ATLAS OF PALEOCLIMATES AND PALEOENVIRONMENTS OF THE NORTHERN HEMISPHERE

Late Pleistocene – Holocene

Edited by

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PREFACE

I. The Quaternary period (Pleistocene and Holocene) comprises the past 2,000,000 years of the geological history of the Earth and the evolution of life during that time. In comparison with earlier geological ages, its particular significance is associated with the emergence of the early man and his distribution on the Earth in various geographical environments.

Climatic changes, showing a cyclic recurrence, have greatly influenced man's evolution and the development of human culture. They have also been responsible for the repeated changes which occurred in the geographical environment in general and for the transformation and spatial shifts of landscape zones in particular. Man has been playing an ever increasing role in shaping the environment, which is why Quaternary studies has been calling for cooperation between the representatives of earth sciences, experts in the physical, chemical and biological sciences, and specialists in anthropology and archeology. More than 60 years ago the International Union for Quaternary Research (INQUA) was established by scholars engaged in solution of scientific problems of the Quaternary to encourage joint research, and the discussion and publication of results of their investigations.

For a better evaluation of the climatic cycles several commissions of INQUA have subsequently become specialized in studying the reasons and frequency of the climatic and environmental changes during the Quaternary period. To elaborate and publish the wealth of knowledge acquired on the spatial and temporal paleogeographical changes on the Northern Hemisphere during the last climatic macrocycle the Commission on Paleogeographic Atlas of the Quaternary was established 14 years ago under the presidency of A. A. VELICHKO. Since the late 1970s the activities of this commission and those of the Laboratory of Paleogeography, Institute of Geography, Academy of Sciences USSR have also been supported by the INQUA Commission on Loess. At that time the conveners of this Commission, M. PÉCSI and B. FRENZEL put a strong emphasis on reconstructions of the climatic and environmental changes during the Quaternary based on investigations of loess-paleosol sequences. Further on, the Research Project Group "Terrestrial Paleoclimatology" of the Ministry of Research and Technology, Federal Republic of Germany, headed by B. FRENZEL, became involved in the compilation of the atlas. The final drawing of the maps, the lithographic works, the printing of the map sheets and of the explanatory text were accomplished by the Geographical Research Institute, Hungarian Academy of Sciences and its Laboratory of Cartography, under the guidance of M. PÉCSI and Z. KERESZTESI.

The compilation, design and publication of the atlas initiated by the above mentioned Commissions have been carried out over the past 14 years with the participation of several institutions and of a great number of experts. This work has involved a great deal of intellectual and manual effort, considerable expenses and significant contributions by international and national scientific institutions. Moreover, the importance of the atlas has recently been enhanced by the "Global Climatic Change" programme launched by the International Geosphere-Biosphere Programme and forms a basis for the seventh core project of this programme entitled PAGES.

II. This atlas contains several climatic reconstructions compiled for the same or similar time intervals. Since these reconstructions have been made by the application of different approaches and methods; consequently, they are not always in full agreement, and offer somewhat varied climatic characteristics. We included such reconstructions in the atlas for the purpose of introducing state-of-the-art Quaternary studies as well as to demonstrate the possibilities offered by the various methods.

Paleogeographic computations and estimations on the basis of factual data show that there were phases during the geological past when mean global annual temperatures were approximately 1°C higher (about 6,000 years B.P., the optimum of Holocene) and ca. 2 °C higher [about 120,000 years B.P., the optimum of Last Interglacial (Eem, Mikulino, Sangamon, etc.)] than today.

In the atlas reconstructions of climatic conditions and landscape elements allow judgements about the hydrothermal regime of the Northern Hemisphere and provide an assessment of the environmental response including that of vegetation, permafrost, hydrological elements, geomorphic processes, etc. Moreover, in the atlas sections will be found which contain map reconstructions showing the changes which occurred during the warm stage of the Last Glaciation between 35,000 and 25,000 years B.P. and during the maximum cooling of the Last Glacial epoch (Würm, Vistula, Valdai, Wisconsin, etc.) between 20,000 and 18,000 years B.P. These reconstructions enable the reader to get acquainted with changes in climate and landscapes that result from an abrupt cooling and enable landscape stability and susceptibility to be assessed. The fourth time span chosen for the paleogeographic reconstruction is the Holocene climatic optimum.

Taken a whole, this series of paleoenvironmental maps comprises the last climatic macrocycle of the past 130,000 years and exemplify fluctuations in the climatic and landscape systems. A cyclic mechanism of this kind is considered to be typical of the Earth's biosphere for the last 2,000,000 years. Detailed analysis of this macrocycle is considered to be a relevant contribution to the understanding of natural evolution on the Earth.

Climatic research revealed that the use of fossil energy sources and modern technological systems has given rise to a permanent change in atmospheric gas composition (increases in the content of carbon dioxide, methane, freons) and this in turn means that the global annual temperature is expected to rise of about 1°C by 2010 and of 2 °C after 2030. Paleogeographic reconstructions should therefore be part of any exercise in environmental forecasting, aimed at the formulation and application of suitable agricultural and industrial production strategies.

III. The Editorial Board would like to express its gratitude to the sponsors of the atlas: the INQUA Executive Committee, for providing a special grant; the Hungarian Academy of Sciences; the Academy of Sciences and Literature, Mainz, Germany; the Federal Ministry of Research and Technology, Germany, the Institute of Geography, Academy of Sciences USSR; the Institute of Botany, University of Hohenheim, Germany; the Geographical Research Institute, Hungarian Academy of Sciences, who undertook a large part of the work and provided substantial financial support. Thanks are also due to the Enkidu Foundation (Berne-Tübingen) who offered a loan under favourable conditions.

The Editorial Board feels also indebted to members of the two INQUA commissions, to the experts of the national working groups, to the consultants, revisers of the map sheets and explanations, as well as to translators and members of the technical staff for their valuable work.

The Editors

The Globe's ecological situation poses a serious threat to mankind. Some of the difficulties anticipated concern changes in climate, since according to measurements on air bubbles in the inland ice of Greenland and of Antarctica, and on instrumental data obtained for the last decades, the concentrations of the greenhouse gases CO_2 and CH_4, and recently also of the freons, are steadily increasing. This would mean a rise in the global annual mean air temperature of at least ca. 1°C by the year 2010 AD, and of about 2 °C or more after 2030 AD. Thus, agriculture, forestry, economics, and politics need to develop new strategies and technologies all over the world to counteract these difficulties.

In order **to take measures in due time**, one should be aware of the possible temporal and spatial changes in the future, if the predictions mentioned turn out to be correct. This can either be done by **climatic modelling** or by **analysing similar situations of the past**. In fact, for the last 130,000 years, the Earth experienced at least twice periods of pronounced warmer climates, i.e., during the **Last Interglacial** and during the **Holocene climatic optimum**. In our reconstructions they will be dealt with in some detail, together with two phases of cold climates between the warm phases; the **interstadial complex** immediately preceding the last major advance of the inland ice and the **pleniglacial phase** itself.

By comparing these four episodes in the Earth's history of climate with each other, the general trends in the natural evolution of the climate will be understood much better than was possible heretofore, including consequences for the biosphere, as well as for the cryosphere and hydrosphere, respectively. Thus the amplitude and the timing of natural changes will become apparent.

Due to the number of paleoecological observations available, their geographical distribution pattern and accuracy, this atlas focusses on the Northern Hemisphere, only.

Two major problems must be dealt with: firstly the **correlation of events** which occurred in various regions of the Northern Hemisphere at the same time, even if it is at present impossible to date them reliably by physical methods; secondly, the fact that **various methods used to reconstruct the ecological conditions** of the past in terms of climate history may result in somewhat diverging values.

Dealing with the **Last Interglacial** we focussed on those phases which can directly or indirectly be shown by biostratigraphical methods to be synchronous. This holds true for the **Eem**— in a paleobotanical sense —, which can be correlated accurately with the **Mikulino Interglacial** of Eastern Europe. Both, the Eem and Mikulino can be correlated reliably with marine sequences of the North Sea, Baltic Sea, and Norwegian Sea. Marine sediments are characterized by their floras and by their marine faunas. So the marine faunas enable the correlation with the **Kazantsevo Interglacial** of Western Siberia and of the northwestern part of Central Siberia. On the other hand, some episodes of the Eemian pollen sequence of Western Europe are recorded in the **Oxygen Isotope Substage 5e**. The same is valid for the same deep sea stage and for some parts of the **Sangamon Interglacial** of western North America.

The **Holocene** can easily be dated by radiocarbon analysis or by dendrochronology. Thus, for the Holocene climatic optimum, no major problems should be expected in synchronizing events in various parts of the globe.

The maximum cooling during the **last pleniglacial**, based on sea surface temperatures reconstructed by **CLIMAP**, occurred about 18,000 years B.P. That is the reason, why for this atlas the ecological and climatological situation of the continents are reconstructed for the same period, though the accuracy of the datings available is not very high. This means that one has to rely very often on sediments or landforms which were formed during an extreme cold phase after the preceding interval which was predominantly interstadial in character (between about 35,000 and 25,000 ^{14}C years B.P.) and before the beginning of the late glacial. The phases of warmer climates cause fewer problems in age determination than the last pleniglacial does. Other approaches to date the last pleniglacial are to trace endmoraine systems over vast regions of the continents, or to correlate Oxygen Isotope Stage 2 off the West African coast or in the northwestern part of the Indian Ocean by means of pollen analysis which reflects paleoecological changes on the continents.

Thus the **Upper Weichsel, Late Valdai, Sartan, Late Wisconsin**, and **Oxygen Isotope Stage 2** can reliably be correlated with each other.

One of the major paleoclimatic problems is to explain, when and how those inland ice masses began to form actively, which later on extended over vast areas of the Northern Hemisphere during full-glacial times. Isotope records of inland ice masses in Greenland and Antarctica, as well as in several deep sea cores show that this must have happened during **Oxygen Isotope Stage 3** or in its equivalent **interstadial complex**, the traces of which are frequently encountered on the continents. They can provisionally be dated by ^{14}C or U/Th techniques. Yet, due to the difficulties in reliable dating, in our reconstructions we concentrated on the broad time interval between approximately 35,000 to 25,000 years B.P., even though from a paleoclimatological point of view this was a phase of repeatedly changing climates.

To **reconstruct paleoclimates quantitatively** or to map past vegetation types requires reliance on exact methods and criteria. Yet since the biosphere has changed so profoundly during the Last Interglacial to Holocene cycle, the problem of **formulating transfer functions** or of **finding modern equivalents** of past biota is extremely difficult. For the solution of general problems, it became necessary to present various characteristics obtained using different approaches. Because of this, different versions are sometimes given in the atlas which may not be in full accord and which may provide regional differences. Moreover, in some cases variable parameters were mapped (e.g., mean temperature deviations from present-day values for January and July or minimum deviations of mean temperatures for February and August from present-day values). At our present state of knowledge it seems better to illustrate the different reconstructions than to rely on one version only.

Spatial reconstructions in the maps of the atlas show climates and the state of the environmental components during characteristic phases of the evolution of the biosphere during the last complete global climatic macrocycle. The assumed position of the chronological levels in concern to the curve of the **atmos-**

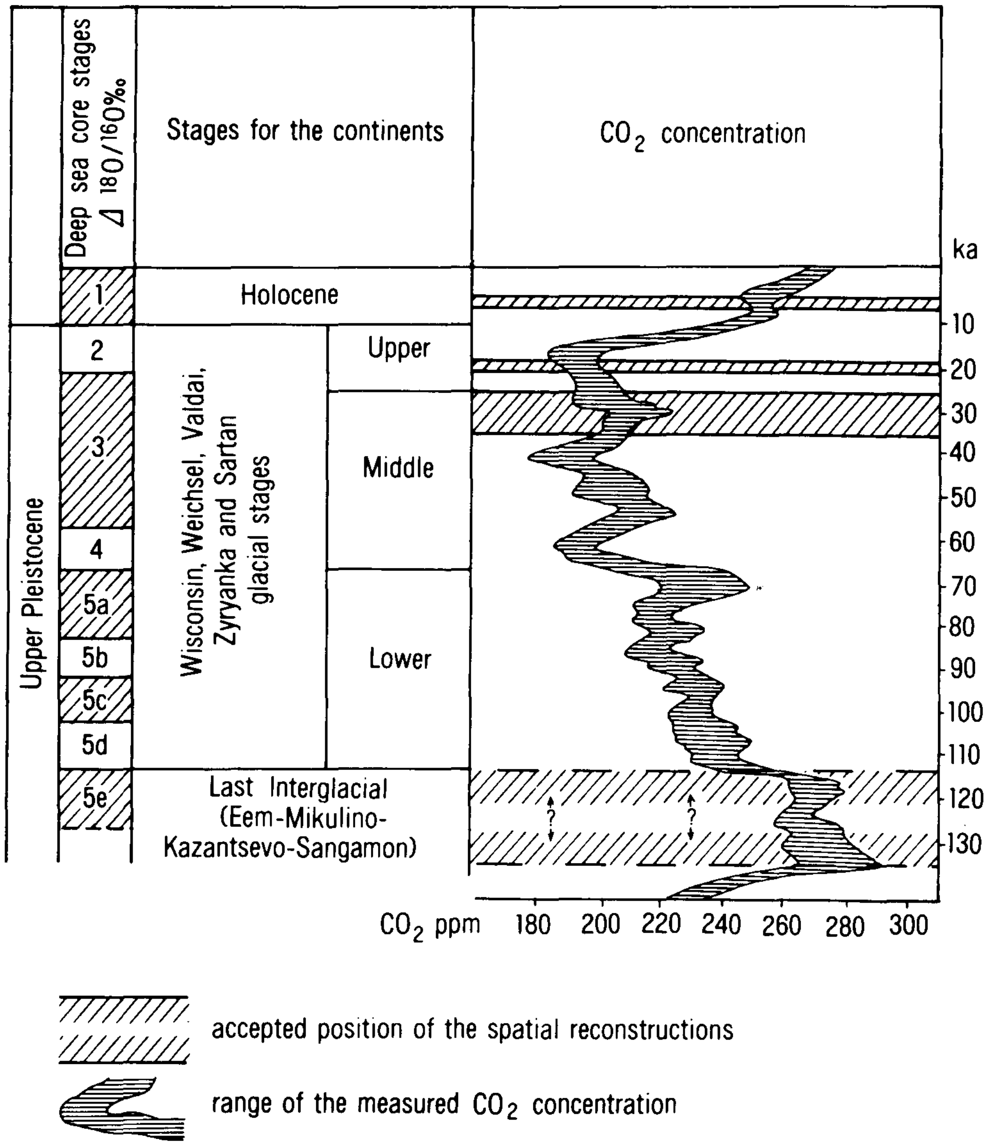

Fig. 1. Changes of the atmospheric CO_2 concentration during the last glacial cycle (after JOUZEL *et al.*, 1989)

pheric CO_2 concentration (based on the ice core investigations at Vostok research station, Antarctic) is demonstrated in *Fig 1.* Their position is not equally reliable in different parts of the curve, just as the curve itself is not equally well calibrated throughout. The latter is more adequate in the upper (younger) section, within the limits of the radiocarbon dating.

Paleogeographic reconstructions of the Last (Eem) Interglacial time are more ambiguous. It is generally accepted to correspond to Oxigen Isotope Substage 5e; the CO_2 curve, however, indicates an earlier date for the carbon dioxide maximum to occur. Thus the chronological interval corresponding to the warmest (interglacial) epoch as well as the position of its optimum appears somewhat shifted to the past, which is indicated in the graph as an alternative.

A more detailed chronological calibration of the atmospheric CO_2 concentration curve, as well as of the ^{18}O isotope curve of deep sea cores for the earlier phases of the macrocycle requires further investigations.

LAST INTERGLACIAL CLIMATIC OPTIMUM (about 120,000 yr B.P.)

VEGETATION DURING THE LAST INTERGLACIAL

by V. P. Grichuk

(Explanatory notes, to map on page 11)

The map shows the distribution pattern of reconstructed vegetation types over the continents of the Northern Hemisphere. It wasdrawn on the basis of a critical revision of published paleobotanic data of interglacial sediments, as well as on historical floristic materials (WULF, 1944; MALEEV, 1940; CHANEY, 1936 and others). The latter were applied mostly to areas such as arctic or tropical regions, with little or no paleobotanic information. Boundaries were drawn taking topography into consideration. Continental shorelines were derived from the recently published Quaternary maps and from other data in the literature.

The construction of the map was difficult since it was the first attempt to map such a vast area. Up to now paleobotanic maps covered only separate parts of Eurasia [e.g., the Russian Plain (GRICHUK, 1964), Siberia (GITERMAN *et al.*, 1968), the USSR territory (VELICHKO, 1984), extratropical Eurasia (FRENZEL, 1968), and Europe (GERASIMOV and VELICHKO, 1982)].

The most difficult task was to choose the most representative chronological interval within this rather long interglacial which could be identified in sections all over the hemisphere. The resulting map characterizes the interval which corresponds to the maximum heat supply at the culmination of the **climatic optimum**. This phase can be easily identified on any pollen diagram as the **transition from a thermoxerotic to a thermohygrotic phase** in the Pleistocene rhythm (GRICHUK, 1960).

The map was primarily composed using paleobotanic data, including information from 162 sites in Europe, and 119 sites in Siberia. For other territories, fairly conclusive data were obtained from 72 sites chosen from the reference publication "Bibliographie Palynologie" published from 1956 to 1983 (Montpellier, University of Languedoc).

The relevant sections had to be complete enough as to be sure that the horizon in concern really belonged to the climatic optimum. In some sections where the climatic optimum was only partially represented, this interval was identified more or less arbitrarily.The presence of closely spaced sections in many regions helped to reveal some serious errors. If, on the other hand, sections were scattered over vast areas, only a conceptual control was possible. We could not show vegetation in West Africa due to the lack of pertinent data.

The map shows that vegetation zones during the interglacial climatic optimum were similar to present-day ones. The absence of **polar deserts** and a more limited distribution of **tundras** were the main differences observed. Typical tundra formations appeared only in northeastern Siberia and in northeastern North America. Palynological data indicate that tundra was present in the south and southwest of Greenland, where it expanded farther south than on the continents, while the central part of the island was presumably ice-capped.

Boreal and **broad-leaved nemoral forests** were, just like today, most important. A marked northward expansion of the forest zone is recorded on the map, especially in North America. Paleobotanic evidence for this was obtained from only two sections in America while in northern Asia it was recorded at 12 sites.

Boreal forests, based on coenose-forming species, were divided into northern and southern subzones. The southern zone was distinguished by the occurrence of broad-leaved species on the plains, and of *Pinus sibirica* in the mountains. Data available at present do not permit to subdivide the boreal forests of North America in a similar way. In Eurasia, interglacial boreal forests differed from their modern analogues in the constant presence of birch (ex sect. *Albae*). Dark coniferous forests of *Picea*, *Pinus sibirica*, and *Abies* only grew in mountainous regions.

During the Last Interglacial broad-leaved nemoral forests were more widespread in the East European Plain than today. Their northern boundary shifted markedly northward up to the Gulf of Finland, which formed part of a strait connecting the Eemian and Boreal sea basins. In North America, the northern boundary of broad-leaved forests was displaced considerably northward near the Great Lakes, as well as further to the east. This displacement of broad-leaved forests was less pronounced in Eastern Asia. Eastern European formations of nemoral type were characterized by a particular composition of dominant species (primarily *Carpinus betulus*), which has no analogue in the present- day vegetation of Eastern or Western Europe.

Formations similar to present-day subtropical tree and bush communities were recorded within the present areas of their habitat. The same is true of sub-xerophilous trees and shrub communities dominated by deciduous and evergreen species. Although paleobotanic data are rather limited, the presence of this type of vegetation has been unambiguously established. It was most widely distributed in North Africa.

Sub-xerophilous and xerophilous herb-grass and sagebrush grass **steppes** were found only in southeastern Europe, near the Black and Caspian seas. In Western Siberia, the southern boundary of this vegetation lay far to the south of its present position; the same is valid for Central Asia (Mongolia). In North America the area of steppes seems to have been approximately the same as it is today.

The reconstruction of all the other types of vegetation is based almost entirely on **historico-floristic data**. The boundaries are mostly hypothetical and take topography and surface deposits into account. Alpine vegetation was reconstructed completely from historio-floristic data. This, however, does not diminish the reliability of the reconstruction. Endemic species (paleoendemics) with narrow distributional ranges indicate that alpine vegetation existed in the mountain regions for a very long time (much longer than the Late Pleistocene).

CLIMATES DURING THE LAST INTERGLACIAL

(Deviations from present-day values)

by A. A. Velichko, V. P. Grichuk, E. E. Gurtovaya, E. M. Zelikson, M. S. Barash and O. K. Borisova

(Explanatory notes to maps on pages 13, 17, 21 and 25)

General remarks. Since our theoretical knowledge about past climates, their spatial distribution, and causes for their change is inadequate at present, precise data (especially quantitative) on paleometeorological parameters for some of the more representative time intervals are of primary importance. Climatic reconstructions for the Last Interglacial (Eem—Mikulino—Kazantsevo—Sangamon) are especially interesting and can be used as models for the climates of the nearest future. The compilation of such paleoclimatic maps for the Northern Hemisphere was undertaken by a team of researchers at the Institute of Geography, Academy of Sciences of the USSR as a part of a large scale reconstruction of interglacial climates in Europe.

In the first place quantitative estimates for paleoclimate parameters were obtained for a series of localities in the oceans as well as on land. These were used to compile maps for the distribution of the January, July, and annual mean temperatures and that of the annual mean precipitation during the interglacial optimum, as well as for their deviations from present-day values. Differences between sea surface and air temperatures defined for present-day conditions in various latitudes (STROKINA, 1982) were also taken into account in the reconstruction of the air temperatures during the Last Interglacial.

Methods of obtaining paleoclimatic data. Reconstructions of paleoclimates on *land* were done on the basis of temperature and precipitation estimates from fossil plant data. The method was developed by GRICHUK (1969, 1973) based on concepts by SZAFER (1946) and IVERSEN (1944).

Two methods can be used to obtain paleoclimatic parameters. The first one consists of identifying the center of concentration of the fossil flora, that is, a region where all the species grow at the present time. The region is determined cartographically by superimposing the modern areas of all the species found in the fossil flora. Since climate is a key factor in plant distribution, the region where all of the plants grow together today is assumed to be a modern climatic analogue of the region represented by the fossil flora. Ranges of present-day climatic parameters for the analogous region are obtained from meteorological stations. If they are close enough, their average values can be taken as approximate values for paleoclimatic indices (July, January, and annual mean temperatures).

Another method consists of determining the climatic ranges of each species found in the fossil assemblage (climagrams) and combining them to establish common climatic fields. In practice, usually two climagrams are composed for each species, one showing temperatures of the coldest and warmest months (January and July), and the other for the total annual precipitation and the duration of the frost-free period.

These two methods permitted us to estimate July and January temperatures to the accuracy of ±1 °C, and total annual precipitation to the accuracy of ± 50 mm.

Paleotemperatures over the *ocean* were calculated from planktonic foraminifera. New methods were developed for estimating paleotemperatures using quantitative correlations between modern thanatocoenoses and sea surface temperatures and then comparing the fossil thanatocoenoses with present-day ones. (BARASH, 1970, 1980; BARASH and BLYUM, 1975; BARASH and OSKINA, 1978). The mean annual temperatures estimated by this method are accurate to ± 1.5 to 2 °C.

