.e. on paleoclimatic reconstructions only, because of impossibility to apply actualistic climatic models. The reconstructions indicate that the non-glacial global climate was also non-uniform. At present, it can be subdivided in two, the *non-glacial humid* and *non-glacial arid* varieties (Zharkov and Chumakov, 2001a).

The typical non-glacial humid climate prevailed in the Late Cretaceous beginning from the terminal Albian. The climatic zonality peculiar of this climate is considered in several works (Ronov and Balukhovskiĭ, 1981; Vakhrameev, 1988; Krassilov, 1985; Chumakov *et al.*, 1995; Zharkov and Chumakov, 2001a; Valdes *et al.*, 1999; Barrera and Johnson, 1999). Polar areas in both hemispheres were occupied at that time by temperate and temperate warm belts with positive annual temperatures (Ditchfield *et al.*, 1994; Spicer *et al.*, 1996, Kennedy, 1996) and very rare frosts (Falcon-Long *et al.*, 2001). Widespread in these belts were deciduous broad-leaved and coniferous forests, where dinosaurs herded, and lakes populated by crocodiles (Vakhrameev, 1988; Chumakov *et al.*, 1995; Clemens and Nelms, 1993; Tarduno *et al.*, 1998; Falcon-Long *et al.*, 2001). The warm humid belts were substantially wide, occupying the middle (Fig. 8c) and partly the high latitudes of both hemispheres (Cenomanian and Turonian ages). Judging from occurrence of the monoxylic wood remains, the climate in these belts was mostly frost-free. The arid belts located in tropical and, partly, middle latitudes were moderately developed. At the beginning of the Late Cretaceous, they occupied approximately 35% of land, and their area gradually reduced to 25% by the end of this epoch, (Zharkov and Chumakov, 2001a; Chumakov and Zharkov, 2003; Fig. 8b) giving way to the widening humid equatorial belt. In total, humid belts covered 75% of land at the end of the Cretaceous.