CLIMAP project members used the composition of planktonic foraminifera to reconstruct the surface characteristics of the ocean by applying factor analysis and regression equations (IMBRIE and KIPP, 1971). The standard error (at the 80 per cent confidence level) ranged from 1.5 to 2.8 °C for various regions of the World Ocean (KIPP, 1976).

Data used. Paleoclimatic recontructions for terrestrial areas are given for the last interglacial optimum which corresponds to the M_6 pollen zone in GRICHUK (1961) and the "g" zone in JESSEN and MILTHERS (1928), and represents a **transition from a thermoxerotic to a thermohygrotic phase** which is associated with the maximum heat supply. The map for the Northern Hemisphere was compiled using data from 47 points on land. Most of these are from Europe, fewer from northern Asia, and only 3 from eastern North America.

In the deep sea cores, the last interglacial horizon is identified by using paleotemperature curves, sedimentation rates, and sediments lithology. It corresponds to the 5e stage of oxygen isotope record as identified by SHACKLETON and OPDYKE (MANGERUD, 1970; MANGERUD *et al.*, 1981) and dated to 125, 000 years B.P.

Reconstructions for the northern Atlantic region are based on the distribution of planktonic foraminifera and on paleotemperatures obtained by the research teams at the Institute of Oceanology, Academy of Sciences of the USSR, and at Moscow State University. Both used the method developed by BARASH. Numerous data published by CLIMAP project members were also involved. The total number of data points in the northern Atlantic was 55 *(Fig. 2)*.

Information on northern Pacific was gathered and summarized by BLYUM and NIKOLAEV (BLYUM, 1981; BLYUM *et al.*, 1982; NIKOLAEV, 1981; NIKOLAEV *et al.*, 1982). Some data were obtained by BLYUM herself, others were based on the composition of fossil foraminifera published by other specialists. Methods used, again were those developed by BARASH. In some cases the temperatures shown were calculated from Radiolaria, coccoliths, and oxygen isotope data (BELYAEVA, 1978; CHEKHOVSKAYA, 1975; ICHIKURA and UJIIE, 1976; KELLER, 1978; MOORE, 1973 and others). Only 30 points with paleoclimatic parameters were available for the northern Pacific, and, unlike the case for the Atlantic and for the continents, only mean annual temperatures and their deviations from present-day values could be reconstructed.

Data available at present come from a rather limited number of localities. They do, however, permit us to present a tentative picture of the distribution of climatic features over the Northern

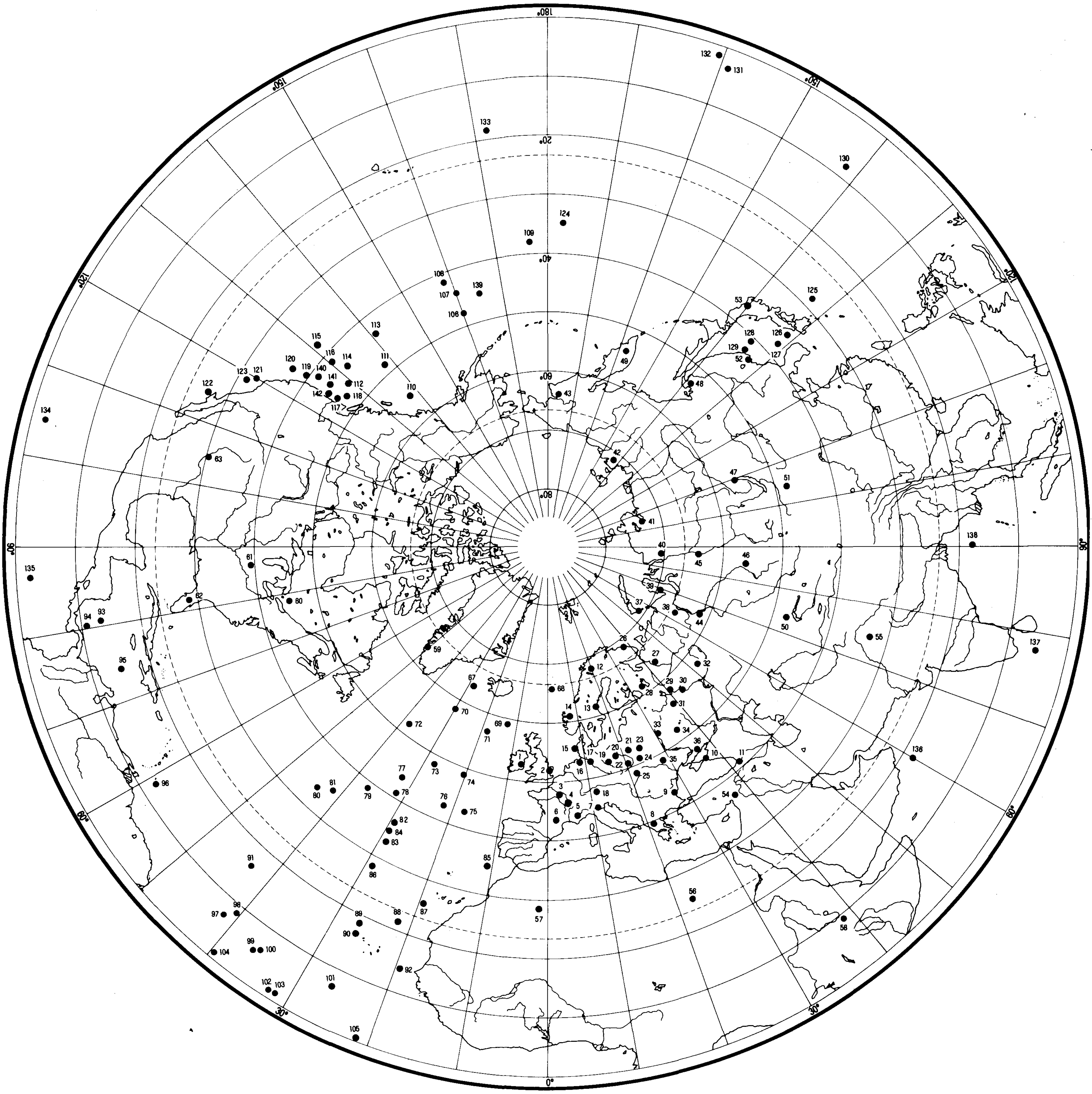

Fig. 2. Sites used for the construction of maps on pages 13, 17, 21, and 25. Deviations from present-day values: ΔT_I = January mean temperature; ΔT_{VII} = July mean temperature; ΔT_Y = annual mean temperature; ΔP = annual precipitation

No	ΔT_I	ΔT_{VII}	ΔT_Y	ΔP_{mm}	No	ΔT_I	ΔT_{VII}	ΔT_Y	ΔP_{mm}	No	ΔT_I	ΔT_{VII}	ΔT_Y	ΔP_{mm}
1	1	0	0	450	22	3	0	0	700	43	-1	4	1	0
2	2	0	0	240	23	4	0	2	700	44	4	0	2	100
3	-1	1	0	300	24	4	0	1.5	500	45	8	1	3	-50
4	0	2	2	100	25	6	2	1.5	0	46	7	0	1.5	150
5	-2	3	1	-	26	2	1	1	100	47	8	4	5.5	50
6	-1	2	1	300	27	3	1	3	100	48	4	3	1	200
7	1	0	-1	0	28	7	1	5	120	49	4	0	1	90
8	-3	-2	-1	0	29	10	0	6	0	50	-1	-3	-3	100
9	-1	-2	0	600	30	11	0	6	-20	51	5	1	1	120
10	2	-1	0.5	600	31	8	0	4	30	52	0	0	0	0
11	0	0	0	0	32	11	1	4.5	50	53	0	-2	-1	-
12	7	3	5	100	33	6	0	3	0	54	-1	-1	-1	-
13	2	1	1	100	34	7	1	3	50	55	0	0	0	0
14	0	2	0	200	35	3	1	1	300	56	-2	-1.5	-2	150
15	1	3	3	-	36	1	-2	1	300	57	-2	-4	-3	150
16	0	2	0	500	37	5	8	6	50	58	0	0	0	0
17	0	2	2	-	38	3	4	4	50	59	0	4	2	-
18	2	2	2	0	39	3	8	5	200	60	4	2	4	-
19	2	0	1	500	40	7	8	6	200	61	-0.9	-0.4	0	-
20	2	0	1	700	41	13	8	2	200	62	-0.3	-0.1	-1	-
21	2	0	0	500	42	12	4	5	40	63	-3	-2	-2.5	-

Hemisphere. Due to the scanty paleofloristic data, the reconstructions for the southernmost tropical parts of the land are speculative.

Winter temperatures (map on page 13).The reconstructed parameters for both land and oceans are in good agreement. This provides additional evidence that the data are synchronized correctly. The reconstructed temperatures, on the whole, reveal a pattern similar to the present-day one. January isotherms follow a meridional direction over Eurasia, and a latitudinal direction over the Atlantic Ocean. This indicates that the position of the atmospheric centers at that time was similar to those of today. The values for the reconstructed parameters, however, significantly deviate from the present-day ones.

The most distinctive feature of the interglacial optimum was a warming of the north. This was pronounced in northeastern Europe and northern Asia. Within the zone now dominated by the Siberian anticyclone, January temperatures rose by 6 °C (or slightly more) compared to present-day values. This indicates that the Siberian High was weaker then over all of the area, and that it hardly influenced the Russian Plain.

Considerably higher winter temperatures (by 12 °C) in the Siberian sector of the Arctic, as well as in Northern Europe, indicate the increased activity of the Gulf Stream. This is also confirmed by distinctly higher temperatures of the northern Atlantic between Iceland and Greenland. The warm current was stronger and penetrated deeper eastward than nowadays.

Temperatures in Western Europe (unlike those in Asia and in Eastern Europe) were but slightly higher than at present. Some regions show no deviations from present-day values. In the Northern Hemisphere, as a whole, winter temperatures were higher during the interglacial optimum than today. This regularity was distinct on the continents north of about 45° N, and in the Atlantic north of 35° N. The same trend is recorded in North America.

A zone of some cooling is noted south of these latitudes. The decrease in temperature was small (about 1 °C or less in most regions) but it was a constant feature. It can be traced in eastern Kazakhstan as well as farther to the west, including all of Mediterranean Europe and a considerable latitudinal band in the Atlantic (at about 15° to 18° N).

Summer temperatures (map on page 17).The pattern of July isotherms for the last interglacial climatic optimum resembles the present-day one which is mostly latitudinal. The isoanomalies show the same trend, which was noted above for past and present winter temperatures. The deviations from present-day values, however, were less than those for winter temperatures. For some regions no deviation was recorded. This regularity is typical for thermal changes throughout the Late Cenozoic when climatic fluctuations were most distinct in winter temperatures.

Within a certain circumpolar area, including high and partly middle latitudes, interglacial summer temperatures show positive anomalies. Increased temperatures were recorded for Arctic Siberia (from Yamal to Taimyr). Temperatures in Scandinavia and in the North Atlantic were markedly higher than at present (by about 3 °C). These data suggest that the influence of the Gulf Stream was more pronounced at that time than today and this accounts for the warming of the Arctic.

The range of positive deviations in summer temperatures (compared to present-day values) within the circumpolar regions increased eastward from the Atlantic Ocean, towards Siberia.

The latitudinal zone of negative deviations for July temperatures was wider than that for January. In Eurasia, summer cooling was more distinct than the winter one, and reached 2 °C. In the Atlantic, the deviations did not exceed -1 °C. Higher temperatures than today were recorded both within a narrow band between 10° and 20° N and in the equatorial zone.

Annual mean temperatures (map on page 21). This map showing the general picture of heat distribution is based on more abundant data than the maps of seasonal temperatures, because additional information on annual mean temperatures was available for the Pacific (for example, data for points off the western coast of North America). In the Arctic, the area of low temperatures was considerably reduced. The lowest temperatures were found in the east-central Arctic.

Annual mean temperatures were also above present-day

Fig. 2 (continuation)

No	ΔT_{VII}	ΔT_{I}	ΔT_{Y}	No	ΔT_{VII}	ΔT_{I}	ΔT_{Y}	No	ΔT_{VII}	ΔT_{I}	ΔT_{Y}
67	4.38	2.86	3.62	93	1.3	-0.3	0.5	119	-	-	3
68	1.58	-0.22	0.68	94	0.2	-0.9	-0.35	120	-	-	0
69	-1.7	-2.0	0.5	95	1.5	-1.2	0.2	121	-	-	0
70	2.2	3.0	1.5	96	0.2	0.3	0.2	122	-	-	0
71	1.0	1.8	0.9	97	-0.4	-0.7	-0.5	123	-	-	1
72	1.0	0.5	0.5	98	1.0	0.7	1.0	124	-	-	0
73	0.2	0.7	0	99	-0.2	-0.1	-0.2	125	-	-	1
74	0	0	0	100	0.0	-2.2	-1.1	126	-	-	1
75	-3.4	3.0	-0.2	101	-0.2	-2.6	-1.4	127	-	-	1
76	-0.8	2.6	0.9	102	-0.7	-0.2	-0.45	128	-	-	0
77	0.5	0.3	0.4	103	0.0	0.0	0.0	129	-	-	1
78	1.0	1.0	1.0	104	0.1	-0.9	-0.4	130	-	-	1
79	-0.7	1.3	0.3	105	0.3	1.4	0.85	131	-	-	1
80	-1.0	0	-0.5	106	-	-	-1	132	-	-	0
81	-1.9	0.7	-0.6	107	-	-	0	133	-	-	3
82	-0.4	0	-0.2	108	-	-	0	134	-	-	-
83	-	-	-1.3	109	-	-	0	135	-	-	1
84	-2.5	-2.2	-0.2	110	-	-	3	136	1.5	-0.5	-
85	0.1	0	0.6	111	-	-	0	137	-2.0	-1.5	-
86	-1.2	-1.0	1.1	112	-	-	2	138	-1.9	-0.5	-
87	-1.2	-1.6	-1.4	113	-	-	1	139	2	-	-
88	-0.5	-0.7	-0.6	114	-	-	1	140	-1	-	-
89	-0.3	0.5	1.0	115	-	-	0	141	2	-	-
90	-0.8	-2.0	-1.4	116	-	-	1	142	1	-	-
91	0.4	0.8	0.6	117	-	-	2				
92	-0.7	4.7	2.0	118	-	-	2				

values over vast areas in high and middle latitudes of the Northern Hemisphere. Temperatures lower than today were recorded mainly in a belt between 20° and 45° N.

A more complicated pattern of positive and negative deviations was found in the Atlantic. Here, an area of steady positive deviation was found in the north. Temperatures were also higher than at present between 10 °and 20° N. They were lower between 10° N and the equator, and higher at the equator itself.

In general, most of the reconstructed parameters indicate that during the period in concern considerable warming occurred compared to modern climates in middle and especially in high latitudes.

Annual mean precipitation (map on page 25). The distribution of precipitation during the climatic optimum of the Last Interglacial was more uniform than at present. This was primarily due to a marked positive deviation in precipitation as compared to present-day values within an area between 40° and 50° N. A general trend towards increased precipitation was observed in northern regions, too. The absolute values for increased precipitation, however, clearly cannot characterize their influence on the water balance of the different areas. For example, a 50 mm increase in the total annual precipitation today will represent an 8 per cent increase in the precipitation on the central Russian Plain, and a 70 per cent rise along the Arctic coast of Siberia.

Some distinct patterns can be observed in these precipitation deviations. While the most pronounced warming during the last interglacial optimum was reported for northeastern Eurasia, the most significant increase in precipitation occurred in western and southern regions of the continent (by more than 200 to 300 mm). Areas adjacent to the tropics, in general, do not reveal any substantial deviation in temperatures from present-day values. In some regions, however, somewhat lower temperatures and higher precipitation were registered than those observed today.

CLIMATES DURING THE LAST INTERGLACIAL

(Minimal deviations from present-day values)

by B. Frenzel, in charge of the Research Project Group "Terrestrial Paleoclimatology", Federal Republic of Germany
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(Explanatory notes to maps on pages 15, 19, 23 and 27)

General remarks. In general, climates of the Pleistocene Interglacials were warmer during their climatic optima than today. The same holds true for the Holocene climatic optimum. Thus, a comparison of the climatic optima of these interglacials has relevance for the near future, when a rise in temperature is expected due to increases in the concentrations of greenhouse gases in the atmosphere.

The Last Interglacial lasted from approximately 125,000 to 115,000 years B.P. Its climatic optimum, at least in Europe, consisted of two phases, the earlier of which was characterized by warm summers and only somewhat warmer winters than today (phases IIIc to IVb, according to SELLE, 1961; for further information see FRENZEL, 1991a). The immediately following phase experienced generally cooler summers, yet remarkably mild winters, when inland ice masses began to accumulate in polar areas (TURON,1984). As shown by investigations of ice cores from Antarctica, the CO_2 concentration in the atmosphere at the time approached the present-day one (LORIUS, 1989; LORIUS *et al.*, 1988). Although both these phases in the evolution of climate during the European optimum of the Last Interglacial are characteristic, the same cannot be shown for the other continents, or for the oceans. For the continents, this is due to the different vegetation history of Europe compared to the other continents, and to a reduced number of reliable indicators of former climates for continents other than Europe as far as the Last Interglacial is concerned. Thus the climatic data for the Last Interglacial given in maps on pages 15, 19, 23 and 27 are not of the same precision everywhere.

During the Last Interglacial climatic optimum, some coastal areas were flooded by the rising ocean. This is well known for most parts of Eurasia and for southern North America. Yet, records of interglacial transgressions in the Canadian Arctic are scanty, if they exist at all.

On the other hand, the Polar Sea became warmer and transgressed at the time south in northern Eurasia. This enabled certain marine warm-water species to penetrate from the Atlantic Ocean far northeast, even up to the Taimyr Peninsula (e.g., ANDREEVA, 1982; ANDREEVA and KIND, 1982; GUDINA, 1976). The climatological consequences of these "warm-water transgressions" in Northern Siberia have not been sufficiently analysed up till now.

It is suggested that the Late Khazar Transgression of the Caspian Sea dates from that time (FEDOROV, 1978; ARSLANOV *et al.*, 1988).

Methods used. Due to the difficulties in dating and in definition of the climatic optimum of the Last Interglacial, data are given in maps on pages 15, 19, 23 and 27 for the transition period between the two European episodes of the interglacial climatic optimum mentioned, or for the warmest climates in other regions of the Northern Hemisphere. The data reconstructed for the continents by the Research Project Group "Terrestrial Paleoclimatology", Federal Republic of Germany are presented in maps on pages 15 and 19. It is suggested that the coldest month on the Northern Hemisphere was as a rule February, though January can also be included in these calculations.

The terrestrial data are obtained from former distribution patterns of various plant and animal taxa and of whole plant communities *(Fig. 3)*. These plant communities were not exact analogues of present ones, due to migration differences and different types of competition. Nevertheless, the paleoecological situation of certain interglacial plant communities was comparable with those of closely related modern plant communities. This mainly holds true for the area distribution of boreal forest, the composition of which during the climatic optimum of the Last Interglacial was clearly similar to its Holocene counterpart.

The most widely used method was to compare former vegetation communities and soil types to their nearest modern equivalents, in order to obtain quantitative estimates for paleoclimates of the Last Interglacial. Comparisons were made only for similar latitudes, thus avoiding those between northerly and southerly regions, since it is assumed that differences in daylength may exert a profound influence on biological responses. Moreover, our reconstructions did not exclusively rely on the former distribution patterns of plant and animal species or on the suggested present-day climatic boundaries of the distribution patterns. This is due to the observation that neither during the Last Interglacial, nor during the Holocene have all plant and animal taxa had sufficient time to occupy their climate-controlled optimal distribution areas. On the contrary, there exists substantial evidence that spontaneous migrations of various plant and animal taxa are still active today. This must negatively influence the climatic interpretation of distribution areas. These problems were extensively discussed by v. KOENIGSWALD (1988) for some large herbivores of the German Last Interglacial fauna.

If possible, the distribution pattern of former permafrost was also taken into consideration. It was impossible to differentiate between various types of permafrost (i.e., continuous permafrost, discontinuous permafrost, isles of permafrost in prevailing talyk) and only the southern boundaries of permanently frozen ground were taken into consideration.

At any rate, only the minimum deviations of former climates from their present-day equivalents are given in maps on pages 15, 19, 23 and 27, not the maximum ones. We only relied on data of the fossil record, not on computer models.

Winter temperatures (map on page 15). It is manifest from this map that during the climatic optimum of the Last Interglacial the mean temperature of the coldest month was about 1.5 to 3 °C higher than today almost everywhere on the continents of the

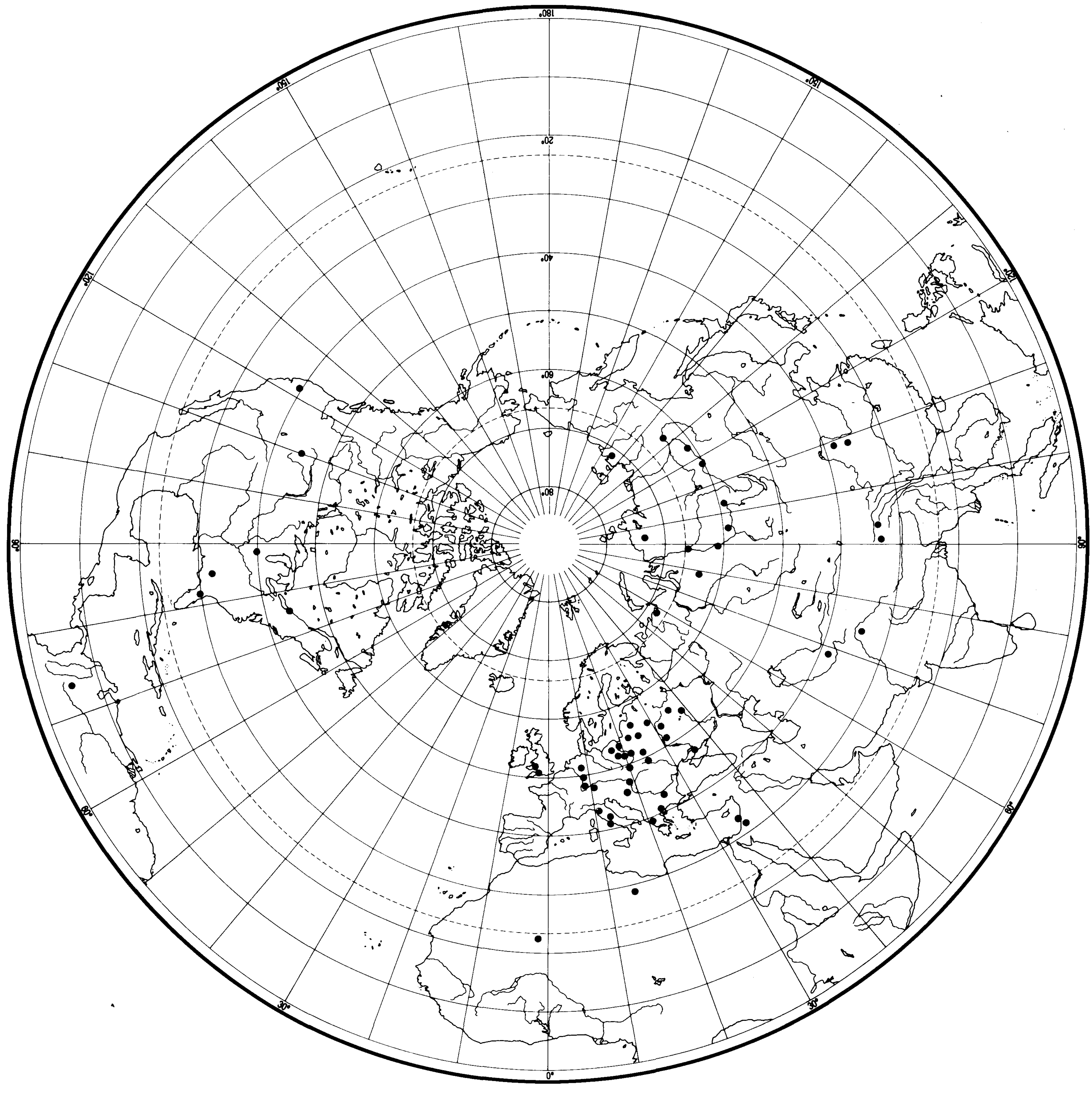

Fig. 3. Sites used for the construction of maps on pages 15, 19, 23 and 27

Northern Hemisphere. According to the international literature, the same seems to have held true for the Southern Hemisphere as well. The **CLIMAP** data (1984) indicate a similar pattern for the oceans, perhaps with the exception of cooling in equatorial waters, primarily in the western subtropical Atlantic and in the Caribbean Sea. This seems to be corroborated by the air temperatures in the Andes of present-day Colombia. These lower tropical and subtropical sea surface temperatures were probably caused by strong evaporation.

The combined terrestrial data on the boreal and subpolar areas indicate even higher deviations (about 4 °C), of interglacial air temperatures, than those prevailing farther south. This conforms with the results of climate models constructed for the near future, when the air temperatures are expected to rise more dramatically in northern regions than in the south.

Some data point to considerably warmer winter climates in the central parts of Eastern Europe than today, yet these values are not always corroborated by independently obtained data. Various indicators of climate might have influenced the reconstructions in a different way. Another explanation can be that in Eastern and Central Europe, data for the two climatically different phases of the Last Interglacial optimum were superimposed. Yet, perhaps a third explanation is the correct one: it can be shown that the annual mean precipitation during the climatic optimum of the Last Interglacial was much higher in Eastern and Central Europe as compared with the present day, than in western Eurasia owing to a pronounced influx of moist air masses during winter far into Eastern Europe. This seems to be supported by comparable trends in the Holocene history of climate.

Summer temperatures (map on page 19). Comparable to what was stated for the coldest month during the climatic optimum of the Last Interglacial, the mean air temperatures of the warmest month of that time seem to have been nearly everywhere appreciably higher than today. It is suggested that August was the warmest month, at least in more continental climates. The interglacial phases investigated were the same as for the coldest month. Though CLIMAP (1984) reconstructed sea surface temperatures for the Last Interglacial Atlantic Ocean, these data were not given on page 19, because they are too scarce and no isotherm maps can be drawn.

The general impression is that the mean summer temperatures were higher by about 2 to 3 °C than today. Yet, farther north, there is no comparable rise in summer temperatures like the coldest month increases. This again shows a general increase in humidity or oceanity of winter climates within the boreal and subpolar belts during the Last Interglacial.

Sea surface summer temperatures in tropical and subtropical oceans seem to have been similar to today or even slightly cooler, including those for northernmost South America. This suggests increased evaporation from tropical and subtropical oceans, similar to the explanation for cooler coldest month temperatures.

In contrast to winter temperatures observed for Eastern Europe, summer temperatures of this region did not show any indications for a positive anomaly compared to present conditions. This again supports the idea on increased oceanity during the winter, caused by relatively mild and moist air masses invading Eastern Europe from the north, much more frequently than today.

Annual mean temperatures (map on page 23). For annual main air temperatures over the continents, the data were not calculated from those of the warmest and of the coldest months, but independent indicators were used. In this respect, the southern limit of permafrost in Siberia turned out to be important. This can be determined using syngenetic permafrost features and frozen cadavers of various animal species, dating from the Last Interglacial or from earlier periods of the Pleistocene.

All the indicators taken together show that mean annual air temperatures of that time were approximately 2 to 3 °C higher than today, with somewhat higher values for Eastern Europe. Moreover, some indicators point to a relative rise in annual mean air temperatures for the present-day northernmost boreal forest and for the tundra belt. Both these observations suggest that the estimates of mean temperatures of the coldest month for the climatic optimum of the Last Interglacial are realistic.

Thus it seems that the mean temperatures of the climatic optimum of the Last Interglacial period are an interesting starting point for speculations about the future development of climate, when the greenhouse effect should cause a significant increase in global temperatures, since data obtained for the Last Interglacial are similar to mathematical predictions for the near future. This leads to the problem of available moisture at that time (see map on page 27).

Annual mean precipitation (map on page 27). The distribution patterns and characteristics of all past moisture indicators were controlled by water balances and not by precipitation alone. Thus, in studying the problem of past precipitation, only the moisture budget available for the growth of plants and plant communities or reflected in lake-level oscillations can be obtained, not the exact data for the precipitation itself. Since all the relevant observations for the past were compared with present-day precipitation under similar biotic, edaphic or geomorphological conditions, one cannot rely on absolute values, but one should be aware of the general trends. Moreover, anomalies in humidity or aridity for various regions of the Northern Hemisphere during the Last Interglacial optimum can only be compared to today's conditions, without any direct data on "precipitation".

For a better understanding of paleoclimate it would be necessary to collect information on the seasonal distribution of precipitation. This might be reconstructed on the basis of distribution patterns of plants which depend on summer rains or on winter moisture, as was demonstrated for the Holocene of southwestern North America (MARTIN, 1963). No data of this type are yet available for the Last Interglacial. Although the data shown in map on page 27 are of limited value, it still seems worthwhile to discuss them.

It is often suggested that a rise in air temperatures by about 2 to 4 °C, as expected for the near future through increased greenhouse gases, would cause further aridity in presently dry regions. As seen in the map in concern, just the opposite happened almost everywhere during the climatic optimum of the Last Interglacial. It should be added that the same also occurred in all other interglacials for which paleovegetation and paleoclimate reconstructions were possible. Yet, regional deviations existed from this general picture. This applies to the present prairie regions of North America, which became drier in contrast to the increased moisture for the whole of the Mediterranean area, the Sahara, and presumably for Middle and Central Asia. The same general tendency prevailed during the Holocene (see map on page 79), with a somewhat changed distribution pattern of arid regions, in comparison with the Last Interglacial. Thus, although generally warmer climates of the past were accompanied by increased humidity over most of the present semiarid or arid regions, there existed regional differences in atmospheric circulation, causing variable distribution patterns of moisture, during the interglacial epochs.

During the Last Interglacial, moister conditions in subtropical and tropical continental belts may have been caused by increased evaporation from the adjacent oceans, as was suggested earlier for sea surface temperatures of the warmest and coldest months.

Several observations draw attention to relatively greater moisture availability in Eastern Europe, as already discussed for temperatures of the coldest month. Simultaneously the oceanic fringes of Eastern Asia and of North America also seem to have received a larger amount of moisture than today. This is also apparent for the loess regions of Northern China and of northern, eastern and southern Tibet (ZHOU *et al.*, 1981; LI *et al.*, 1979; LIU, 1988), thus indicating a probable increase in monsoon activity.

INTERSTADIAL OF THE LAST GLACIATION
(about 35,000 to 25,000 yr B.P.)

CLIMATES DURING INLAND ICE FORMATION

by B. Frenzel

(Explanatory notes to maps on pages 29, 31, 33, 35 and 37)

General remarks. During the Last Glaciation, inland ice masses formed several times, yet their distribution patterns were not always identical. For this stage of the Last Glaciation maps on pages 29, 31, 33, 35 and 37 are shown *(Fig. 4)*. It is well known that those ice advances, which belonged to the classical Weichsel, Upper Würm, Late Wisconsin, Late Valdai glaciations had their maximum extensions to the south at about 22,000 to 18,000 years B.P. Yet it remains an unsolved dilemma, when and how these ice

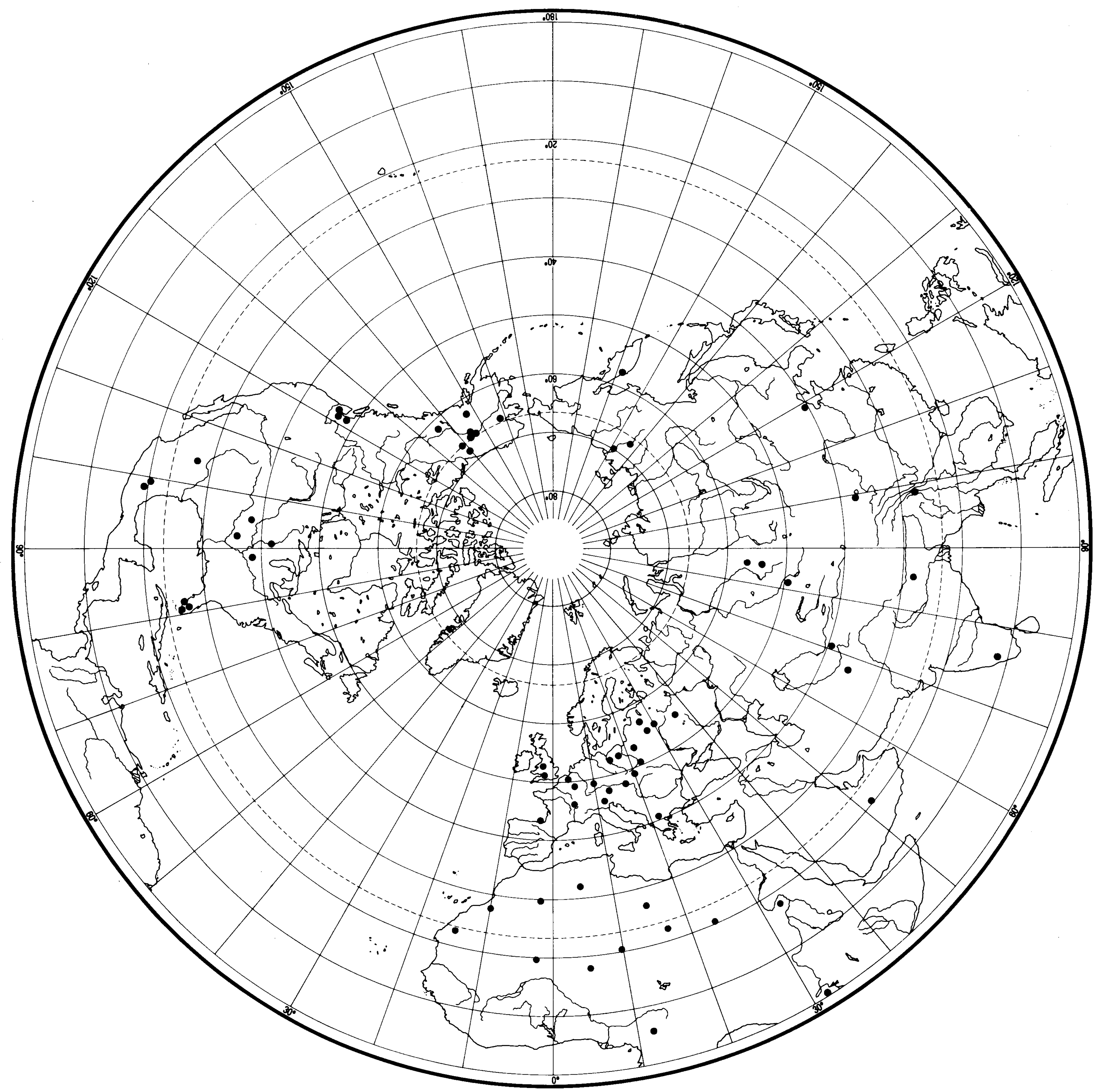

Fig. 4. Sites used for the construction of maps on pages 29, 31, 33 , 35 and 37

masses could form (for further information see FRENZEL, 1991b). At about 25,000 years B.P., the Central Alps were still free of ice. Thus the classical Alpine Weichsel or Würm glaciers could not have formed during the first part of the Last Glaciation, but only appeared by the end of this glacial epoch. On the other hand, the North American evolution of inland ice masses during the Last Glaciation is a very extended and complex one. The same seems to be valid for the **ice sheet formation** and **mountain glaciations** in Siberia (e.g., ALEKSEEV, 1978; KIND, 1974; KIND and LEONOV, 1982; ARKHIPOV *et al.*, 1986). In this context it is assumed that the inland ice formation causing the Late Wisconsin Glaciation, or the German Jungmoränen occurred between about 35,000 and 25,000 years B.P.

Evidence suggests that between 35,000 and 25,000 years B.P., climate had changed at least twice. Thus the dates mentioned do not encompass a stage with a homogeneous climate. On the other hand the "absolute" datings of this period are far from exact. This means that to restrict oneself to a shorter interval within the 10,000 year period mentioned does not increase the accuracy of evaluations. Thus a broader interval was chosen, yet it was attempted to concentrate on the younger phase with a somewhat milder climate within this 10,000 year interval wherever possible.

From the fossil material available one gets the impression that at that time climates of different regions varied from each other. In some regions ice sheets which had already formed seem to have grown steadily, as e.g., in Scandinavia (MANGERUD, 1991), whereas elsewhere the former glaciers and inland ice masses may have disappeared between 35,000 and 25,000 years B.P., like in some regions of Siberia or presumably in the Alps (e.g., KIND and LEONOV, 1982; KELLER and KRAYSS, 1991; SCHLÜCHTER, 1991). The climatic reconstructions of this period for the Northern Hemisphere will contribute to a better understanding of what might have happened in various regions of the world at that time (see the maps mentioned above).

On the other hand, KUHLE (1987a, b; 1988; 1991) has repeatedly stressed that during the first part of the Last Glaciation a huge inland ice mass formed on the Tibet Plateau, the surface area of which is calculated to have been approximately 2,400,000 km^2. In his opinion these vast ice masses, the albedo of which might have been approximately 80 to 90 per cent, triggered the formation of inland ice in North America and in Scandinavia *(Fig. 5)*. Yet LI *et al.* (1989) claim that according to their geomorphological, geological and geochronological research carried out along the southern border of the Western Kunlun Mountains, the western Tibet Plateau experienced two phases with the formation of huge lakes during the Last Glaciation, between more than 47,000 and 35,000 years B.P., and between 21,000 and 15,000 years B.P. The same is valid for Central Tibet, between the Kunlun Pass and the mountain systems of the northern Himalayas, to the southwest of Lhasa, according to our investigations on fossil lake terraces, surrounding extant lakes on the Tibet Plateau (several thermoluminescence dates between 104,000 ±7,000 and 12,000 ±1,000 years B.P., FRENZEL *et al.*, 1991).

Thus it is suggested that the Tibet inland ice sheet did not exist during the last glaciation and accordingly only a minor glaciation within the higher Tibet mountain systems is indicated in maps on climates during inland ice formation. Nevertheless, the stimulating papers of KUHLE merit serious study and further testing.

Winter temperatures (map on page 29). The data in this map are given as deviations from present-day conditions. It is suggested that in general the coldest month was February, though

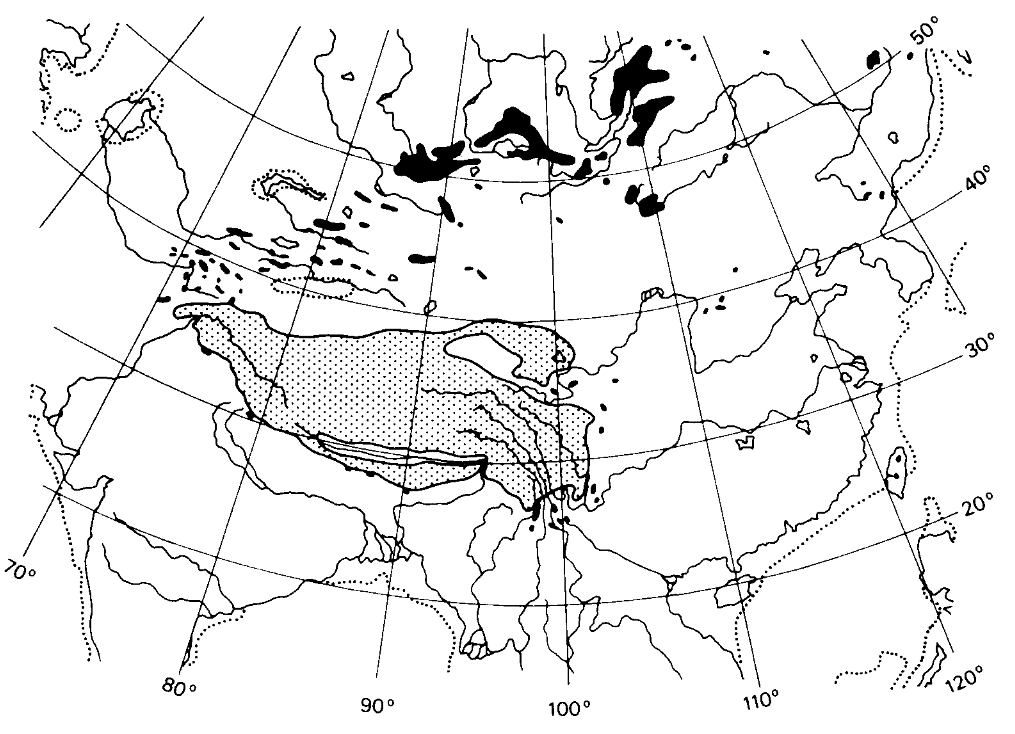

Fig. 5. The Tibet inland ice, according to M. KUHLE (1987a, b; 1988; 1991)

it may be that similar values prevailed in January, at least in some regions of the Northern Hemisphere.

The maps, as a rule, were compiled according to the distribution patterns of various plant and animal taxa and to the types of the contemporaneous vegetation, which were reconstructed from fossil records. Another climatic indicator used was paleosol development during the intervals of interest.

Though traces of syngenetic permafrost are quite uncommon for the period concerned, all indicators of climate show that within the belt of present cold temperate climates, winter temperatures were much lower than today. On the other hand, a steep climatological gradient seems to have existed in Europe and North America, which indicates that inland ice already existed at that time, even if this gradient was not as steep as in full-glacial times. It may be concluded that the ice masses, particularly in North America, were much less extensive than during the last glacial maximum (for further information see DREIMANIS, 1991; FULLERTON, 1986a; RICHMOND and FULLERTON, 1986).

It is evident that tropical regions were nearly as warm as today. This reveals that cooling was concentrated on North America, Greenland, the North Atlantic and northwestern Europe. The contemporaneous situation in northern Siberia seems problematic. Here sediments of the Karginsky Interglacial (or Interstadial) are worth mentioning (as to the history of research: KIND and LEONOV, 1982). Some authors claim that at that time climate in northern Siberia was at least as warm as today or even milder. This is shown by — in general — infinite radiocarbon ages, by former distribution patterns of terrestrial plant species and communities, and by marine molluscs and foraminifera. Yet, the marine faunas of the Kazantsevo Interglacial (equivalent to the Eem Interglacial) and of the **Karginsky** Interglacial (or **Interstadial**) of the Taimyr Peninsula closely resemble one another (cf. ANDREEVA, 1982; ANDREEVA and KIND, 1982). Thus it is assumed that both these transgressions of the Polar seas are either of the same age, or that the Karginsky Transgression dates from the early warm phases at the very beginning of the Last Glaciation, but not from the interval 35,000 to 25,000 years B.P.

The extremely steep gradient in mean air temperatures of the coldest month along the coast of Western Canada points to a striking thermal contrast between the ocean and the continent at that time. This would have induced abundant precipitation in the mountains, which should have triggered glaciation, see maps on pages 35 and 37.

Summer temperatures (map on page 31). When attempting to investigate the climatological conditions responsible for the formation of the Late Wisconsin, classical Weichsel, or Upper Valdai glaciations, the time period between 35,000 and 25,000 years B.P. was chosen, as in the other maps on climates of inland ice formation. It is suggested that August was in general the warmest month, though occasionally the highest temperatures may have occurred in July.

At present, no data are available for sea surface temperatures between 35,000 and 25,000 years ago. This is regrettable, because during this interval, decisive climatological processes are assumed to have started, which eventually led to extensive glaciation.

Comparing maps on pages 29 and 31, it becomes clear that during the summer, a strong cooling of air masses in the North Atlantic Ocean region was typical, but this cooling was much more intensive in North America and extended much farther south than in Europe. From this it may be concluded that relatively larger ice masses already existed in North America, while they only began to form or were very restricted in northwestern Europe. However, owing to the configuration of the isotherms in northwestern Central and Northern Europe, it becomes evident that in Northern Europe, some inland ice masses were already present (for more information see MANGERUD, 1991). Moreover, it can be assumed that in the oceans near the coasts of northeasternmost Siberia and northern Alaska a minor centre of cooling existed. According to map on page 31, its influence did not reach far south.

Similar to air temperatures of the coldest month, a steep gradient in mean summer air temperatures existed at that time between the northeastern Pacific Ocean and the neighbouring mountain systems of Western Canada. This is thought to have stimulated the accretion of ice masses there, even during the summer.

As with air temperatures of the coldest month in the tropical and subtropical regions, mean summer air temperatures seem to have been identical to those of today.

Comparing the maps for the air temperatures of the coldest and of the warmest months with one another, it becomes evident that the centre of cooling, which initiated inland glaciation in the Northern Hemisphere, was situated in the northern parts of North America.

Annual mean temperatures (map on page 33). Again data are given as deviations from present-day indexes. The basic data for constructing this map are not obtained by interpolating between the mean annual air temperatures of the coldest and of the warmest months, but other indicators of climate were used instead.

Unlike present-day cold temperate and cold climatic belts, the subtropical and tropical regions did not experience any pronounced cooling at that time, as it is apparent in the previous maps, see pages 29 and 31.

The deviations of annual mean air temperatures from those of today show three major areas of pronounced cooling, namely the northern parts of North America, northwestern Europe, and northeastern Siberia. Of these centres of cooling, climatic reconstructions for northeastern Siberia have proven to be the most problematic, though a distinct cooling during summer is already indicated in the map.

Maps on pages 29 and 33 taken together suggest deeply depressed air temperatures over the Arctic Ocean as well as over the North Atlantic, compared with present-day conditions. The same did not occur simultaneously over the North Pacific Ocean. From this it may be concluded that the Arctic and the North Atlantic oceans were covered by sea-ice, unlike the North Pacific. It may be that simultaneously the Sea of Okhotsk and Sea of Japan also experienced a long seasonal cover of sea-ice, but it has not yet been proven.

Annual mean precipitation (map on page 35). As already stated when discussing the precipitation during the climatic optimum of the Last Interglacial, it must be stressed again that from a paleoclimatological point of view it is impossible to reconstruct the precipitation rates of the past with high precision. What can be deduced, however, from the distribution patterns of sediments, soil types, plant and animal taxa or from vegetation types and geomorphological processes of the past, are water budgets. The latter are then compared with those of the present, where precipitation rates are known. To enable a comparison between present-day conditions and those of the past, the paleoecological situation between 35,000 and 25,000 years B.P. is converted into precipitation data in the map discussed. These data can only show the relevant trends in past water budgets. Moreover, it must be taken into consideration that there is a lack of information regarding seasonality of moisture supply in various regions of the world at that time.

Nevertheless, as can be seen from the map, a distinct difference existed between the amount of **moisture available** in subtropical and tropical regions and that in the cold-temperate and cold regions. Between approximately 35,000 and 25,000 years B.P., the tropical and subtropical regions experienced a remarkable increase in water supply compared with present-day conditions. If it is accepted that during this time interval the air temperatures of the coldest and of the warmest months were only lower by about 1 to 1.5 °C than today, this increased moisture supply can chiefly be explained by relatively high **evaporation rates** from the neighbouring ocean surfaces.