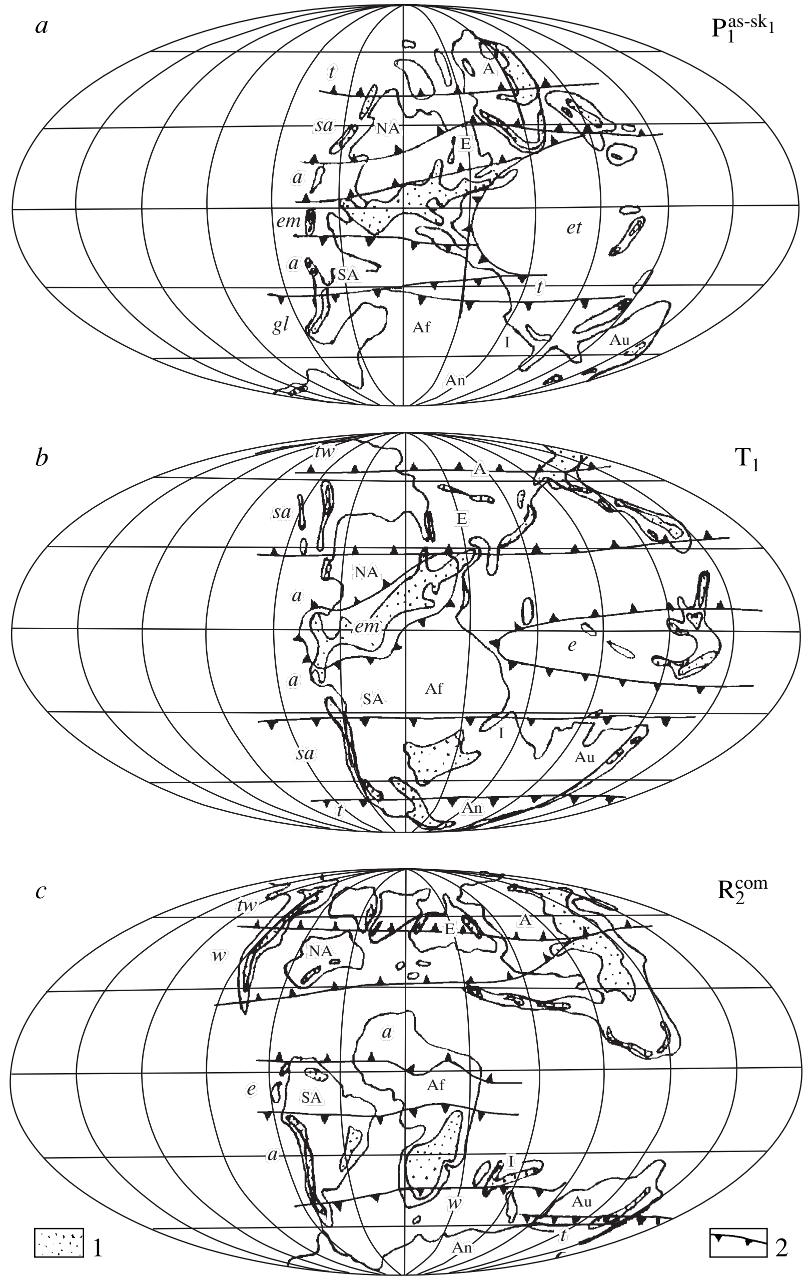

Fig. 8. Climatic zonality (a) during the Asselian–early Sakmarian time (after Chumakov and Zharkov, 2002), (b) in the Induan Age of the Early Triassic (after Chumakov and Zharkov, 2003), (c) in the Maastrichtian Age of the Late Cretaceous (after Zharkov and Chumakov, 2001a): (1) mountainous belts and regions; (2) boundaries of climatic belts (after Zharkov and Chumakov, 2001a; Chumakov and Zharkov, 2002, 2003). Continents: (A) Asia, (Au) Australia, (An) Antarctica, (Af) Africa, (E) Europe, (I) India, (NA) North America, (SA) South America. Climatic belts: (*gl*) glacial, (*t*) temperate, (*tw*) temperate warm, (*w*) warm, (*sa*) semiarid, (*a*) arid, (*et*) equatorial–tropical, (*e*) equatorial, (*em*) equatorial mountainous.

The Induan Age of the Triassic was a time of typical non-glacial arid climate (Zharkov and Chumakov, 2001a, 2001b; Chumakov and Zharkov, 2003; Fig. 8b). The humid temperate belts of northern and southern high latitudes and the equatorial humid belt (mountainous areas with a vertical climatic zonality) occupied, in total, 20% of land at that time. The remainder was covered by arid and semiarid belts covering almost all the middle and partly high latitudes (Fig. 8b). Semiarid belts were particularly widespread. They were about 40° wide in both hemispheres. The narrow equatorial belt was discontinuous and, therefore, the northern and southern arid belts joined locally each other. The typical equatorial humid climate was preserved at that time only within the Cimmerian arc that represented an archipelago of microcontinents between the Paleo- and Neo-Tethys.

The genesis of the Induan Age climate represents a certain problem. An increase of the global surface temperature should be theoretically favorable for elevated evaporation in oceans and seas and, correspondingly, for a higher humidity of the atmosphere (Barron *et al.*, 1989). This is likely correct for the Late Cretaceous epoch. Nevertheless, the situation was more ambiguous, because the moisture distribution over land was largely controlled by macrogeographic peculiarities of the Earth.[4] A combination of these factors determined the remoteness and latitudinal position of moisture sources, paths of moisture transfer in the atmosphere, and distribution of baric and orographic barriers along them. Therefore, the breakup of supercontinents and opening of new oceans coupled with transgressions and wide distribution of epicontinental seas and seaways favored development of the non-glacial humid climate in the Late Cretaceous, while a combination of mountainous supercontinents with regressions in marginal orogenic belts resulted in prevalence of non-glacial arid climate at the beginning of the Triassic (Zharkov and Chumakov, 2001a; Chumakov and Zharkov, 2002, 2003).

4.3. Asymmetry of climatic zonality. Typical of the present-day climatic zonality of the Earth is a slight asymmetric position of climatic belts relative to the equator. Similar asymmetry is likely a usual phenomenon characteristic of planets with the atmosphere. It is characteristic of Mars, par example, the polar caps of which are substantially different in size. The climatic asymmetry of the Earth is thought to be a consequence of asymmetrical position of continents, oceans, seas, and, correspondingly, of circulation systems in the Northern and Southern hemispheres. Moreover, these systems are autonomous to a certain extent. Inasmuch as land, oceans, and seas were hardly located symmetrically relative to the equator at any time before, one can assume that the climatic asymmetry of hemispheres existed throughout the geological history. Paleoclimatic reconstructions confirm this assumption. An insignificant climatic asymmetry existed in various forms during the entire Cretaceous period. At that time, the warm humid belt in the Southern Hemisphere (dissimilar to the Northern one) reached sometimes (Neocomian, Cenomanian, Turonian) the high latitudes (Vakhrameev, 1988; Chumakov *et al.*, 1995). The arid belt of the Southern Hemisphere was slightly wider in the Early Cretaceous, Cenomanian, and Turonian, whereas the equatorial humid belt advanced noticeably southward in the Maastrichtian. It seems that, dissimilar to its present-day state, the Southern Hemisphere was slightly warmer in the Cretaceous Period than the Northern Hemisphere. At the beginning of the Triassic, climatic asymmetry was weak, like in the Cretaceous and nowadays, but the Southern Hemisphere was slightly colder (Chumakov and Zharkov, 2002, 2003).