On the other hand, it is very unlikely that the strongly reduced moisture in the northern cold temperate and cold climate belts was only caused by a highly reduced rate of evaporation from the North Atlantic and the Arctic oceans. It seems more realistic to accept the hypothesis that extensive sea-ice cover had already appeared by that time. Although the same did not occur on the North Pacific Ocean, the influence of long-lasting seasonal sea-ice in the seas of Okhotsk and of Japan may be deduced from the map. As shown by coring, these masses of sea-ice in the North Atlantic Ocean can be traced in deep sea sediments.

The difference between the "annual mean precipitation rate" during inland ice formation and maximum glaciation (deviations from the pleniglacial values during Upper Weichsel or Late Wisçonsin time, map on page 37).The phases compared are those between 35,000 and 25,000 years B.P. and between about 20,000 to 18,000 years B.P. Positive values in the map indicate regions with higher "mean annual precipitation rate" at about 35,000 to 25,000 years B.P. than at about 20,000 to 18,000 years B.P., while negative values show the reverse.

What has already been said regarding the reliability and accuracy of reconstructions of mean rates of annual precipitation for episodes of the past, must be stressed again for this map. Here the difficulties are even greater due to the relatively poor accuracy of the available data. So again the general trends are emphasized, not the absolute values.

As seen in the map in concern, there was a more humid climate nearly everywhere during the time of inland ice formation than in full-glacial times. This was especially valid for tropical South America and Central America, for tropical and subtropical Africa, for Southeast Asia, for northwestern North America, and, to a lesser extent, also for Western and Southwestern Europe.

The situation in the tropical and subtropical regions can be explained by more abundant moisture supply in these climatic belts between 35,000 to 25,000 years B.P. than during full-glacial times, presumably owing to the still warm tropical and subtropical oceans, and by slightly reduced terrestrial evaporation rates compared to the present. This strong reduction in available moisture during the glacial maximum indicates that full-glacial atmospheric circulation patterns were completely different from those during the inception of this glaciation. High precipitation rates in Southeast Asia in comparison to full-glacial conditions may have initiated an early and extensive glaciation in southeast Central Asian mountain systems, though this does not necessarily prove the existence of inland ice in Tibet at that time, as was suggested by KUHLE (1987a, b; 1988; 1991) and by HÖVERMANN and SÜSSENBERGER (1986).

It seems realistic to suggest that the relatively high precipitation rates in the Mediterranean region and in the mountains of the Middle East were caused by an intensified latitudinal circulation there during inland ice formation, due to a marked longitudinal thermal gradient at that time (maps on pages 29, 31 and 33) and to the relatively humid air masses during inland ice formation, compared to full-glacial times.

As shown in maps referred, during the time of inland ice formation, particularly in Western Canada, an extremely steep temperature gradient existed between the North Pacific Ocean and the mountain systems of western North America. As can be seen in map on page 37, this situation also coincided with a difference in the oceanic moisture supply from the Pacific Ocean to westernmost North America during the time of inland ice formation, compared to the glacial maximum. This water supply from the Pacific Ocean is expected to have caused föhnwind systems on the leeward sides of the mountain ridges. Yet during full-glacial times the regional paleoecological situation was controlled by another climatological condition. Both these influences were likely responsible for the pattern shown in the map for the eastern part of the Rocky Mountains: it seems that in contrast to the föhnwind situation, which prevailed between 35,000 to 25,000 years B.P., the moisture supply was fairly abundant in the southwestern part of the present-day USA in full-glacial times, owing to depressed temperatures, which resulted in highly reduced evaporation, and also due to the extremely strong thermal gradient during the glacial maximum.

MAXIMUM COOLING OF THE LAST GLACIATION
(about 20,000 to 18,000 yr B.P.)

CLIMATES DURING THE LAST GLACIAL MAXIMUM

by B. Frenzel, in charge of the Research Project Group "Terrestrial Paleoclimatology", Federal Republic of Germany
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(Explanatory notes to maps on pages 39, 41, 43 and 45)

General remarks. It is generally held that the last glacial maximum dates at about 18,000 years B.P. This is supported by several age determinations in North America, Europe and Siberia, indicating that the maximal advance of the inland ice masses occurred later than 22,000 years B.P., yet before 17,000 years B.P. It must be noted that deviations from this general principle do exist, see map on page 49.

It may be questioned whether this glacial maximum coincided with climatological conditions typical of an intensive glacier advance. There exist observations, which indicate that the most appropriate climate for maximum expansion of ice sheets prevailed some time before the maximal extension of the inland ice masses occurred (see maps on pages 35 and 37; for further information: HAEBERLI, 1991; KELLER and KRAYSS, 1991). On the other hand, once the inland ice had reached its maximal extension climate seems to have already favoured its disintegration, rather than advance. Since **CLIMAP** (1981) chose 18,000 years B.P. for the reconstruction of sea surface temperatures, the same period is considered in the maps discussed below. It must be recognized, that it is impossible to provide exact datings everywhere for this event. Instead, the above maps only show the interval of the most extreme climatic conditions, at about 20,000 to 18,000 years B.P. *(Fig. 6).*

In map on page 39, sea surface temperatures and the southern limits of compact sea-ice are given for the coldest month of maximum glaciation, according to **CLIMAP** (1981).

At about 18,000 years B.P. global sea levels had dropped to about 100 to 120 m below the present. It is also held that the Upper Khvalyn Transgression of the Caspian Sea broadly coincided with this time (FEDOROV, 1978; VARUSHCHENKO *et al.*, 1980). So did presumably the maximal transgression of the Aral Sea during the Upper Pleistocene. Recently new data have been obtained which suggest different chronological position for the Late Khvalyn Transgression and correlate it with the time of the melting of Late Valdai glaciers after 15,000 years B.P. (ARSLANOV *et al.*, 1988).

Methods used. When compiling the maps in concern, the main indicators of climate were former distribution patterns of various plant and animal taxa, and of certain plant communities. As to the sites used, see *Fig. 6*. Yet the plant communities only resemble some recent ecosystems and this lack of correspondence is one of the main shortcomings for the quantification of full-glacial climates, rendering exact comparisons by mathematical procedures between present-day and past climates nearly impossible. Although useful mathematical methods for the quantitative evaluation of full-glacial climates have been worked out (GUIOT *et al.*, 1989) they were not used in our reconstructions. Instead it was attempted to incorporate as many different indicators of former climates as possible, provided they enable quantitative estimates to be made. Preferred indicators were those which are independent of one another and datable. Besides the biological indicators, distribution of relict permafrost features and pediments, stable isotopes and noble gases in ground waters, lake-level oscillations in closed basins, loesses and other eolian sediments were used. Each of these parameters was evaluated regarding its reliability as an indicator of climate. Then all the useful indicators within a region were combined, enabling a better comparison with present climates. Here again, the values mapped were not calculated as an arithmetic mean of the figures derived from various indicators of climate, but by a comprehensive paleoecological analysis of the meaning and weight of each indicator of past climates. Only the general trends were incorporated, not the extreme values.

Winter temperatures (map on page 39). In tropical and subtropical regions, climate of the coldest month (presumably February) was about 10 to 12 °C colder than today. This was the background situation, with greater extremes in most of North America, Europe (except its southernmost parts), the Tibet Plateau and Northwestern China. The extremely cold areas were situated in North America and in Europe. A steep thermal gradient led from the tropical and subtropical regions to those of greatest cold. This must have provoked an appreciable intensification of atmospheric circulation beyond the inland ice masses, compared to present conditions. The strong upwelling phenomenon off the West African coast (e.g., SARNTHEIN *et al.*, 1987) is claimed to have been caused by these strong winds.

During the inception of inland ice formation (see map) the most drastic cooling occurred in the northern part of North America. Yet map on page 39 shows that during the maximum of the Last Glaciation the lowest temperatures (compared to present conditions) occured in Northern Europe. The Siberian center of cooling (during the period of inland ice formation) had evidently not experienced a drop of temperatures as great as the other two centres.

It must be noted that no information is available about the actual temperature values over the inland ice surfaces.

If it is correct that an intense deterioration of winter climates in North America and in Europe was caused primarily by the influence of the inland ice masses, it may be deduced from our map that this impact extended about 1,100 km beyond the south-

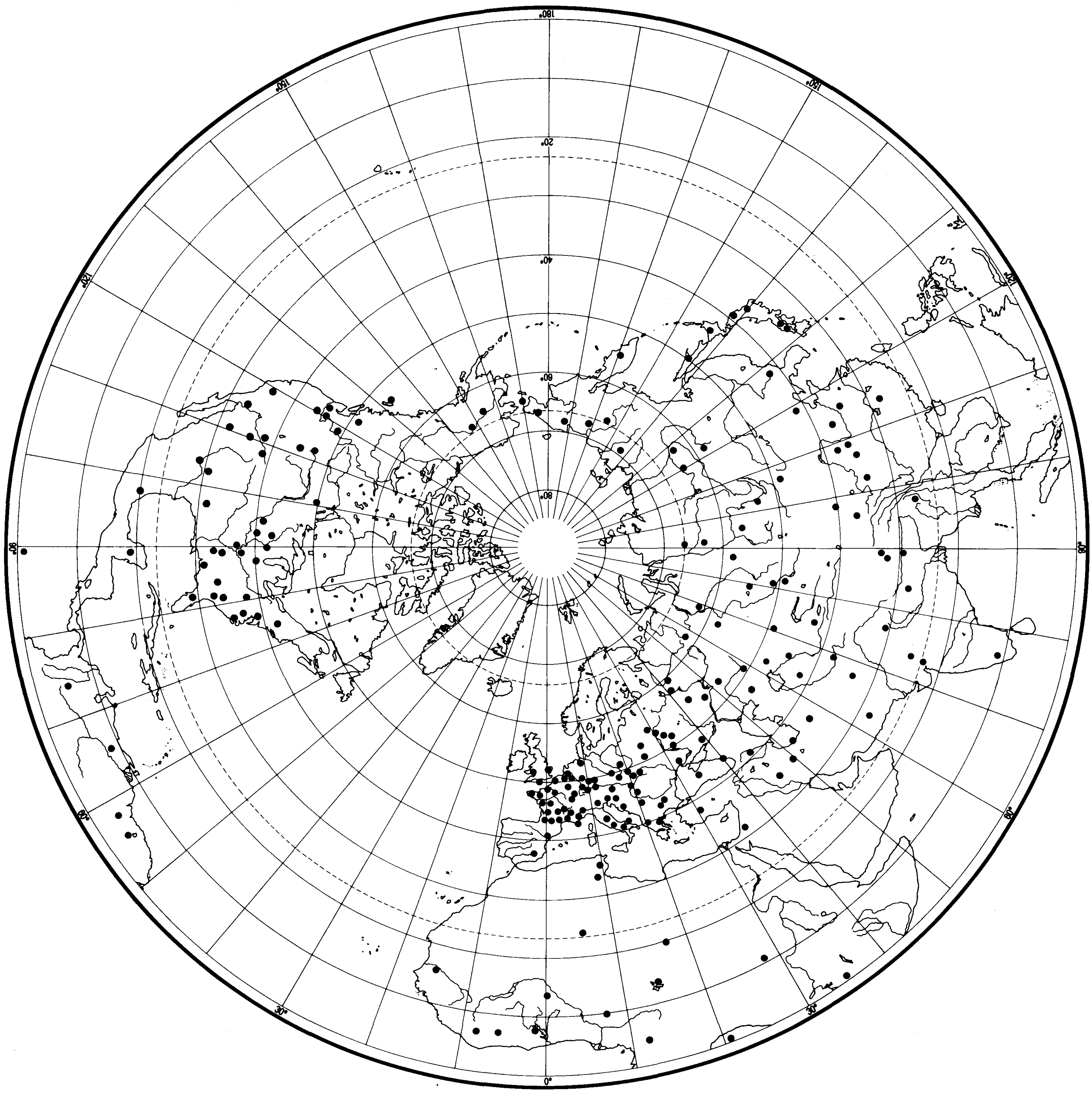

Fig. 6. Sites used for the construction of maps on pages 39, 41, 43 and 45

ern limits of the ice sheet in North America, and some 600 to 700 km in Europe. This may have been due to the Alps and Pyrenees, which blocked the southerly flow of cold air in Europe, and restricted the glacial influence to ca. 600 km here, compared with the lowlands of Eastern Europe where glacial conditions extended about 700 km south.

The configuration of the isolines of temperature deviations in southwestern France shows an extremely weak influence of oceanic air masses in Europe at that time. The same also seems to have been valid for North America (see the southern part of North America; the other flanks of the continent were sheltered by mountain systems).

Summer temperatures (map on page 41). It is suggested that during the time in concern, the warmest month was generally August, although in some regions of the Northern Hemisphere July must be taken into account as well.

Sea surface temperatures and compact sea-ice cover were taken from **CLIMAP** (1981).

Remarkably, the deviations of mean summer air temperatures from those of today were everywhere much smaller than during the coldest months. In general, summer climates seem to have been colder in tropical and subtropical regions by about 6 °C in comparison to present-day conditions, and vast areas of Northern Asia had only 6 to 8 °C colder summers than today.

Again the centres of cooling were situated in northern North America and Europe, with the exception of its southernmost parts. It is worthwhile to note that the differences in the intensity of cooling between North America and Europe during the winter could not be observed for the summers of about 18,000 years B.P. Moreover, the cooling influence over the Tibet Plateau was very faint.

This general lack of significant cooling is assumed to have maintained a relatively high evaporation rate during summer at about 18,000 years B.P. Thus the moisture supply of the extraglacial ecosystems should have dropped considerably during summers. The extremely sparse vegetation cover in the Northern Hemisphere was partly due to this phenomenon of aridity, not merely to the decrease in temperatures.

In the southeastern part of North America, interesting paleobotanical investigations were carried out by P.A. DELCOURT (1980) and H.R. DELCOURT (1983). They reconstructed mixed broad-leaved and coniferous forests during glacial time. This is in sharp contrast to the reconstructions of winter and summer temperatures for the south- eastern part of North America given in maps on pages 39 and 41. Yet the pollen floras analysed were found only in minerogenic sediments and the number of pollen grains analysed or their concentrations are extremely low. Moreover, the pollen floras in general change abruptly with the beginning of peat formation and the widely spread loesses do not provide evidence for the existence of thermophilous and hygrophilous forests within the southeastern USA. This is the reason why it is thought that these most interesting investigations cannot be referred to full-glacial sediments proper. Instead, it is suggested that the temperate pollen floras were redeposited from older sediments, perhaps of interglacial or interstadial origin, and that long hiatuses strongly impede a correct interpretation (for further information see WATTS, 1970; 1971; 1980; 1983).

Annual mean temperatures (map on page 43). Just as was stated for maps on pages 23 and 33, this map is not drawn by interpolating between the values for the deviations of temperatures of the coldest and of the warmest months from present-day conditions, but is based on specific climate indicators. The most important ones are vegetation types having then existed.

The general background situation seems to have been a drop of annual mean air temperatures by about 8 °C at approximately 18,000 years B.P., compared to the present. CLIMAP (1981) did not reconstruct mean annual sea surface temperatures, so only the continental data are given in map on page 43.

Again three centres of maximum cooling are clearly discernible: North America, most parts of Europe, and Central Asia. As was mentioned in the explanations to map on page 41, the influence of the Tibet Plateau was not as pronounced as that of the North American or the European ice masses.

It is reasonable to suggest that the influence of these ice masses on annual mean air temperatures was weaker than during the winter season. Nevertheless, the sheltering effect of the Pyrenees, Alps and Carpathians is clearly manifest, since the cooling influences of the Scandinavian ice masses on annual mean air temperatures extended much farther in the southeast than directly south. This influence extended approximately as far as the Ural Mountains, suggesting that the area covered by glacier ice in Siberia could not have been very large. In general, this is corroborated by geological and geomorphological observations in Siberia (e.g., ALEKSEEV, 1978; KIND and LEONOV, 1982; ARKHIPOV *et al.*, 1986, VELICHKO *et al.*, 1989; see also map on page 49), though there are divergent views on the extent of former glaciation in the northern part of Western Siberia and in the northwestern part of Central Siberia (GROSVALD, 1983).

As already stated regarding the air temperatures of the coldest months, the warming influence of the Atlantic and Pacific oceans on the climates of North America and of Europe was negligible.

Annual mean precipitation (map on page 45). The climatic indicators on which this map is based are chiefly the distribution patterns of contemporaneous vegetation types, of certain animal taxa, of loesses and loess-like sediments, and of former pediments, provided these pediments could be dated. Stable isotopes of ground waters were also used. To some extent the position of the equilibrium line on former mountain glaciers and its depression compared to its present-day altitude were evaluated as an indicator of humidity or aridity (v. WISSMANN, 1959; FRENZEL, 1960; MESSERLI, 1967; KUHLE, 1987b).

As to the problem of the notion "precipitation rate" see explanations to maps on pages 27 and 35.

Though on the continents the amount of moisture available was appreciably reduced at about 18,000 years B.P. compared with present-day conditions, there did exist two areas of deviation from this general pattern; Central Asia and the southwestern part of North America. According to present knowledge, in the regions mentioned the amount of moisture available was either similar to today or even greater. It is suggested that this was caused by a radical decrease in potential evaporation in Central Asia, due to the significant fall of temperatures. The latter, having occurred mostly during the summer season, was more marked in Central Asia and in North America than in North Africa. Moreover, the Pacific storm tracks had shifted at that time far to the south, thus bringing southwestern North America more moisture. On the other hand, the area with the greatest increase in humidity in Central Asia was the Tibet Plateau and the adjacent mountain systems of the Tien Shan. This might have initiated glaciation, though from the data used for maps on pages 29 and 39, the former existence of inland ice in Tibet could not be established (cf. explanation to maps on climates during inland ice formation, general remarks).

The areas with maximum decreases in available moisture were the mountain systems of southern Europe, of the Middle East, the Western Ghats, the Himalayas and the adjoining western Burmese Mountains, together with Abyssinia, the Alleghenies and the present-day humid inner tropic areas. This indicates a widespread decrease in precipitation at that time. Since the decrease in summer temperatures was not so marked as it was during the winter at about 18,000 years B.P., this in addition should have increased evaporation rates during the summers, causing the water balance to become critical in vast regions of the globe. It is suggested that this must have influenced the CO_2 turnover by slowing down global fermentation and respiration in soils. It is assumed that a **strongly reduced input of CO_2 into the atmosphere** would have been the result (see explanations to map on vegetation, page 55).

Map on page 45 indicates a decrease in precipitation by about 100 mm per year for the Central and Northern Sahara. Though a wealth of information on these regions points out extreme aridity of these deserts, which was even more pronounced at that time than today, the data on the negative deviation of about 100 mm are of course only relevant for areas in which the annual precipitation is presently higher.

No data are available on the interannual variability of precipitation at about 18,000 years B.P., although a wide range is likely.

From the map it becomes understandable why, in the present-day cold temperate and boreal climatic belts, no or very few traces of human hunting activities are found in the archaeological record of the glacial maximum.

THE SUMMER SURFACE ALBEDO AT ABOUT 18,000 B.P.

by B. Frenzel, in charge of the Research Project Group "Terrestrial Paleoclimatology", Federal Republic of Germany
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(Explanatory notes to map on page 47)

General remarks and methods. The albedo is one of the most important variables in the equation of the Earth's radiation budget, although even today it is very difficult to determine the albedo for various types of land surfaces, clouds or sea surfaces. Any attempt to reconstruct the past albedo of the globe is thus prone to serious misinterpretations. There do not exist any methods to reconstruct directly the planetary albedo of the past, since there is a lack of knowledge about the past cloudiness of the sky, types of clouds, their seasonal occurrence, and so on. We attempted to approximately reconstruct only the albedo over the vegetated land surfaces. Yet since the areas covered by snow during winter in the past are unknown, one is restricted to reconstructing the **summer surface albedo**, such as for 18,000 years B.P., as shown in the map in concern. The data used were obtained by reconstructing dominant **vegetation types** as exactly as possible, i.e., more accurately than for map on page 55, and to apply values for surface albedos of the present comparable vegetation types *(Table 1)*. By subtracting the values of the reconstructed surface albedo at 18,000 years B.P. from those of today, the deviations between the glacial maximum and present-day can be obtained from a network of many reference sites.

Table 1. Percentage solar radiation reflected by different surfaces (albedos by LAMB, 1972; GEIGER, 1961)

Surface	Albedo%	Surface	Albedo%
Newly-fallen snow	85-95	Clean glacier-ice	30-46
Old snow	50-70	Contaminated glacier-ice	20-30
Thawing snow	30-65		
Dune sand	34-60	Clayey desert	29-31
Sandy soil	14-40	Salt deposits from dried-up lakes	50
Fir tree canopy	10	Spruce tree canopy	14
Pine forest	6-19	Oak tree canopy	18
Green deciduous forest	16-27	Yellow deciduous forest in autumn	33-38
Meadow	12-30	Green grass	8-27
Parched grassland	16-30	Stubble field	15-17

It was already stated that atmospheric circulation is suggested to have intensified considerably during full-glacial times in comparison to the present. This would mean an increase in the roughness of the ocean surface, in turn strengthening reflectivity of the ocean surfaces. Since this phenomenon has not been sufficiently studied, only approximate indexes of the summer surface albedo of the ocean could be evaluated.

Some years ago **CLIMAP** (1981) published a map showing the summer surface albedo for full-glacial times of the Last Glaciation. Since much more is known today about the contemporaneous terrestrial vegetation over the Northern Hemisphere, map appearing on page 47 was compiled anew.

Results. Our map shows that at about 18,000 years B.P. the summer surface albedo increased in areas where interglacial or present-day forests were replaced by various types of steppe or semidesert vegetation. This represents a considerable increase in reflectivity, most of all in recent cool temperate to northern boreal climatic belts. New-fallen snow increased the summer surface albedo in the centres of inland ice accumulation.

On the other hand, the present-day tundra belt, as well as the steppe and desert belts had not experienced any pronounced changes in their surface albedos, with the exception of those smaller areas, where a present-day steppe or desert vegetation had been replaced by bushes or forests such as in the southwestern part of North America. In eastern Kazakhstan and Turkestan lower values of albedo can be attributed to a much denser steppe vegetation than today, whereas in northeasternmost Siberia the eustatic regression of the Arctic Ocean enabled tundra vegetation to thrive on the former shelf, thus decreasing the surface albedo.

These changes in the surface albedo during the summer season are suggested to have caused an overall decrease in air temperatures, which in turn enhanced the deterioration of climate in areas adjacent to the inland ice sheet. This influence is suggested to have been even more pronounced during winters at about 18,000 years B.P., due to the extensive land surfaces covered by thin layers of snow and firn.

All the foregoing interpretations suggest an increased stability of climate and a constancy in the distribution patterns of the huge ice masses during the glacial maximum. Yet just the opposite happened. Immediately after 18,000 years B.P. a rapid **disintegration** and retreat **of the ice masses** began. It is suggested that this process was triggered by the sublimation of ice under an extremely dry climate, by the dust storms contaminating ice surfaces and by an increase in solar radiation, due to changing orbital parameters of the Earth.

CORRELATION OF THE LATE PLEISTOCENE EVENTS WITHIN GLACIATED AREAS OF THE NORTHERN HEMISPHERE

by A. A. Velichko

General remarks. Geological, geomorphological, paleontological, physico-chemical and radiometric data have considerably clarified the history of the Late Pleistocene Glaciation. These data are especially important for our understanding of the complicated paleogeographic problems of the Northern Hemisphere where vast ice sheets covered large regions of the middle and high latitudes. Work conducted by the **International Geological Correlation Programme (IGCP), project "Quaternary Glaciation in the Northern Hemisphere"** made a significant contribution to our understanding of past glacial events. This work, in particular, permitted us to undertake the reconstruction of past glacial systems presented in this atlas.

The development of glacial phenomena varied from region to region in the Northern Hemisphere. This occurred because of the variability in climatic and geographic factors, such as the ratios of sea to land, plains to mountains, cold to warm sea currents, the position of the regions vis-à-vis the systems of atmospheric circulation, etc. This section discusses our present state of knowledge about the Late Pleistocene events within the main glaciated regions of the Northern Hemisphere *(Fig. 7)*.

Problems encountered in the interregional correlation of the ice sheets dynamics are associated with an unequal state of knowledge concerning glacial chronology in regions which accounts for different "infillements" of regional schemes. A number of difficulties arises from various interpretations of the glacial history within the same regions (e.g. in North America or Scandinavia). Considering the aims of the atlas, only principal (sometimes generalized) elements of regional chronologies are presented in *Figure 7* and in the *Tables 2* to *5*.

North America. The history of North American glaciation represents an unique case. Here the largest ice sheet developed on the plain and coalesced with the complex glacial body of the **Cordilleran ice sheet**. They formed a single glacial system which spread from the Atlantic to the Pacific coasts over the northern and middle sections of the continent. This process has been extensively studied over the years by a number of scholars beginning with FLINT (1971) and followed by work of numerous research teams (WRIGHT, 1983; ŠIBRAVA *et al.*, 1986). In the 1960s American and Canadian specialists developed a chronology of glacial events which, for the most part, remains valid to this day. Key sections north of Ontario, Canada, for example, reviewed by KARROW (1984), were first used by DREIMANIS and KARROW (1972) as a basis for the chronostratigraphy of North America *(Table 2)*.

Data obtained recently permit us to clarify the sequence of glacial events during the **Wisconsin ice age**. Specifically, the "classical" tripartite subdivision into the Early, Middle, and Late Wisconsin has been replaced by a four-fold scheme which includes a new subdivision of the Eowisconsin to designate a time period between the Sangamon and the Early Wisconsin Glacial Stage (ŠIBRAVA *et al.*, 1986).

The **Sangamon Interglacial** corresponds to Oxygen Isotope Substage 5e of the marine oxygen isotope curve, the **Eowisconsin** correlates with Substages 5d, c, b and the **Early Wisconsin** with Stage 4, the **Middle Wisconsin** with Stage 3 and beginning of Stage 2, finally, the **Late Wisconsin** with the remainder of Stage 2 and the beginning of Stage 1.

The glacial history of the *Great Lakes region* is the best known of all the areas covered by the **Wisconsin ice sheet**. ESCHMAN and MICHELSON (1986) argue that no evidence for

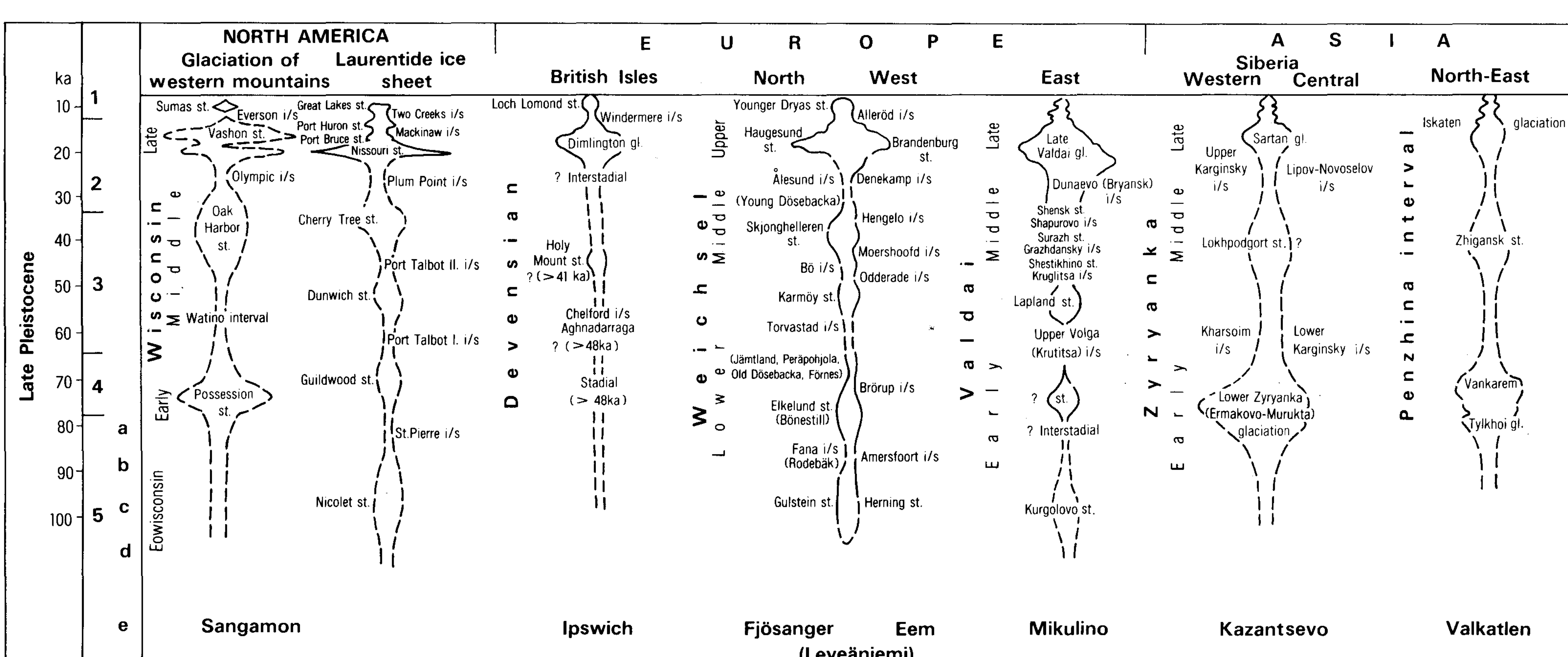

Fig. 7. Correlation of the Late Pleistocene events within glaciated areas of the Northern Hemisphere (by A. A. VELICHKO)

Table 2. Late Pleistocene glacial chronostratigraphy of North America

Late Wisconsin	North Bay Interstadial Great Lakes Stadial Two Creeks Interstadial (11,850 years B.P.) Port Huron Stadial Mackinaw Interstadial (13,000 years B.P.) Port Bruce Stadial Erie Interstadial (16,000 years B.P.) Nissouri Stadial
Middle Wisconsin	Plum Point Interstadial (28,000 years B.P.) Cherry Tree Stadial Port Talbot II Interstadial (45,000 years B.P.) Stadial (Dunwich Drift) Port Talbot I Interstadial (65,000-50,000 years B.P.)
Early Wisconsin	Guildwood Stadial St.-Pierre Interstadial (75,000 years B.P.) Nicolet Stadial
Eowisconsin	
Sangamon Interglacial	

the *Eowisconsin* glaciation has been found in Michigan and Wisconsin. Within the **Huron glacial lobe**, a till is tentatively dated to the second half of the Middle Wisconsin. This till is stratified between organic beds, with the upper horizon radiocarbon dating to between 25,000 and 22,000 years B.P. (probably reflecting the *Plum Point — Farmdale* Interstadial), and the lower organic horizon dating older than 30,000 years B.P.

Glacial deposits formed during the *Nissouri* Stadial, representing the maximum Late Wisconsin advance, dated to ca. 21,000 to 20,000 years B.P., are the most extensive (**Sanilac till, Tiskilwa, New Berlin**, and the lower members of the **Wedron formation**). The *Erie* Interstadial (16,500 to 15,500 years B.P.) and the later recessional *Port Bruce* Stade (15,500-14,000 years B.P.) are represented by the **Fisher Road** and **Wadsworth tills**. The *Mackinaw* Interstadial (ca. 14,000-13,000 years B.P.) was followed by the *Port Huron* Stadial (13,000-12,000 years B.P.), which is represented by various tills belonging to the **Jeddo till** and the **Kewaunee** and the **Copper Falls formations**. The short *Two Creeks* Interstadial (12,000-11,750 years B.P.), recorded in the so called "forest beds", was followed by a new readvance called the *Great Lakes* phase.

Research in the *area between the Great Lakes and the Atlantic coast* also yielded no evidence for the Eowisconsin Glaciation (FULLERTON, 1986b). Here the Early Wisconsin is represented by a series of tills (e.g., the **Titusville till**) interstratified with the **New Paris interstadial deposits**. The **Sidney** and **Farmdale soils** (dating between 65,000 and 23,000 years B.P.) separate the Early Wisconsin tills from the Late Wisconsin glacial formations. The youngest of the latter date to ca. 12,000 years B.P. Interstadial deposits, presumably corresponding to the *Erie* and *Mackinaw* interstadials, have been identified on the Appalachian Plateau as well as in the regions west of it.

STONE and BORNS, Jr. (1986) delimited two main glacial stages in New England. They attribute these to the Early Wisconsin, ca. 80,000 to 70,000 years B.P. (the **Drumlin till**) and to the Late Wisconsin, ca. 21,000 years B.P. and younger (the **Lexington till, Backersville till**, etc.).

Three till horizons have been described in the central part of the **Laurentide ice sheet** (Hudson Bay). The lower (**Adam till**) presumably dates to the Early Wisconsin, while the upper (**Kipling till**) to the Late Wisconsin.

ANDREWS *et al.* (1984) have noted the asynchrony in the dynamics of the ice spread from the different centers — the **McClintock dome, Keewatin, Foxe**, and the **Labrador ice divides**. On Baffin Island, the Wisconsin correlates with the *Foxe* Glaciation and includes the oldest *Ayr Lake* Stade (ca. 100,000 years B.P.), the *Quajon* Interstadial (ca. 80,000 years B.P.), the middle stadial, the *Iron Strand* Interstadial (event?) (ca. 50,000-25,000 years B.P.), and the last *Baffin Land* Stade (20,000 to 10,000 years B.P.).

The *western part of North America* is especially interesting in as much as this zone witnessed the interaction of the ice sheet with the mountain glaciers. FENTON (1984) identified three main stages of the Wisconsin on the Canadian Prairies. The *Burke Lake* Glaciation, whose lower age remains undated but whose upper limit is placed at ca. 65,000 years B.P., correlates with the Early Wisconsin. The Early and the Late Wisconsin were separated by the *Watino* interval which featured a cool temperate climate. FENTON (1984) dated this interval between 52,000 years B.P. and 23,700 years B.P. This was followed by the spread of the **Lostwood glacier** during the Late Wisconsin. This spread reached its maximum extent at ca. 20,000 years B.P. when the glacier probably coalesced with the **Cordilleran ice sheet**.

The dynamics of the glaciation differed from region to region in the western mountains. Two main phases of the Wisconsin glaciation have been distinguished in *Alaska* (Brooks Range and St. Elias Mountains; HAMILTON, 1986). During the Late Pleistocene this was probably the locus of the earliest glaciation. Volcanic ash from the **Siruk Creek** and **Itkillik glacial deposits** have been dated by fission tracks to 120,000 years B.P. The Early Wisconsin phase of active glaciation lasted to ca. 65,000 years B.P. but did include a recession interval at ca. 80,000 years B.P. During the end of the Early and throughout the Middle Wisconsin non-glacial environments predominated, with small glaciers probably present in the Brooks Range. A series of radiocarbon dates indicate that while this non-glacial interval ended here by 25,000-24,000 years B.P. The rapid decay of the glaciers began ca. 14,000-13,000 years B.P.

In the North (Canadian) Cordillera two main glacial phases were recognized, too (FULTON, 1984). DENTON and STUIVER designated them as the *Icefield* (Early Wisconsin) and the *Cluane* (Late Wisconsin) glaciations which were separated by the *Boutelier* non-glacial interval (Middle Wisconsin). Due to the nonsynchrony in the advances of the ice sheet and mountain glaciers, from time to time there existed an ice-free corridor. In southern Alberta the Cordilleran ice reached its maximum extent during the Early Wisconsin (*Waterton II*) while the Laurentide ice sheet reached its maximal spread later (*Erratics Train* advance), during the time when the mountain glaciers were already retreating. Here both ice margins were separated by an ice-free belt of a considerable width throughout the entire Late Wisconsin (RUTTER, 1984).

Studies of mountain glaciers and their correlation with those on the Puget Lowland show that several glacial phases existed in Washington during the Wisconsin Glaciation (EASTERBROOK, 1986). The *Possession* advance occurred during the Early Wisconsin (before 50,000 years B.P.) while the subsequent *Oak Harbor* one, marked by the **Bogachiel Drift**, ended ca. 30,000 years B.P. (*Salmon Springs* phase). This was followed by the *Olympic* Interstadial and by the *Vashon* advance which began at ca. 21,000 years B.P. A rapid recession occurred from 13,000 years B.P. onward which was accompanied by a rise in the ocean level and by glacio-marine sedimentation (**Everson Drift**). Between 12,000 years B.P. and the Holocene there occurred a brief *Sumas* readvance.

In the Great Basin and Colorado mountain systems, south of the Cordilleran ice sheet, distinct signs of glaciation date primarily to the Late Wisconsin. In some regions, however, such as in the Yellowstone National Park and in the adjacent regions of Idaho, Montana, and Wyoming, data on hand appear to also indicate ice advances during the Eowisconsin, at ca. 110,000-90,000 years B.P., and again during the Middle Wisconsin, between 65,000-50,000 years B.P. Here the Late Wisconsin Glaciation reached its maximum extent ca. 20,000-18,000 years B.P. (**Pinedale 1 till**). This was followed by three recessional phases (**Pinedale tills 2, 3, and 4**) during the Wisconsin, and by three more phases during the Holocene (RICHMOND, 1986).

In the Sierra Nevada mountains two main ice advances — the *Tahoe* and the *Tioga* — are assigned to the Early and Late Wisconsin, respectively. The Middle Wisconsin here was non-glacial (FULLERTON, 1986a).

Northern, Western, and Central Europe. Here the Late Pleistocene Glaciation was distributed very irregularly. In addition to those on the islands in the Arctic Ocean, there were two main ice sheets in the lower and middle latitudes — centred on the British Isles and on Scandinavia. The later was considerably greater in extent.

Using data from the marginal glacial zones of Denmark, ANDERSEN (1980), BERTHELSEN (1979) and SJÖRRING (1977) developed a rather extensive chronostratigraphic scheme for the Late Pleistocene Glacial epoch — particularly for the early and the middle phases. Their data indicate no glacial deposits dating to the **Lower** and **Middle Weichsel** cold phases (stadials)which followed the **Eem Interglacial**. Two **Upper Weichsel** ice advances (the *Norwegian* and the *Old Baltic*) occurred between 25,000 and 20,000 years B.P. The *Ne* maximum advance began ca. 20,000 years B.P. and was followed by the *Asnaes* Interstadial, dating 15,000-14,000 years B.P. At ca. 13,000 years B.P. this recession was interrupted by a short advance (the *Late Baltic*) *(Table 3)*.

The two Lower Weichsel interstadials delimited in the table below probably correspond to the *Amersfoort* and *Brörup*.

The *western flank of the* **Scandinavian ice sheet** covered southwestern Scandinavia (Norway). MANGERUD (1981) outlined the following dynamics sequence for the ice sheet here. The *Fjösanger* Interglacial was followed by the initial glacial phase (*Gulstein* Stadial). Later the *Fana* Interstadial occurred, which may correspond to *Rodebäk*. The Early Weichsel Glaciation reached its maximum extent during the *Elkelund* Stadial (**Bönes till**). This, in turn, was followed by the well pronounced *Förnes* Interstadial, which is probably analogous to the Brörup. An almost complete deglaciation, which existed even in the mountains of southern Norway, occurred during this time period. In the coastal areas no ice was present to 28,000 years B.P., including during the *Nygaard* (ca. 55,000-45,000 years B.P.) and the *Ålesund* (*Sandnes*) interstadials and the intervening short ice advance. After the Ålesund, which ended the Middle Weichsel interval, the ice reached its maximum extent during the *Haugesund (Rogne)* advance. A full deglaciation of this territory occurred only during the Holocene.

In *Sweden* (LUNDQVIST, 1986), within the central zone of the glaciation, evidence for the Lower and the Middle Weichsel has preserved poorly. Here, however, there is evidence for the *Leveäniemi* Interglacial (Eem) and for the *Jämtland* Interstadial, which corresponds to the Brörup. Data on hand indicate that a considerable warming occurred at that time and that deglaciation occurred even in Central Sweden (LUNDQVIST, 1986). A warm episode, dating to ca. 27,000-21,500 years B.P. (the *Younger Dösebacka — Ellesbo* interstadials), is recognized in Southern Sweden. Following the glacial maximum at 20,000 years B.P., here deglaciation occurred after 14,000 years B.P. and continued until 8,500 years B.P.

Table 3. Late Pleistocene glacial chronostratigraphy of Northern, Western and Central Europe

Upper Weichsel	Younger Dryas Stadial (11,000-10,300 years B.P.) Alleröd Interstadial (11,700-11,000 years B.P.) Old Dryas Stadial (12,000-11,700 years B.P.) Bölling Interstadial (13,000-12,000 years B.P.) Late Baltic advance Oldest Dryas Stadial (14,000-13,000 years B.P.) Asnaes Interstadial (15,000-14,000 years B.P.) Ne ice advance (20,000-15,000 years B.P.) Interphase Old Baltic and Norwegian advances
Middle Weichsel	Denekamp Interstadial (ca. 25,000 years B.P.) Cooling (stadial) Hengelo Interstadial (Oxygen Isotope Substage 3a; ca. 37,000 years B.P.) Cooling (stadial) Moershoofd Interstadial (Oxygen Isotope Substage 3c; ca. 47,000 years B.P.) Karmöy cooling (Oxygen Isotope Substage 4)
Lower Weichsel	Odderade, St. Germain II. Interstadial (Oxygen Isotope Substage 5a) Melisey II Stadial (Oxygen Isotope Substage 5b) Brörup, Amersfoort, St. Germain I. Interstadial (Oxygen Isotope Substage 5c) Herning cooling (stadial, Oxygen Isotope Substage 5d)
Eem Interglacial (Oxygen Isotope Substage 5e)	

In *Finland*, Lower Weichsel data indicate an interstadial — the *Peräpohjola* Interstadial (probably corresponding to the Jämtland and the Brörup) — which featured a considerable deglaciation accompanied by the development of coniferous forests (HIRVAS *et al.*, 1981; MÄKINEN, 1979; KORPELA, 1969). The main advance of the ice took place during the second half of the Weichsel (**till bed II**) while the late glacial advance is represented by **till bed III**.

LUNDQVIST (1986) notes that active deglaciation of the entire Scandinavian ice sheet began after 13,000-12,000 years B.P., but that it was interrupted by an ice advance during the *Younger Dryas* (11,000-10,300 years B.P.). Data obtained on the ice-free status of the North Sea (BOWEN *et al.*, 1986) indicate that the western margin of the Scandinavian ice sheet did not extend far west even during its maximal extension. A single glacial system did, however, exist on the *British Isles* and the ice did cover the Irish Sea (the **Devensian ice sheet**). In spite of this, Southern England and Southern Ireland remained mostly ice-free.

The chronostratigraphy of the **Devensian ice age**, which followed the **Ipswich** (Eem) **Interglacial**, has not been elaborated to date. No firm data exist about the presence of the ice sheet during the Early and the Middle Devensian. BOWEN *et al.* (1986), along with other specialists, do allow the existence of mountain glaciers during this period.

Late Devensian glacial deposits are widespread in the area. This depositional sequence begins with the **Dimlington till** (radiocarbon dated on organic matter to 18,000 and 18,240 years B.P.). The organic lake deposits which overly this till date between 16,700-13,000 years B.P. While the ice sheet probably retreated considerably by 14,500 years B.P., this retreat was uneven and was interrupted by various readvances (the *Welsh*, the *Irish Sea*, and the *Drumlin* readvances).

While the glaciers did somewhat increase in size after the *Windermere* Interstadial (ca. 14,000 years B.P.), they were confined to the mountains of Scotland (*Loch Lomond* Stadial). This stadial dates to ca. 11,000 years B.P. and can thus be correlated to the Younger Dryas.

In *Poland*, the **Weichsel (Vistula) ice age** is subdivided into 5 main parts consisting of 3 stadials and 2 interstadials. MÓJSKI (1988) attributes the first *Kaszuby* Stadial to the Eovistulian (110,000-90,000 years B.P.), the second — *Pregrudziadz* — to a period between 60,000-50,000 years B.P., and notes the existence of the *Konin* Interstadial between them. During both stadials the ice cover was not very extensive. The whole Middle Vistula (50,000-20,000 years B.P.) corresponds to the complex *Grudziadz* Interstadial which consisted of two warm and one cool intervals. The maximum extent of the ice during the main phase between 20,000-18,000 years B.P. is recorded in the **Poznan** and **Pomerania marginal morainic belts**. The period after 13,000 years B.P. witnessed a steady retreat of the ice sheet.

The Russian Plain of Eastern Europe. There is a detailed Late Pleistocene chronostratigraphy for the glaciated areas of the East European — Russian — Plain. In spite of this, the dating of the maximum extent of the **Valdai ice sheet** still remains under discussion. Some scholars argue that the Early and the Late Valdai ice sheets were not only comparable in size but that the former was larger than the latter one (ZARRINA *et al.*, 1989). The majority of opinion, however, holds that the maximal extent of the ice sheets took place during the Late Valdai (e.g., VOZNYACHUK, 1985; CHEBOTAREVA and MAKARYCHEVA, 1974; FAUSTOVA and BORISOVA, 1986; VELICHKO and FAUSTOVA, 1987; also see the map on page 49 of this atlas). In spite of these differences, almost all scholars recognize the same or similar sequence of events within the glaciated zone.

After the **Mikulino Interglacial** the onset of the **Early Valdai** Stage was marked by a short *Kurgolovo* (or *Åland*) cool period which was followed by the more pronounced *Upper Volga* Interstadial. During this interstadial the heat and moisture supply permitted the growth of coniferous (mostly spruce) forests which probably included some broad-leaved species. This interstadial is probably correlated with the Brörup and the Amersfoort while in the periglacial zone it corresponds to the *Krutitsa* Interstadial.

ZARRINA *et al.* (1989) have developed a slightly different chronostratigraphy. They recognize the existence of three warm phases during the Early Valdai and consider the second of these (and not the first — Upper Volga one) to have been the warmest as well as the equivalent of the Brörup interval. They see this period followed by the *Lapland* cool phase and the *Kruglitsa* warming. The Early Valdai stage terminated with the *Shestikhino* cold period (52,000-47,000 years B.P.). ARSLANOV (1975), ZARRINA *et al.* (1989) see this period as the time of the maximum extent of the Early Valdai Glaciation.

The **Middle Valdai**, which we date between 50,000 and 35,000 years B.P., featured great instability in climatic and environmental conditions. It consisted of three warming and three cooling phases. Paleobotanical data indicate that these cold phases were less severe than those during the Early and the Late Valdai. Given that the climate at that time was rather cool and wet, we can define this Middle Valdai as a long-interstadial.

The *Dunaevo* Interstadial (the *Bryansk* interval in the periglacial zone), radiocarbon dated between 30,000 and 25,000 years B.P., preceded the **Late Valdai** Stage. Data indicate that at this time the Russian Plain was covered by taiga forests.