The present-day zonality and examples mentioned above imply that climatic asymmetry in periods of the interglacial and, particularly, non-glacial climate was usually insignificant even in case of highly asymmetrical position of land and oceans as it was characteristic, for instance, of the Early Triassic.

In periods of the glacial climate, particularly during "great glaciations," asymmetry in position of climatic belts substantially increased and was multiply manifested in its extreme form of the "monopolar glaciations". Although immediate causes of climatic asymmetry on the Earth might differ, asymmetrical position of continents, oceans, large orogenic belts, and oceanic currents were the main factors responsible for this phenomenon. In the Pleistocene, ice sheets advanced to 38°N in the Northern Hemisphere and only to 60°S in the Southern Hemisphere, because they were kept back by oceans surrounding Antarctica. The Late Ordovician glaciations developed only the Southern Hemisphere, because there was no land in high and middle latitudes of the Northern Hemisphere. The most remarkable example of extreme climatic asymmetry and its dependence on glaciation was a climatic zonality of the Asselian–Early Sakmarian time (initial Early Permian) (Chumakov and Zharkov, 2002). As was mentioned, the glacial belt occupied almost entire high and middle latitudes of the Southern Hemisphere at that time. The southern temperate humid belt was reduced to a narrow (10°) westward pinching-out band, and the semiarid belt disappeared (Fig. 8a). To the contrary, the Northern Hemisphere lacked the glacial belt. The temperate belt of this hemisphere occupied the high and nearly all the middle latitudes, while semiarid belt was as wide as 20°. Despite a similar disposition of land, seas, and oceans at the beginning of the Triassic, when the glacial climate on the Earth was replaced by the non-glacial one, climatic asymmetry immediately became minimal (Fig. 8b).

CONCLUSION

The discussed trends in climate changes are mainly inferred from geological data being consequently empirical. At the scale of an eon, they outline irreversible changes from non-glacial climate of the Early Archean to quasi-periodic glacial climate of the Late Riphean, Vendian and Phanerozoic. The stable hierarchy of global, synchronous, and cophasal climatic fluctuations of 10–12 ranks is distinguishable within the general gradual climatic trend. The fluctuations determined gradual transformations or rapid reorganizations of the climatic zonality on the Earth depending on the scale of fluctuations and induced feedbacks.

This implies that evolution of the Earth climate was relatively regular, not chaotic. The inferred regular climatic changes seem unexpected for such an intricate

[4] In addition to main planetary trends in moisture distribution, which are controlled by the form and rotation of the Earth and by circulation cells responsible for creation of primary latitudinal belts, the arid and humid ones inclusive.

open system as the climatic system of the Earth. In this connection, two assumptions are logical. First, the Earth climate system was quasistationary and this phenomenon was most clearly manifested during the large time intervals (eons, erathems, and periods) and by highest-rank fluctuations. Ultrashort and weather oscillations in particular were more chaotic, and this is not surprising. The non-linear development of natural processes, the geological ones included, usually changes depending on the analysis time scale (Grachev, 1998). Second, among all the diverse nonlinear processes that influenced the Earth climate there was likely a limited number of most influential "governing" processes with fixed or periodical attractors. These processes significantly differed from each other in the frequency of characteristic variations. The following "governing" processes can be outlined at present.

(1) Variations in the endogenic activity of the Earth were responsible for superlong (150 m.y.) and long (several tens of million years) climatic fluctuations. Contrary to the widely accepted standpoint, periods of superlong fluctuations substantially differ from duration of the galactic year. They were probably related to the self-oscillating geodynamic processes in the Earth interiors (Dubrovskiĭ, 1998; Dobretsov, 1999), although the influence of astronomical factors cannot completely be ruled out.

(2) The other factor is the carbonate and organic carbon burial in the sedimentary shell, lithosphere, and mantle of the Earth, the intensity of which depends, in particular, on geodynamics, latitudinal position of continents, weathering processes on land, and accumulation rates of organic carbon in seas.

(3) Variations in orbital parameters and inclination of the Earth axis are responsible for the Milankovitch-type short fluctuations. Their astronomical origin is inferred in many works and, what is important, this is supported by the fact that such fluctuations occurred throughout the entire Phanerozoic regardless of repeated changes in geological, geographic, and biotic situations on the Earth.

(4) Variations in the solar activity control the ultrashort climatic fluctuations.

There are probably the other "governing" processes, for example, those, which caused middle climatic fluctuations.

The other accompanying processes only modified, to some extent, the superlong and long climatic fluctuations determined by "governing" factors. They strengthened or weakened influence of the latter, being unable to control the alternation of glacial and non-glacial periods.

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