The coldest phase of the entire Valdai occurred between 23,000 and 17,000 years B.P. when the Scandinavian ice sheet reached its maximal extent. However, the northeastern part of the Plain, an area covered by the **Novaya Zemlya ice sheet**, witnessed an earlier glacial maximum, which occurred here no later than 33,000 years B.P.

During the Late Glacial there were at least three extensive periods during which there was either a halt in the ice retreat or even, possibly, ice readvance. The *Edrovo* phase dates to ca. 16,000 years B.P. and the *Vepsovo* phase between 16,000 and 14,000 years B.P. This was followed by a brief *Rauniss* warming (between 14,000-13,000 years B.P.) and two subsequent phases — the *Luga* and the *Neva* — during which the ice margin reactivated. These two phases were separated by a warming interval, dated between 12,750 and 12,250 years B.P. which probably corresponds to the *Bölling*. After that the ice sheet retreated rapidly, especially during the *Alleröd* warming (11,700-11,000 years B.P.). A new advance occurred during the *Younger Dryas* (11,000-10,300 years B.P.) when the ice sheet expanded over the major portion of Karelia *(Table 4)*.

Table 4. Late Pleistocene glacial chronostratigraphy of the Russian Plain in Eastern Europe

Late Valdai	Dryas III Alleröd Interstadial Neva advance Bölling Interstadial Luga advance Rauniss Interstadial (14,000-13,000 years B.P.) Vepsovo cold phase Maximal (Usvyach) phase (23,000-17,000 years B.P.)
Middle Valdai	Dunaevo (Bryansk) Interstadial (30,000-25,000 years B.P.) Shensk cooling (ca. 35,000 years B.P.) Shapurovo warming Surazh cooling Grazhdansky Interstadial Kashin cooling Krasnogorsk warming (minimum 50,000 years B.P.)
Early Valdai	Shestikhino cold phase (Early Valdai glacial stade) Kruglitsa warming Lapland Stadial Upper Volga warming Kurgolovo cold phase
Mikulino Interglacial	

Siberia. The chronostratigraphies of *Central* (the Yenisey River basin, the North Siberian Lowland, the Taimyr Peninsula) and *West Siberia* (primarily the Ob River basin) are usually treated separately. The schemes for both regions, however, do have much in common (ARKHIPOV and ISAEVA, 1987). The Late Pleistocene here began with the **Kazantsevo Interglacial** which is analogous to the Mikulino in Eastern Europe.

According to the recently developed stratigraphic scheme (ARKHIPOV, 1971; ARKHIPOV and VOTAKH, 1981; ARKHIPOV *et al.*, 1986), the Late Pleistocene glaciations belong to a single period — the **Zyryanka**. In this scheme the **Early Zyryanka** corresponds to the Zyryanka proper in the earlier stratigraphic schemes while the **Late Zyryanka** corresponds to the **Sartan Glaciation**.

The Early Zyryanka Stage is represented by the *Ermakovo* glacial horizon which consists of two **tills** — the earlier **Kormuzhikhanka** and the later **Khashgort**. The deposition of these two was separated by a pronounced warm interstadial with environmental conditions close to present-day ones.

The **Middle Zyryanka** — the *Karginsky* Stage — featured repeated fluctuations in environmental conditions. It began with a cool *Kharsoim* Interstadial (50,000-40,000 years B.P.) and was followed by a cold *Lokhpodgort* interval (40,000-30,000 years B.P.) reflected in the moraine found at the mouth of the Ob River.

The Middle Zyryanka Stage ended in the *Karginsky* Interstadial (29,000-25,000 years B.P.) — seen as a chronostratigraphic counterpart of the Dunaevo—Bryansk Interstadial on the East European Plain (although the latter is often attributed to the Late Valdai).

The Late Zyryanka Stage (younger than 25,000 years B.P.) consists of the Sartan Glaciation with its maximal phase (the *Salekhard-Uval*) and two recessional stages (the *Sopkei* and the *Polar Urals*).

Along the lower Yenisey and on the North Siberian Lowland, the Lower Zyryanka Stage is represented by the two-part *Murukta* glacial horizon. Here the Middle Zyryanka (Karginsky) Stage began with a warm interval — the *Lower Karginsky* Interstadial — which was followed by the *Konoshchelye* cooling (38,000-35,000 years B.P.). This stage terminated in the *Lipov–Novoselov* Interstadial when, as KIND and LEONOV (1982) have argued, climatic conditions were probably warmer than today. The chronostratigraphic position of this interstadial is akin to that of the Karginsky Interstadial in West Siberia.

In North Siberia the **Sartan ice sheet** reached its maximal extent between 20,000 and 19,500 years B.P. — a period reflected in the *Gydan* marginal formation. This was followed by a period of ice retreat which featured a number of stadials (the *Tiutei* 16,000-15,000 years B.P., the *Nyapan* at 15,000-13,000 years B.P.; and the *Norilsk* at 11,400-10,700 years B.P.).

Data on hand clearly indicate the extensive presence of the ice not only during the Late Zyryanka (Late Valdai) but also in the beginning of the Late Pleistocene. Moreover, here the earlier ice sheet was even more extensive — a case considerably different from that on the Russian Plain. This was due to the smaller dimensions of the Siberian anticyclone at the beginning of the Late Pleistocene which permitted the western cyclones to penetrate into the Siberian interior. Subsequently, with progressive cooling, a strong anticyclone blocked the penetration of the cyclones (VELICHKO, 1987). *Table 5* correlates the West and Central Siberian stratigraphic schemes.

The Northeast of the USSR. While Eastern Europe and the Siberian Plain were dominated by the ice sheets, mountain glaciers were predominant in northeast Asia. Traces of mountain glaciers are poorly preserved here — a situation common to most of the glaciated mountain regions of the Northern Hemisphere. BESPALYJ and GLUSHKOVA (1987) noted that the glacial history of this region was broadly synchronous to that of adjacent areas.

The Late Pleistocene began with the **Valkatlen Interglacial** Transgression recorded on the coasts of the Chukchi Peninsula. The glacial epoch is subdivided into three stages. The first stage (the *Tylkhoi* Glaciation in local schemes), represented by the **Vankarem** and the **Tylkhoi tills**, terminated ca. 75,000-60,000 years B.P. The second (*Karginsky*) stage, represented by the **Amguema marine beds**, was characterized by climatic conditions similar to present-day ones. During this stage, ca. 33,000-30,000 years B.P., there was an ice advance (the *Zhigansk*) . The lowest temperatures were reached during the late stage some 20,000-19,000 years B.P. This time interval is reflected in the

Table 5. Glacial chronostratigraphical correlation between West and Central Siberia

West Siberia	Stages	Central Siberia
Polar Urals Stadial Sopkei Stadial Salekhard-Uval Stadial	Late Zyryanka (Sartan)	Norilsk Stadial (11,400-10,700 years B.P.) Nyapan Stadial Tiutei Stadial (16,000-15,000 years B.P.) Gydan Stadial (20,000-19,500 B.P.)
Karginsky Interstadial (29,000-25,000 years B.P.) Lokhpodgort cold phase Kharsoim warming	Middle Zyryanka (50,000-30,000 years B.P.)	Lipov-Novoselov Interstadial (29,000-23,000 years B.P.) Konoshchelye cold phase (38,000-30,000 years B.P.) Lower Karginsky Interstadial
Khashgort glacial phase (Ermakovo Glaciation) Bogdashkino Interstadial Kormuzhikhanka glacial phase	Early Zyryanka	Murukta Glaciation
	Kazantsevo Interglacial	

Khaimikino Glaciation on Kamchatka and the *Iskaten* Glaciation on the Chukchi Peninsula.

The dimensions of the glaciers show the same pattern asobserved in Siberia — namely, that the older Late Pleistocene Glaciation (the Tylkhoi) was more extensive than the younger (Khaimikino) one.

Conclusions. In spite of local specificities, the glaciated areas of the Northern Hemisphere show an overall similarity in the sequence of glacial events.

These are as follows:

1. *The Late Pleistocene began with a well-documented interglacial stage*: the Sangamon, Ipswich, Eem, Mikulino, Kazantsevo, correlated with the Substage 5e of the oxygen isotope record.

2. In most regions, the Late Pleistocene *glacial epoch is divided into three stages.*

a. The *first of these saw glaciation* in some regions, most pronouncedly in eastern Eurasia (Siberia, the Northeast). *This stage contained one of the most distinct warm intervals* — the one probably quasi-synchronous to the Brörup in Europe.

b. The *middle stage* featured rather unstable climatic conditions which changed from cool to temperate humid. *An interstadial*, dating between 30,000 and 25,000 years B.P. — the Plum Point, Denekamp, Dunaevo, Bryansk, Karginsky — *is well documented* in almost all of the regions. Thus this interstadial represents an important chronostratigraphic and paleogeographic component of the Late Pleistocene.

c. The subsequent *third stage included the maximal cooling* (between 21,000 and 19,000 years B.P.) and, in many regions, *glaciation*. It was followed by active deglaciation which began ca. 13,000 years B.P.

GLACIATION DURING THE LAST GLACIAL MAXIMUM

by A. A. Velichko and M. A. Faustova

(Explanatory notes to map on page 49)

The geomorphic and chronostratigraphic aspects of the Late Pleistocene glaciations (Valdai, Vistula, Devensian, Wisconsin) are reasonably well studied for the whole Northern Hemisphere. As the data accumulated, different reconstructions of ancient glacial systems were developed. Some of them almost completely eliminated large ice sheets, while others presumed the existence of vast coalescing ice domes which dominated both polar and temperate latitudes in the Northern Hemisphere, including the Arctic shelf. These differences in reconstructions seem to have resulted not only from the different conceptual approaches used by different specialists, but also from a somewhat inadequate knowledge of former glacial phenomena. Reliable data obtained only recently enable us to model the spatial and temporal differentiation in the evolution of the **ice sheets.**

Detailed chronostratigraphic studies, including radiocarbon dating, indicate the existence of two main cold stages within the Late Pleistocene ice age both in North America and Eurasia. They included an early stage (from 90,000-80,000 to 50,000 years B.P.) and a late one (from 30,000-25,000 to 10,000 years B.P.). The two stages were separated by an interval when the ice sheets were reduced in size. Several warm phases are identified within this interval, but these were not exactly synchronous in the different regions. The global character of the climatic warming in the middle of the Late Pleistocene cold epoch indicates a somewhat uniform development of the last glaciation in the Northern Hemisphere. At the same time, some regional differences have also been recorded. In some regions, the phase of active expansion of the ice sheets began about 25,000 to 23,000 years B.P., in other regions it occurred earlier. While the coldest phase is dated between 20,000 to 18,000 years B.P., it did not coincide everywhere with the maximal expansion of the ice sheets.

Recent research revealed a rather complex structure of the surface of ice sheets and equally complex dynamic interaction between separate sheets and their parts. These results contradict some previous concepts about ice sheets being stationary in character and one-centered in structure.

The **Late Wisconsin glacier complex** over North America was the largest in the Northern Hemisphere. It consisted of several units. The dynamics of its central part were controlled by the development of a number of **ice domes: New-Quebec, Keewatin, Innuitian, Labrador, Foxe** and others. These domes maintained separate flow patterns but formed a confluent ice mass (FULTON, 1984; ANDREWS, 1982; WRIGHT, 1983). The eastern and western flanks of the North American glacial complex were formed by separate ice masses which spread from the Appalachian and Cordilleran mountains. The Appalachian ice basically coalesced with the central ice sheet. The Cordilleran ice sheet included systems of intermontane, piedmont and valley glaciers.These glaciers flowed eastward, out of the mountains, and coalesced with the westward flowing Laurentide ice sheet only occasionally. An **ice-free corridor** existed between the two ice masses during most of the Late Wisconsin stage. At times, ice tongues passed this corridor from one or the other (or both) of the ice sheets. Numerous nunataker rose above the ice sheet surface in the mountains and, occasionally, on the midcontinent. The main part of the ice spread southward forming several ice lobes. Detailed stratigraphical studies of the marginal zone point to a series of ice advances. Since the magnitude of advances often differed in adjacent lobes, these advances appear to have culminated at somewhat different times (from 21,550 to 19,000 years B.P.).

In the high latitudes of North America, ice sheets were thinner and spatially quite restricted. Sizeable parts of the Alaskan lowlands as well as of Banks and Somerset islands were ice free during the Late Pleistocene (FULTON, 1984; PORTER *et al.*, 1983; DYKE, 1978). Geological and geomorphological studies, together with radiocarbon dating carried out during the last decade, indicate that the Late Wisconsin ice occurred at its maximum extent in the Canadian Arctic 10,000 to 8,000 years later than at its southern margin. The maximum advance during the last glaciation in Greenland is dated not much earlier than 15,000 years B.P.

The glaciation of Eurasia was no less complex than that of North America. In the westernmost region (the British Isles), Late Devensian glaciers centered in the mountains of Scotland, Cumberland, Wales, and Ireland, and did not cover the whole area of the Isles (PRICE, 1975). The southern and southwestern regions of Great Britain and Ireland were ice free (SHOTTON, 1981). The ice flowed eastward from the highlands and extended beyond the present-day shoreline onto the then emerged shelf of the North Sea. The **Scandinavian ice sheet** spread into the same area from the east. These two ice advances, however, probably did not culminate at the same time, and a rather broad zone in the central North Sea could have remained ice free (PRICE, 1975; FLINN, 1978; SISSONS, 1981; LARSEN and SEJRUP, 1990).

The Scandinavian ice sheet spread from two independent centers located in the southwest and northeast of the Scandinavian highlands. Its northern and northwestern margins extended onto the nearshore parts of the Norwegian Sea floor. The boundary of this glaciation is represented by the Egga end-moraine complex which dates to about 20,000 to 18,000 years B.P. (although a controversy exists about the age of the moraines). In the south, the ice sheet did not reach its maximum extent everywhere at the same time. Radiocarbon dates indicate the time-transgressive character of this culmination which occurred earlier in the east than in the west (from 24,000 to 17,000 years B.P., respectively).

The Scandinavian ice sheet extended onto the northern part of the Russian Plain which, in the northwest, was also covered by ice from Novaya Zemlya. Marginal zone features such as tills, as well as petrography and radiocarbon dates from underlying sediments, indicate that the advances of the Scandinavian and Novaya Zemlya ice sheets were heterochronous. The possibility of such heterochronous advances has already been suggested for areas of Canada east of the Rocky Mountains and for the North Sea shelf. VELICHKO suggested the term **zones of heterochronous advances** (as distinct from zones of convergence) for such zones which were alternatively covered by ice from different centers. Such zones within the glacial system could have coexisted with **zones of convergence** in their vicinity.

The distribution of ice sheets in Siberia has been a subject of discussion for a long time. Recent geological and geomorphological studies, as well as new radiocarbon data, have provided further arguments in favour of a rather limited extent of glaciers during the **Sartan** Stage of the Late Pleistocene glaciation. Peat and mammoth bones not covered by till dating to about 40,000 to 25,000 years B.P. and younger have been found on Taz and Gydan peninsulas (northwestern Siberia). A mammoth bone from solifluction deposits (discovered by V. D. TARNOGRADSKY and F. A. KAPLYANSKAYA) on the northern Gydan Peninsula was dated to 30,100 ± 300 years B.P. (GIN-3742). Peat from the Late Pleistocene marine terrace on the Gydan Peninsula was dated to 15,830 and 15,360 years B.P. (AVDALOVICH and BIDJIEV, 1984). The ice lobe which spread onto southwestern Yamal Peninsula came from the Polar Ural Mountains and covered only a small part of the peninsula. Glaciers which penetrated the lower Yenisey Valley down to the latitude of the city of Dudinka and which formed typical glacial landforms there, appeared to be older than Sartan. Wood samples from overlying sediments in the "Ice Hill" section were radiocarbon dated to 43,000 to 41,000 years B.P. as well as older than 50,000 years B.P. (ASTAKHOV and ISAEVA, 1985). These data seem to corroborate information obtained earlier which indicated that the North Siberian Lowland and most of the Taimyr Peninsula were ice free during the Sartan stage. A section of the 30 m high lake terrace on the western shore of the Taimyr Lake is characterized by a series of successive radiocarbon dates from 30,000 to 11,000 years B.P. (KIND and LEONOV, 1982). On the Anabar Plateau and on the Taimyr Peninsula only mountain glaciers existed. At the same time, the sediments underlying pronounced morainic ridges in some northern parts of the North Siberian Lowland gave dates from 37,000 to 29,000 years B.P. Conclusive evidence for the extent of Sartan glaciation on the Taimyr Peninsula still remain to be obtained.

Small ice sheets existed on the Putorana Plateau. There was no ice sheet further east, but some mountains were heavily glaciated: Verkhoyansk Ridge (reticulate glaciers), Chersky Ridge, and eastern part of the Koryaki Upland and Kamchatka, where the glaciers advanced onto the exposed shelf.

Recent data on the extent of Late Pleistocene glaciation in the Arctic are consistent with the idea that Sartan glaciers were rather limited. The Kara Sea shelf, except for its western margin, was not glaciated. Only small ice domes existed even at the Severnaya Zemlya islands, leaving part of the islands ice-free. This, in particular, is indicated by discoveries of mammoth bones on the marine terrace not covered with till which date to 24,000 to 19,000 years B.P. (MAKEEV *et al.*, 1979).

The most important data on the **glaciation of the Arctic shelf** were obtained from Novaya Zemlya and adjacent sea areas. Submarine geological studies summarized by MATISHOV (1984) confirm that parts of the Barents and Kara seas were unglaciated (based on data obtained by Yu. A. PAVLIDIS (see AKSENOV *et al.*, 1987) The marginal zones of the Novaya Zemlya glaciers lay only 100 to 150 km beyond the present-day coast (AKSENOV *et al.*, 1987; BIRYUKOV *et al.*, 1988). The Barents Sea's shelf probably was not occupied by an ice dome. Rather, ice lobes from Scandinavia, Spitsbergen, Franz Josef Land, and Novaya Zemlya reached the periphery of the shelf. It seems probable, however, that glaciers were more extensive at earlier stages of the Late Pleistocene glaciation (as indicated by works on Spitsbergen by SALVIGSSEN, 1979; TROICKY *et al.,* 1979, and others). At the time in concern, glaciers were far more limited. They reached the shelf only through a few fiords and a considerable part of the land was left unglaciated.

The reconstruction of the last Late Pleistocene glaciation (25,000 to 10,000 years B.P.) can be summarized as follows:

In the high latitudes, the glaciers were mostly restricted to islands and adjacent shelves. On the continent main ice masses were formed in lower polar and middle latitudes and had their centers in the highlands.

The distribution pattern of the ice sheets indicates a distinct asymmetry: ice sheets in North America were the hugest while those in Europe were less in extent. Thus, both the area and thickness of the ice progressively decreased eastward. It is possible, however, that at the initial stages of glaciation, Siberian glaciers could have been more active and could have grown faster than those in western regions. This phenomenon can be explained by the climatic features of the ice age within the frame of the **concept of the asymmetry of glaciation** (VELICHKO, 1980).

The glaciers had non-stationary regimes and individual ice sheets usually included several ice domes. Because of the differences in their position as to atmospheric centers (including sources of moisture), different ice sheet dynamics were not completely synchronous. This accounts for the fact that margins of adjacent ice sheets could not only merge (forming zones of convergence), but also penetrate, in turn, the same area, never coalescing and constantly leaving ice-free corridors.

CRYOGENIC REGIONS DURING THE LAST GLACIAL MAXIMUM (PERMAFROST)

by A. A. Velichko and V. P. Nechaev

(Explanatory notes to map on page 49)

During the cold stages of the Late Pleistocene, permafrost spread over vast regions of the Northern Hemisphere, and its area exceeded considerably that of the ice sheets. It was not until recently, however, that the great importance of this phenomenon — the so-called **underground glaciation**— was fully understood. Mapping ancient permafrost does not have a very long tradition. Works by TUMEL (1946) for the USSR and by POSER (1948) for Europe represent one of the first attempts at such reconstructions. Later, some small-scale maps were composed for the whole Late Pleistocene cryogenic region. The second half of this period was identified as a separate **cryogenic stage of Pleistocene** prehistory (VELICHKO, 1973). These first results provided a basis for more detailed reconstructions of the cryolithic zone during the latest and most pronounced cooling of climate (about 20,000 to 18,000 years B.P.).

The map presented here is based on numerous data available for the Late Pleistocene cryolithic zone. Studies of periglacial (or, more precisely, paleocryogenic) areas permit to distinguish between several phases of cryogenesis during the Late Pleistocene. These phases are best manifested in the loess-paleosol series in the central Russian Plain. They are recorded in three horizons of ice wedge pseudomorphs and in other **cryogenic structures.** These indicate episodes of the reactivation of cryogenic processes, as well as deep transformations of the ground surface and surficial deposits, which occurred due to low temperatures and specific regimes of humidity. Three **cryogenic horizons** on the Russian Plain, named **Smolensk, Vladimir,** and **Yaroslavl**, date from about 90,000 to 80,000; 25,000 to 23,000; and 20,000 to 17,000 years B.P., respectively. The best developed is the youngest — Yaroslavl — cryogenic horizon. The most active phase of its formation coincided with the coldest time of the Late Pleistocene glacial epoch. Cryogenic structures of this horizon (which include systems of large **ice wedges** and **ice-earth veins)** suggest the lowest temperatures during cryomorphogenesis. At this time permafrost reached its maximum expansion in the Northern Hemisphere. Its spatial regularities are shown in map on page 49.

Reconstructions were done for regions beyond the limits of the ice sheets, i.e., for their "periglacial zone" (in the broad sense of the term). The southern limits of this zone, however, do not exhibit any relationship with the boundaries of the ice sheets. Instead, they were mostly associated with the limits of sea-ice, and, within the continents were controlled by topography.

Our reconstruction is based on the same main features of permafrost used for modern permafrost zonation. They include: lateral continuity, annual temperatures of the frozen ground, and thickness of the permafrost.These three parameters are indicated on the map only if the available cryogenic information dates to the time interval under consideration (the coldest phase of the Late Pleistocene). It should be noted that the distribution pattern of the data is very uneven. Since most of the reported paleocryogenic features are from northern Eurasia, this region was used as the base area for reconstructing all three geocryological parameters. The important aspects of this methodological approach have been discussed elsewhere (VELICHKO *et al.,* 1978; BAULIN *et al.,* 1981).

Data from Northern Eurasia at the end of the Pleistocene reveal some regional permafrost features within this vast area. In northeast Asia, large ice wedges reached a depth of 20 to 30 m. This implies ground temperatures between –15 to –20 °C (KAPLINA and KUZNETSOVA, 1975). In western Siberia and in the European USSR, permafrost spread on an average 2000 km south of its present day boundary. Permafrost was about 200 to 400 m thick at the latitudes of Moscow and Novosibirsk. These regions now are beyond the permafrost area. On the Volyn-Podol Upland (west of the Dnieper River, 49° to 50° N) ice-wedge casts are reported from late Valdai loess. They are 3 to 5 m deep, form polygons about 20 to 25 m wide, and the fossil active layer was about 1 m thick (NECHAEV, 1980). At the same latitude further to the east some wedge-shaped structures were described and diagnosed as pseudomorphs of **polygonal ice veins** (AUBEKEROV, 1982). Ground temperatures in this region were much lower than at present. The region was situated close to the southern boundary of a vast area with **continuous permafrost**. Zones of **discontinuous** and **sporadic permafrost** with deep seasonal freezing lay to the south of this region.

The reconstruction of geocryological parameters for Western Europe is based on recorded data as well as on permafrost maps published by J. KARTE, A. PISSART, J. TRICART, A. VELICHKO, and others (GERASIMOV and VELICHKO, 1982). During the maximum cooling (20,000 to 18,000 years B.P.), most of the areas north of the Alps and of the Carpathians were characterized by **permafrost of** the so-called **"Siberian" type.** This type contains continuously frozen ground with relatively low temperatures (usually not higher than –3 to –5 °C) and features a wide distribution of **ice-wedge polygons.** Geocryological conditions were less rigorous only in the westernmost regions and in the Danube Basin.

To illustrate changes in permafrost along a north to south transect one can compare the reconstructed permafrost from Poland and Hungary with each other. In northern Poland Late Vistula permafrost features are found everywhere. Here even tills and sands are cut by frost fissures and ice veins (DYLIK, 1966; GOZDZIK, 1973). In southern Poland numerous paleocryogenic distortions are also found in loess (JERSAK, 1973; MARUSZCZAK, 1980, and others). On the other hand, continuous permafrost was not found in most of the Hungarian territory despite considerable cooling towards the end of the Pleistocene (PÉCSI, 1969).

Reconstructions of permafrost in low latitudes (Central and Eastern Asia in particular) are still problematic. This is due both to a lack of detailed studies on these regions as well as to more complicated topographic patterns there.

In Mongolia, Late Pleistocene permafrost was distributed (at least, sporadically) up to 45° 30' N (GRAVIS and LISUN, 1974). These authors consider it quite probable that sporadic permafrost may have expanded southward considerably, and that even there isles of permafrost could have existed locally under favourable topographic conditions.

Permafrost was widely distributed in eastern China north of 45° N between 23,000 and 12,000 years B.P. Its southern limit, as indicated by the presence of patterned ground in the Taibai Mountains, Quinling Range, dating from the Late Pleistocene, was probably situated as far south as 40° to 39° N (CUI ZHIJIU and XIE YOUYU, 1982; LI YOUMING *et al.*,1982). It can be supposed that areas of deep winter freezing (and, probably, of sporadic permafrost) adjoined the alpine permafrost regions in the Pamir, Tibet, Tien Shan, etc. For example, relict periglacial landforms (micropolygons, stone stripes, etc.), presumably of Late Pleistocene age, are reported from eastern Anatolian Mountains above 2700 m a.s.l. and from the Elburz Mountains above 2100 to 2300 m a.s.l.

Several levels of fossil **involutions** and of ice-wedge casts were recorded in the Far East (Hokkaido Island) in sections which include the 30,000 to 15,000 years B.P. time interval (ONO, 1984). Paleogeographic data indicate that winter sea-ice existed off the coasts of Hokkaido and that permafrost was present in northern and eastern parts of the island. Thus, in the easternmost regions of Asia, permafrost expanded at least as far south as 44° to 43° N.

On this question considerable research has been contributed in North America during the last 15 to 20 years.The results have been critically reviewed and summarized by PÉWÉ (1983). The data suggest that continuous permafrost encompassed nearly all the unglaciated areas of Alaska and of the exposed shelf (at present, only one third of the Alaskan territory is included into the Wisconsin-age continuous permafrost zone). This permafrost was thicker and its temperatures considerably lower than observed today. Ice-wedge polygons were formed almost everywhere in Alaska. Ice-wedge casts are also reported from the Pribilof and St. Lawrence Islands. We calculated the Alaskan Late Wisconsin geocryological parameters shown on the map on the basis of paleogeographic data as well as from extrapolations from other regions of Beringia.

For the territory of the United States, we agree with PÉWÉ's (1983) opinion that, with the exception of very limited areas, there was no zone of continuous permafrost on the plains south of the ice sheet. The presence of discontinuous permafrost here is suggested by inactive **rock streams, solifluction deposits**, and by other **periglacial features.** The distribution of ice-wedge casts was rather restricted. In many areas of North America the permafrost zone was less than 200 km wide probably because it was situated nearly at latitude 40° N (i.e., farther south than in Europe). During the Wisconsin, an alpine permafrost zone existed in the western mountains of the US territory. There, geocryological conditions could have been more rigorous than in the Alps, because of higher elevations.

Near to the Equator the mean annual temperatures are known to have lowered by 3 or 4 °C during the last glacial maximum as compared with the present values (BONNEFILLE, 1987). the decrease in temperature could bring about the lowering of the **alpine permafrost** limits by a few hundreds of meters in the Kenya Mountains or in high ranges of the Andes. Map on page 49 represents the first attempt to present a differentiated geocryological reconstruction on a subglobal scale. Hopefully, it will serve as a catalyst for a more detailed paleocryological mapping in the future.

The map shows the following fundamental features for the former permafrost zone of the Northern Hemisphere. First, for most parts of the hemisphere, the zone expanded far to the south and well into the southern part of the present temperate zone. The limit of this zone, in contrast to its present position, ran mostly latitudinally.

The distribution pattern of cryogenous regions during the last glacial maximum exhibited a pronounced asymmetry. The zone was widest (more than 4000 km north to south) in the east and narrowed towards the west. It was a few hundred kilometers wide in Western Europe and not more than 100 to 200 km wide in the Western Hemisphere. The climate was also the most severe in Eastern Asia.

There is a noticeable asymmetry in the distribution of this permafrost which is similar to the asymmetry noted for the glacial regions, but it is inversely directed. The interrelationship between glacial and permafrost asymmetry has been considered by VELICHKO (1980). Both of these asymmetry patterns were controlled by the general circulation of the atmosphere. During the Late Pleistocene glacial maximum, atmospheric circulation was characterized by large anticyclones in Eastern Asia. This was less pronounced farther to the west.

LOESS OF THE LAST GLACIATION

by M. Pécsi

(Explanatory notes to map on page 51)

1. ***Loess as a subaerial formation plays a special role in preserving traces of terrestrial life.*** Loess is one of the widest spread Pleistocene deposits, mantling almost 10 per cent of the lands surface of the Earth. It is distributed in well-defined geographic environments: periglacial tundra and steppe, forest steppe, cold and cold temperate belts and semiarid, hot steppe zones around deserts. Loess does not occur over continental surfaces which were buried by inland ice during the Last Glaciation. The largest areal extension of loesses and loess-like sediments was reached in the Last Glaciation. During this period several young loess layers and intercalated soils formed. Older loess layers were, for the most part, also produced during Pleistocene glacial cycles. Loesses dating before the Last Glaciation are known from considerably smaller areas. Finally, loesses older than Pleistocene glaciations and manifesting the properties of true loess are not detectable.

The slow accumulation and pedogenesis of the loess matter did not present a serious obstacle to continuous life. In certain geographic environments loess preserved not only numerous species but also traces of their activities which took place during paleogeographic environments conducive to loess formation. These remains allow us to reconstruct environmental conditions and to outline paleoecological circumstances present at the time that the loess was formed. Such reconstructions rely heavily on the lithological properties as well as on the texture and mineral composition of the loess. Consequently, the multiple changes in past ecological conditions (organic life, physical and chemical processes) can be monitored in loess sequences over long spans of time and the evolution of the loess-mantled surface and their living organisms are documented well.

2. ***Perspectives for the reconstruction of paleogeographical changes in the Last Glacial on the basis of loess profiles.*** In the past decades numerous attempts were made to determine in detail the duration, quality and quantity of the changes the climate or the geographical environment underwent within the **Last Interglacial/Glacial cycle of the Pleistocene**. Most recently, the investigation of global climatic change — for the forecast of future environmental changes — has become part of several international research programmes. For this reason interregional loess chronological correlations were developed. The establishment of a global chronology is hindered by many factors, first of all, by major differences in the lithostratigraphy of young loess-paleosol sequences in various regions of the Earth.

If representative loess profiles are intended to be used for reconstructing the paleogeographic events during the Upper Pleistocene, we have to rely on well-studied sections. For the chronological evaluation of key sections most of all regionally repeating phenomena can and have to be considered, whereas local ones can only play an ancillary role *(Figs. 8 and 9). Figure 9* indicates *4 to 7 loess horizons* and *4 to 7 paleosols* (humous loess, steppe soil, forest steppe soil or forest soil) in various stages of development occurring in the young loess profiles of Western and Central Europe. At the same time, in the loesses of the Russian Plain in Eastern Europe and in the Great Plains in North America less loess and paleosol layers seem to occur, though certain variations may be due to different dating techniques or to the paleogeographic position of the profile investigated. Thus, for instance, in some frequently cited stratotype sections of the Loess Plateau of China (Luochuan and Xifeng), only one paleosol is found for the Last Glacial cycle (LIU, 1985, 1987), while in the recently analysed young loess profile on a terrace surface (LI *et al.*, 1989), 6 to 7 paleosols and an equal number of loess horizons alternate. In other cases, in the exposures of the Middle Danube Basin, TL data suggest a rather different evaluation of the Last Interglacial soil *(Fig. 8)*.

The **lithostratigraphic variation** in young loesses may often have geomorphological and sedimentological explanations. In some extensive, non-subsiding lowlands with subdued relief, where generally the rate of silt accumulation was also low, loess and intercalated paleosols of small thickness or **polygenetic soils** developed. Under such conditions several subsequent climatic phases are represented by one single polygenetic soil or **soil complex** in the profile. On terraced valley sides and on slopes of interfluves loess sequences are more minutely subdivided than in lowlands or on flat watersheds.

In general, more complete loess sequences seem to occur most of all in *sediment traps* like low-lying small basins, buried dells (small dry vallies), foot-slopes and subsided basins.

Sedimentation gaps, erosional, deflational hiatuses, intercalations of blown sand, loessy soil deposits and other stratified loess-like deposits can also occur in young loess sequences.

It is assumed that stratified slope loesses are partly formed in periods between loess and soil formation, whereas the dells were filled by loess with soil sediment during the solifluction phase after soil formation. Transitional paleogeographical conditions are assumed to have happened in the phase between soil and the ensuing loess formation. Thus, phases of *loess formation, soil formation* and of transitional *blown sand accumulation, dell erosion* and locally *cryogene processes* are distinguished. The **climatic equivalents of the corresponding strata** can be found for these stages. However, there are uncertainties in finding the relevant climates due to obstacles caused by the relief on a local, regional or global scale.

The formation of humous loess and paleosol layers in the Pleistocene periglacial belt are usually associated with *moister-warmer climates*, while loess formation is assumed to have taken place in the cold, but not the driest phases of the glacials. Compared to the climate of loess formation, the preponderance of an even *drier and colder climatic type* seems probable during the time of *ice wedge formation*, the remnants of which occur in 3 to 8 levels in the youngest loess horizons, depending on the region. In the fillings of some young ice wedges blown sand and other subaerial deposits have unambiguously been identified. The degradation and filling of such ice wedges may represent the driest climatic phase. In contrast, the cold and moist environments are represented by cryoturbations, involutions and solifluction phenomena in loess.

The intercalated sand layers of various thickness in the

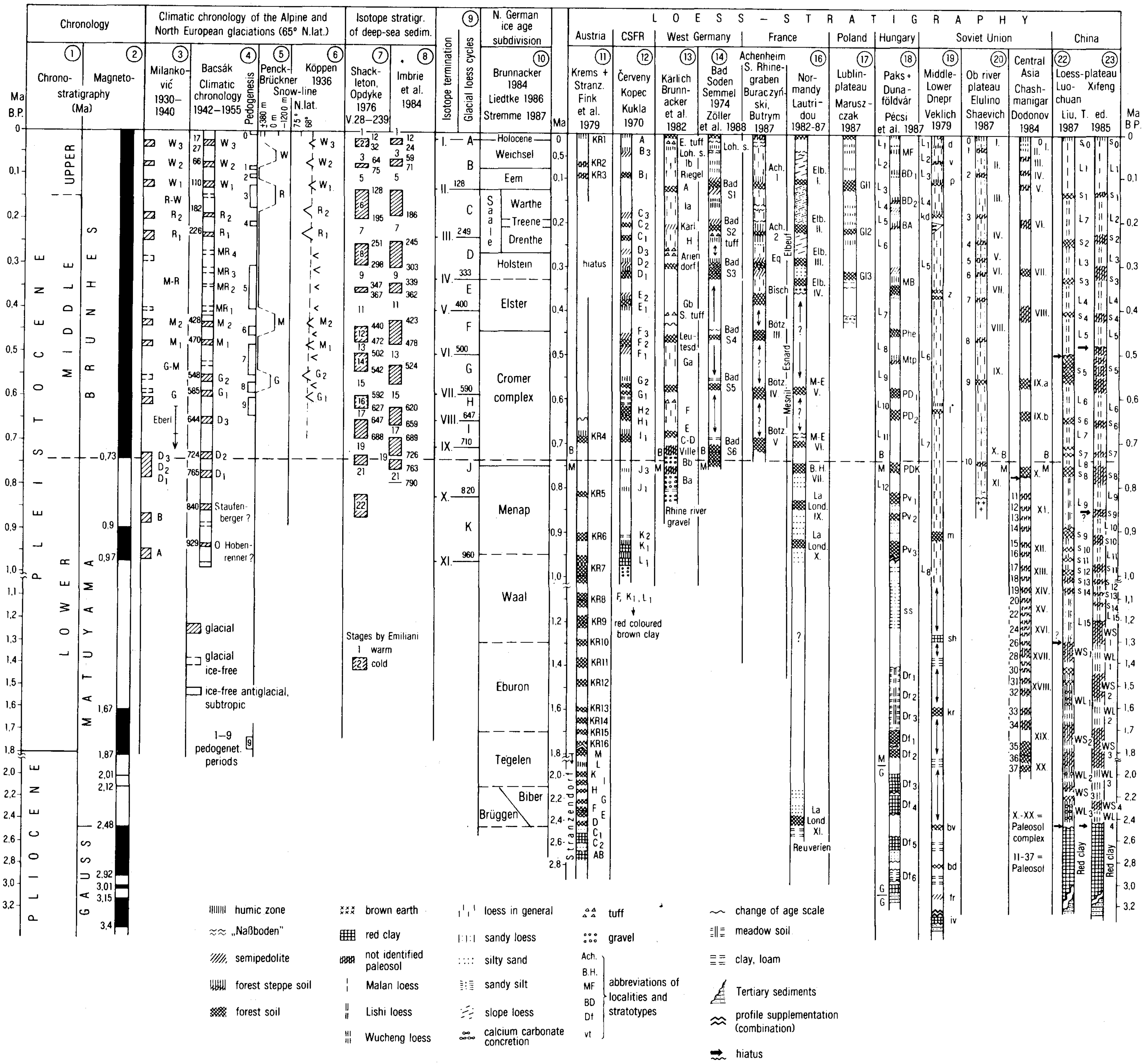

Fig. 8. Provisional correlation between Pleistocene climatic chronology, oxygene isotope stages and loess-paleosol stratigraphy

young loess sequence are hold to be blown cover sand; on alluvial fans and terraces they are fluvial deposits and are associated with floods or meltwater.

The problem of a sound chronostratigraphic evaluation of **double** or **triple paleosols** and of polygenetic soils is unsolved. A polygenetic soil (e.g., from the Last Interglacial) may indicate several phases of soil formation, but it is assumed that a double or triple soil profile, e.g., in a dell, belongs to one single soil formation phase only. Yet the possibility cannot be ruled out that the two members of a double soil are superimposed as a result of more intense soil erosion or slope wash and of eolian sedimentation in the intermediate period. Thus, the time difference between the formation of the two members of such a double soil may be considerable (they may have formed in two interstadials or even interglacials).

The reconstruction of the lithogenesis of the above-mentioned layers is based on their internal structures, mineralogical composition and alteration, the respective periglacial and other physico-chemical, geological and geomorphological phenomena, remnants of soil types or of fossil life. Through the interpretation of its origin the *lithostratigraphical backbone* of the given loess section can be reconstructed. Correlating the lithostratigraphy with a Pleistocene time-scale, local, regional or global loess chronostratigraphy can be achieved. However, this is only a provisional information for the further use of correlation perspectives.

3. ***Perspectives for the subdivision of Last Glacial loess based on ^{18}O isotope stratigraphy.*** Initiated by KUKLA (1970, 1975) the time-scale of deep sea oxygen isotope stages, i.e., their boundaries, are preferred to correlate the loess chronostratigraphy. The so-called **SPECMAP oxygen isotope time-scale**, developed by IMBRIE *et al.*, (1984), in close correlation with changes of ice masses on Earth, contains average age values for stages calculated from five different oceanic boreholes[1] *(Fig. 8)*.

The time-scale by SHACKLETON and OPDYKE (1976), based on the ^{18}O isotope stages of the Pacific borehole (V28-239) close to the Equator (1° S, 160° E), is also often applied. The two oxygen isotope time-scales show similar, but not exactly identical stages for the Upper Pleistocene *(Fig. 8)*. In the Middle and Lower Pleistocene the boundaries of oxygen isotope stages considerably vary in the different deep sea boreholes. The differences are caused by the fact that the ages of the particular stages are founded on the

1 V28-238, V30-40 and DS PD 502 from low latitudes and V22-174 and RC11-120 from midlatitudes.

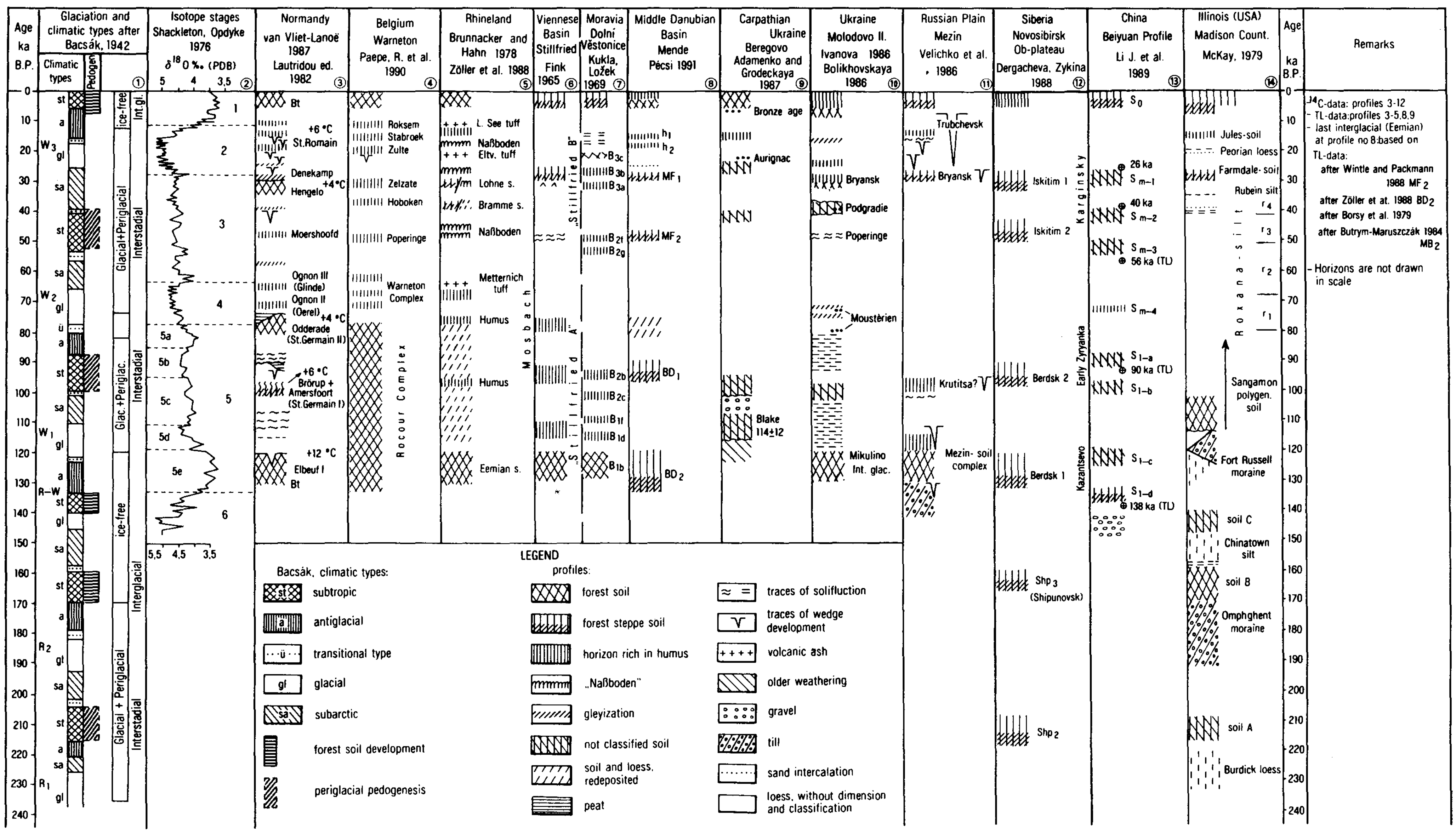

Fig. 9. A tentative correlation of Last Glacial loess profil

estimated rate of sedimentation in the deep sea. Of course, where possible, absolute dates obtained by radiocarbon, paleomagnetic or other methods are also used.

For the past 130,000 years the oxygen isotope stratigraphy indicates three relatively warm climatic stages (nos. 1, 3 and 5) and two relatively cold or cool stages (nos. 2 and 4). Within the warm stages of longer duration warmer (3a, 3c, 5a, 5c and 5e) and cooler (3b, 5b and 5d) substages were identified.[2]

In the lithostratigraphic sequence of young loess profiles loess layers, solifluction, cryoturbation and other phenomena caused by permafrost are correlated with 'cold' stages. In contrast, paleosols, humous or peaty horizons, blown sand and erosion gaps are thought to correspond to the 'warm' stages of isotope stratigraphy. This procedure — supported by other informations obtained from the loess sequence — allows comparisons of local subaerial profiles with the oxygen isotope time-scale to be made *(Fig. 9).*

The oxygen isotope curve and its stages do not only reflect the changes in the mass of inland ice, glacial and deglacial periods, but also changes in sea-water temperature. The two phenomena shift in time.[3] In the carbonaceous shells of foraminifers of deep sea sediments heavy oxygen shows a somewhat higher concentration than does the adjacent sea-water. This concentration effect increases with the lowering of water temperature. Until the maximum growth of inland ice this process is prolonged with oscillations, in contrast with the relatively short period of ice melting, the **termination phenomenon** described for the first time by BROECKER and VAN DONK (1970).

On a global scale, meltwater and ocean water are relatively rapidly mixed (during ca. 2,500 to 4,000 years) and, therefore, the change in the oxygen isotope composition of ocean sediments is assumed to be a contemporaneous phenomenon. Another assumption is that the sediment sequence of deep sea basins is more or less complete and that hiatuses are not so common as in terrestrial sedimentary sequences.

Consequently, **oxygen isotope stages** are regarded **as units of standard reference** for the global correlation of marine and subaerial sediments.

Among subaerial formations complete sedimentary sequences will be preserved in small basins, i.e., in sediment traps such as the peaty layers of Grand Pile and Les Echets (MANGERUD, 1989; PONS *et al.*, 1989). They are rich in pollen and, thus, for the interpretation of their pollen spectra and for dating the time-scale of oxygen isotope stratigraphy — computed from various data using various methods — they have been used repeatedly *(Fig. 10).*

Some attempts were made to correlate the moraines of the Scandinavian ice sheet with certain stages of the oxygen isotope stratigraphy (ANDERSEN and MANGERUD, 1989). The same holds true as concerns the ^{18}O isotope temperature curve of polar ice-caps and even of the concentration of CO_2 entrapped in ice, was used for the chronological subdivision of Last Glacial oscillations *(Fig. 10).*

There do exist some doubts as to the reliability of the ^{18}O deep sea isotope stratigraphy as a standard Pleistocene time-scale for the exact dating of some continental deposits and of certain climatic phenomena. For instance, when analysing the Last Glacial loess profiles usually more lithostratigraphical units are observed than the stages and substages (altogether 12) of the deep sea sediments exist since the Last Interglacial *(Fig. 9).*

4. ***The MILANKOVITCH radiation curve and BACSÁK's calendar of climatic types.*** Today it is not yet exactly known, which physical processes are jointly reflected in the curve of $^{16}O/^{18}O$ ratio changes in the deep sea foraminifers; however, from the frequency of main and minor changes it was thought that they basically reflect the cyclical alterations of the Sun's radiation during the Quaternary according to the **astronomical theory** proposed **by MILANKOVITCH** (BERGER *et al.*, 1984).

As a consequence changes in the Earth's orbital elements and the oscillations of solar radiation (MILANKOVITCH, 1941)

[2] The enrichment of heavy oxygen isotope (^{18}O) in the skeletons of some deep sea foraminifers is hold to represent a cold stage.

[3] The preponderance of heavy oxygen accumulated in foraminifers is brought about not merely by the oscillation of inland ice.

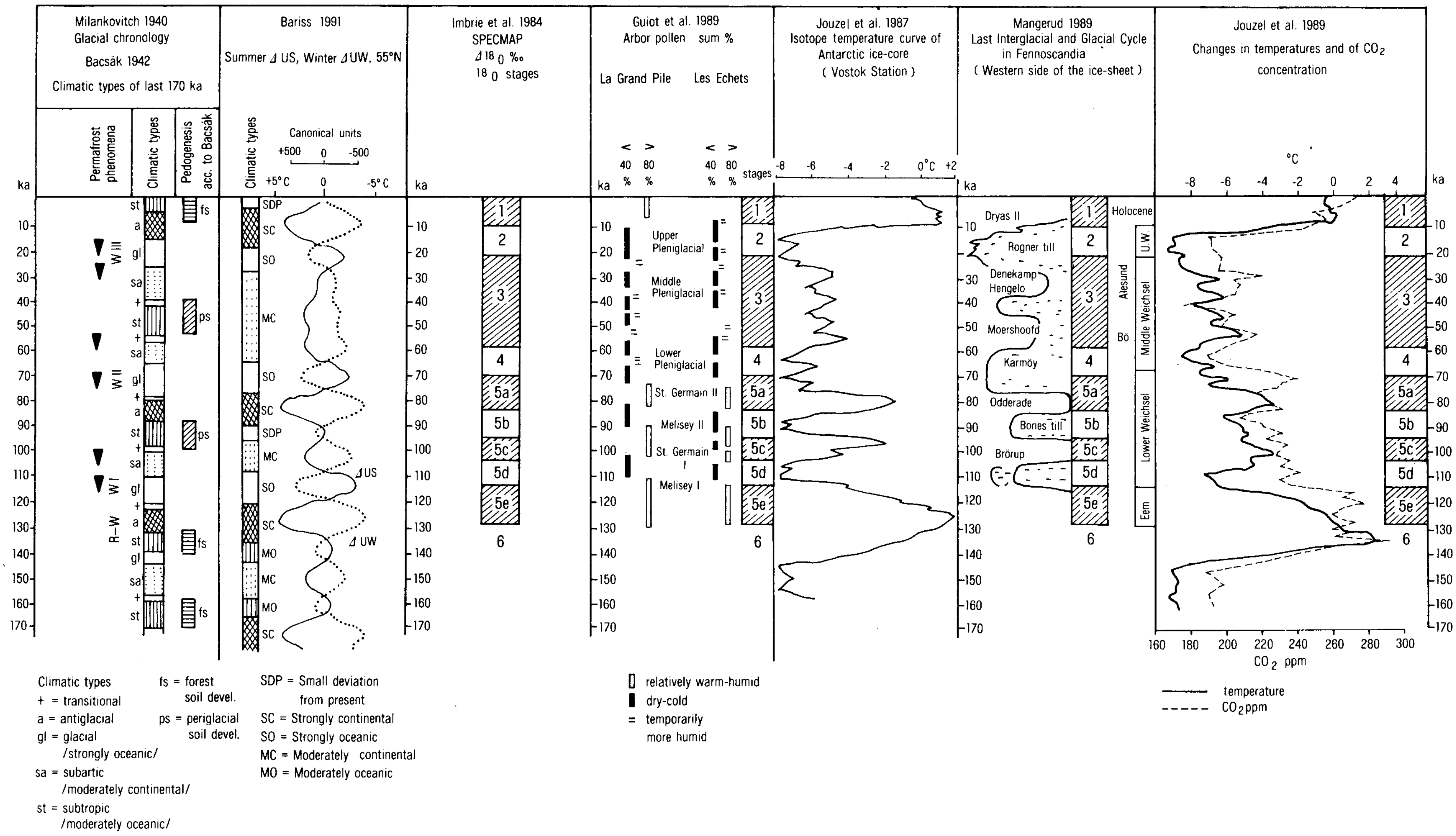

Fig. 10. Time-scales of the Last Glaciation

will have affected most directly the temperature changes of the Earth's atmosphere. In other words, the changes of radiation were felt and reactions occurred much earlier in the atmosphere than in the hydro- or cryospheres. Depending on the geographical distribution patterns of oceans and inland ice, changes of the insolation produced — by means of atmospheric and sea currents and by the situation of atmospheric centres of high and low pressure — the alternating predominance of at least four basic climatic types. These climatic types continued to alternate during the Pleistocene glacials and interglacials, too, and the duration and temporal sequence of these alternations is described by the **climatic calendar of** MILANKOVITCH's first follower, **BACSÁK** (1942). The alternation of the so-called subtropical, subarctic, glacial and antiglacial climatic types results in the existence of particular ice-free interglacials, interstadials and of the glacials or stadials. According to BACSÁK's calculation, there occurred during the last 140,000 years besides the four climatic types mentioned three times of **transitional climatic types**, too. Thus, the climatic types mentioned subdivide the Last Interglacial to Glacial cycle into 18 units lasting from the beginning of the Last Interglacial up to our days *(Fig. 10).*

Recently **BARISS** (1991) reinterpreted these Last Glacial **climatic types** into strong oceanic, strong continental, medium oceanic, medium continental and — based on MILANKOVITCH's data — he reconstructed a more or less continental climatic type with minor oscillations, regarding 55° N for the time 65,000 to 28,000 years B.P. The temperature range between summer and winter was intermediate — compared to the present-day value — for a long time. On the other hand, in BARISS' curves a transitional climatic type occurs each 20,000 years, when continental climatic types rapidly change into oceanic ones. Sometimes this went in the opposite direction (*Fig. 10).* At approximately these transitional phases minor erosional hiatuses in the loess profiles occurred. Their further investigation and inventory seems to be necessary.

In the intensively studied Central European young loess profiles the number of lithostratigraphic units and subunits in general reaches or even exceeds the number of BACSÁK's climatic types *(Figs. 9 and 10).* It is interesting to see that in the nearly 20 m peaty sequence of the Grand Pile 20 **pollen zones** were identified for the time from the Last Interglacial to the Holocene on the basis of arboreal and non-arboreal pollen. A similar value is obtained by an analysis of the temperature in the Vostok ice core, Antarctica *(Fig. 10).*

5. *An attempt to correlate Last Glacial chronostratigraphy with some typical loess profiles.*

5.1 The *young loess sequences of Western and Central Europe* are correlated with the Weichsel or Würm Glaciation. In several key sections the loess covers a well-developed, Last Interglacial, polygenetic brown forest soil. This soil and the young loess sequence are held to correspond to the Oxygen Isotope Stages 1 to 5. However, the number of horizons is different in various profiles. Therefore, the correlation of the particular loess horizons or of the several paleosols with some of the 'cold' and 'warm' isotope stages needs a cautious analysis.

The Last Interglacial paleosols (**Elbeuf I paleosol** — France; **Rocourt paleosol** — Belgium and the Netherlands) in the westernmost loess profiles of Europe are regarded as being polygenetic soils formed during the whole of Isotope Stage 5 (127,000 to 73,000 years B.P.) (LAUTRIDOU, 1982; VAN VLIET-LANOË, 1987; PAEPE *et al.*, 1990).

In other key sections of Central Europe it is suggested however that a soil complex was formed during the same time interval (**Stillfried A paleosol** — Austria, FINK, 1965; **PK III**, **B1** and **B2 paleosols** — Bohemian Basin, KUKLA and LOŽEK, 1961).

According to the interpretation of PAEPE *et al.* (1990) the Rocourt polygenetic soil complex at the type localities of Zelzate and Warneton did not develop during the Eem Interglacial (Stage 5e) proper, but soil formation continued during the relatively warm and humid substages of Amersfoort-Brörup (5c) and Odderade (5a) as well as during the intermediate colder and drier spells. In contrast to this, in the Mediterranean zone distinct paleosols (**Cala complex** of Sardinia and **Koroni complex** in Greece) were formed during the Substages 5c and 5a with intercalated eolian deposits developed during the Substages 5d and 5b.

Regarding the chronological and paleoclimatological reconstruction of the Last Glacial loess, sequences of individual key sections must be used. It cannot be the aim to develop a generalized sequence, compiled from various units of loess profiles, located remote from each other. On the other hand it is no general contradiction that — while the differences in the young loess sequences are repeatedly emphasized — similarities and opportunities for a generalization are looked for.

In the Atlantic loess region of Europe — i.e., in the type profiles of Normandy and Belgium — the Last Glacial loess is subdivided microstratigraphically *(Fig. 9)*. In Belgium the Last Interglacial paleosol (Rocourt) overlies Eem marine-lacustrine deposits in two profiles (Zelzate and Warneton — ZAGWIJN and PAEPE, 1968; PAEPE *et al.*, 1990). It was probably formed after the high sea level of the Eem Transgression. Its evolution — in a polycyclic manner — continued to the early Last Pleniglacial (73,000 years B.P.). During the ensuing pleniglacial, in the interstadial and stadial phases, 6 paleosols and 7 to 8 loess and loessy sand layers were formed. According to PAEPE *et al.* (1990) the tripartite **Warneton soil complex** (GS-5, GS-6 and GS-7) formed during the *Ognon* and *Oerel* interstadials, under basically cool to cold climates (Isotope Stage 4, 73,000 to 55,000 years B.P.).

Regarding the relatively warmer Isotope Stage 3 (55,000 to 24,000 years B.P.) 4 loess horizons and 3 paleosols (**Poperinge paleosol**, GS-8 — 47,000 years, *Moershoofd* and *Glinde* interstadials; **Hoboken paleosol**, GS-9 — 34,000 years, *Hengelo* Interstadial and **Zelzate paleosol**, GS-10 — 28,000 years, *Denekamp* Interstadial) are mentioned.

On the other hand 3 fossil soils (**Zulte paleosol**, GS-11 — 17,000 years, *Lascaux* Interstadial; **Stabroek paleosol**, GS-12 — 12,000 years, *Bölling* Interstadial and **Roksem paleosol**, GS-13 — 11,000 years, *Alleröd* Interstadial) and 3 loess horizons were formed during the Upper Weichsel Glaciation. This series is correlated with the Oxygen Isotope Stage 2.

In the Last Glacial loess profiles of Normandy and Belgium, together with the Eem paleosol, 20-21 lithostratigraphic units are identified, with at least four levels of cryoturbation and remnants of ice wedges *(Fig. 9)*.

5.2 The *young loess profiles in Central Europe, along the German Rhine and the Bavarian Danube* are different from the Western European loess sequence just described, but there are clear similarities (BRUNNACKER, 1964, 1967, 1984). In Germany, particularly in loesses along the Rhine, many profiles have been studied and based on these evidences the development and the climate history of Last Glaciation are attempted to be reconstructed *(Fig. 11)*.

The most detailed subdivision of the young loess sequence seems to have been established in the Nußloch profile, south of Heidelberg (BENTE and LÖSCHER, 1987; BIBUS and SEMMEL, 1977; BUCH and ZÖLLER, 1990; ZÖLLER *et al.*, 1988). This section can be regarded as one of the stratotype profiles of the Last Glacial loess sequence in Germany. The Last Interglacial soil — just as in Belgium — is also a polygenetic and polycyclic soil complex. Three Bt horizons are directly superimposed on each other and the TL age for the middle one was dated to 110,000±12,000 years (ZÖLLER, 1989). Soil formation continued during the Amersfoort-Brörup-Odderade substages, accompanied by slow sedimentation. Simultaneously, three humic zones developed with the intercalation of thin solifluctional loess and soil sediment. The two lower humic horizons were described as

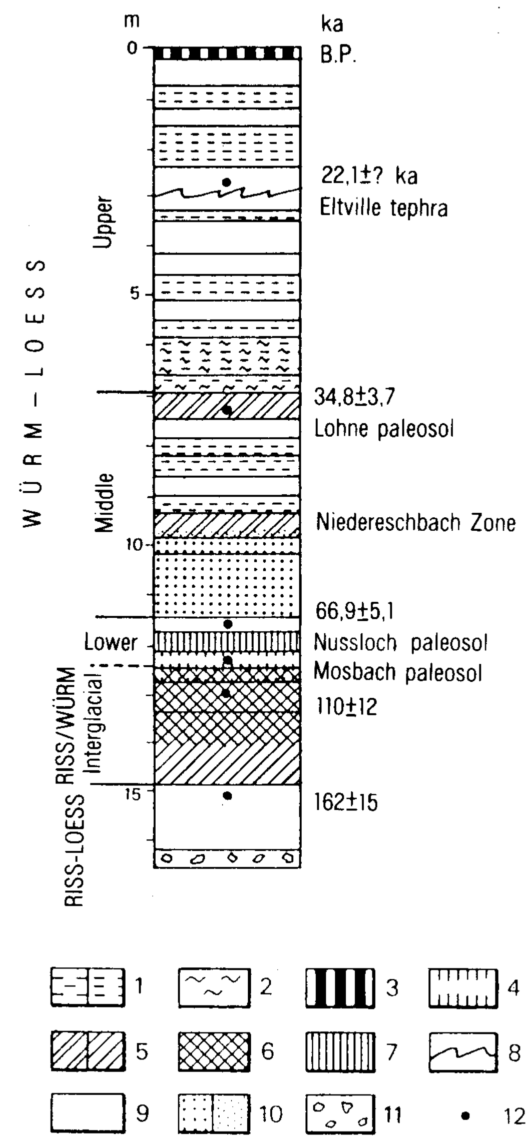

Fig. 11. Litho- and chronostratigraphy of Nußloch profile, Germany (Lithostratigraphy: according to BENTE and LÖSCHER, 1987; TL dating by ZÖLLER, 1989). 1 = Tundra gley; 2 = Solifluction; 3 = Ap horizon; 4 = Al horizon; 5 = Bv horizon; 6 = Bt horizon; 7 = Humic zone; 8 = Eltville volcanic ash; 9 = Loess; 10 = Sand/sandy loess; 11 = Limestone debris; 12 = Sites of TL sampling

Mosbacher Boden I-II, while the uppermost arctic meadow soil is called **Nußlocher Boden** in this section. The latter gave a TL age of 66,900 ± 5,100 years. These humous soil horizons are referred to the lower part of the Würm Glacial.

In contrast to this, the Warneton complex of three humous layers overlying the Last Interglacial Rocourt soil complex in Belgium is recently placed into the lower part of the Middle Würm Glacial (73,000 to 55,000 years B.P., Isotope Stage 4) by PAEPE *et al.* (1990). The chronological difference, however, may be due to the use of different TL dating techniques.

In the Nußloch profile the Middle Würm begins with a 2 metre layer predominantly consisting of sand, which presumably represents a dell fill and an erosional hiatus. The sand layer is overlain by *three faintly developed, incipient soils*, with two sandy loess intercalations. The lowermost is the **Niedereschbacher Zone** overlain by a brownish embryonic soil (Naßboden). Then two thin loess horizons are separated by Naßboden and finally the Middle Würm series is finished here by a very poorly developed, so-called **Lohner Boden,** having an age of 34,800 ± 3,700 years B.P. (ZÖLLER *et al.*, 1988).

The Upper Würm starts with solifluction loess and four or five brownish Naßböden follow with four or five slightly sandy loess horizons. In the loess between the third and the fourth Naßboden the **Eltviller tuff** (ca. 22,000 years B.P.) is observed. The uppermost part of the Nußloch profile must have been eroded since the 11,000 years old **Laacher See tuff**, present in the lower part of many paleosols covering young loess along the Rhine, is missing in this profile.

In summary, in the young loess sequence along the Rhine river the amount of paleosols and of embryonic soils (Naßböden) is 9 to 11 and loess, sandy loess and sand layers appear in a similar number. Thus together with the Eem polygenetic soil complex the number of lithostratigraphic units is also 20 or 21 here, similarly to many Belgian and French type localities, but there are also differences amongst the profiles.

5.3 Several profiles of *the Bohemian-Moravian Basin, Central Europe*, have been analysed stratigraphically in a very detailed manner (KLIMA and KUKLA, 1961; KLIMA *et al.*, 1962; KUKLA and LOŽEK, 1961; DEMEK and KUKLA, 1969; KUKLA, 1970, 1975). These authors were the first to attempt a reconstruction of the climatic history of the Last Glacial cycle and at correlation with the deep sea oxygen isotope stratigraphy.

In the Dolní Věstonice profile 10 to 11 paleosols and about the same number of loessy horizons represent — in **KUKLA's loess chronology cycle B** — the Last Interglacial and Glacial (ca. 128,000 to 13,000 years B.P.). In the young loess sequence of the Bohemian-Moravian Basin the alternation of various types of soils, solifluction loess and soil sediments is more conspicuous and, consequently, more climatic phases and shorter climatic spells can be identified than in the profiles along the river Rhine. These conditions provided the authors from Czecho-Slovakia a good opportunity to device a new and more exact classification of paleosols and of cycles in loess sequences than was possible hitherto and to reconstruct climatic conditions fairly well (DEMEK and KUKLA, 1969; LOŽEK, 1976). The similarity of loess cycles had to be explained by uniform processes over large areas.

A whole glacial cycle in the sense of KUKLA — e.g., cycle B — consists of three subordinate stadial cycles. In each of them six sedimentation or pedogenic phases are identified.[4]

These 18 phases within Last Glacial times precisely agree with BACSÁK's (1942, 1955) climatic types, which are repeated, too, three times to make up 18 units. *Figure 9* shows that — according to BACSÁK — the combination of climatic types allowed prolonged (forest) soil formation on three occasions during the Last Glacial.

5.4 *In Eastern Central Europe — in the* more continental *Middle Danubian Basin* — clearly less stratified (solifluction or slope wash) loess and soil sediments formed in the Last Glacial loess cycle than in the less continental Bohemian-Moravian Basin.

One of the stratotypes is exposed in the Mende brickyard *(Figs. 9 and 12)*. In the upper part of the profile two poorly-developed soils or humous loess horizons (h_1 and h_2, 16,000 and 18,000 years B.P., resp.), two loessy sand layers as dell fills and three loess layers occur and represent the Upper Würm Glacial. Below them follows a duplicate forest steppe-like paleosol (**Mende Upper**: MF_1 and MF_2). Charcoal from MF_1, enabled a ^{14}C dating of 27,000 to 29,000 years B.P. to be made (PÉCSI, 1987b). The TL age of the sandy loess with calcareous accumulation between MF_1 and MF_2 was 43,000 years B.P. (WINTLE and PACKMAN, 1988).

The third and fourth duplicate soil in the Mende profile is of (forest) steppe nature, too, called **Basaharc Double** (BD_1 and BD_2). Above the paleosol BD_1 there are two humous semipedolite layers followed by loess layers nos. 5 and 6, divided by dell (derasional valley) loess.

Below the BD paleosols about 2 metres of loess and a well-developed steppe soil (**Basaharc Lower**, BA) follows. At Mende the paleosols BD and BA — since they are not forest soils — were correlated for a long time with the Würm Interstadial soils. The marked brown forest soil **Mende Base** was correlated with the Last Interglacial. This subdivision was not contradictory to fauna and flora finds or previous TL and paleomagnetic datings (PÉCSI, 1982, 1987b).

The results of more recent TL datings (SINGHVI *et al.*, 1989; ZÖLLER *et al.*, 1988; ZÖLLER and WAGNER, 1990), paleomagnetic and amino acid investigations[5] call for a *reconsideration of the stratigraphic interpretation.* If the new absolute ages are correct, *the paleosol BD_2 of steppe nature probably represents in the Mende profile the Last Interglacial.* The BD soil complex occurs as a similar stratotype in several profiles of the Carpathian Basin. This means that the Last Interglacial soils in this loess region of Eastern Central Europe may not have been brown forest soils and, consequently, the contemporary climate may have differed from that in Western and Central Europe. At the same time, the well-developed Würm Interstadial soils of the Carpathian Basin (such as MF_1, MF_2 and BD_1) do not differ substantially as to their genesis from the — supposed — Last Interglacial BD_2 soil.

Consequently, according to this new interpretation the Last Glacial profiles show four well-developed soils, two young humous horizons, six loess layers, three or four dell fills (sandy loess) and two loessy semipedolites. Thus, in the Mende profile there are — including the recent soil — 21 units of a climate-indicating character *(Fig. 12)*.

5.5 The most extensive loess province of the Northern Hemisphere stretches over the central and southern parts of the *Russian Plain, Eastern Europe*. It is today cut by pronounced geographical zones. A large number of loess profiles has been described and evaluated paleogeographically for the Last Interglacial and Glacial loess cycles there.

The evolution and paleoenvironments of Valdai loesses have recently been summarized by VELICHKO (1990) and VELICHKO *et al.* (1987). In this vast *Eastern European loess region* the young loess and paleosol sequences also show differences besides the clearly existing similarities. The key profiles in the western, central and eastern parts of the Russian Plain reflect this variation *(Fig. 13)*. The reconstruction of the Last Glacial climatic changes was based on the paleogeographic evaluation of layers in the type sections of the central zone of the Russian Plain.

For the reconstruction mentioned stratigraphical data and complex physical and paleogeographical investigation methods obtained and applied over decades were used. In the Russian Plain the loess mantle is almost continuous south of the Lvov-Kiev-Ryazan line. To the north of this line, up to the limits of the former Valdai ice sheet, loess is only of sporadic occurrence, but larger areas are found in the environs of the towns Minsk, Vladimir and on the Smolensk-Moscow Uplands. In the southern portion of the Russian Plain loess forms an extensive cover, up to the Black Sea coast and the Caucasus foreland (see map on page 51). In this belt the loess-paleosol sequence is much thicker and there also are major differences in the appearance of paleosols than in the central zone of the Russian Plain.

At the base of the Valdai loess a well-developed, polygenetic soil complex, the **Mezin complex**, is found. It is the widest spread paleosol of the Russian Plain. In the profiles two phases of the formation of this soil are identified: the earlier *Salyn* phase corresponds to the Mikulino Interglacial. In this period over most of the

4 1. Slope deposit accumulation; 2. Forest soil formation; 3. Chernozem formation + semipedolite and eolian sediment accumulation; 4. Marker loess — 20-30 cm fine loess; 5. Loamy semipedolite formation; 6. Loess formation + pseudogleyification.

5 In addition, the amino acid analysis of a gastropode fauna collected from the loess (l_6) overlying paleosol MB at Mende allows the conclusion — according to McCOY — that this forest soil is older than the Last Interglacial. According to ZÖLLER's recent TL dating the ages of loess layers l_4 and l_5 enclosing paleosol BD_2 at Mende are 114,000 years B.P. and 144,000 years B.P., respectively.

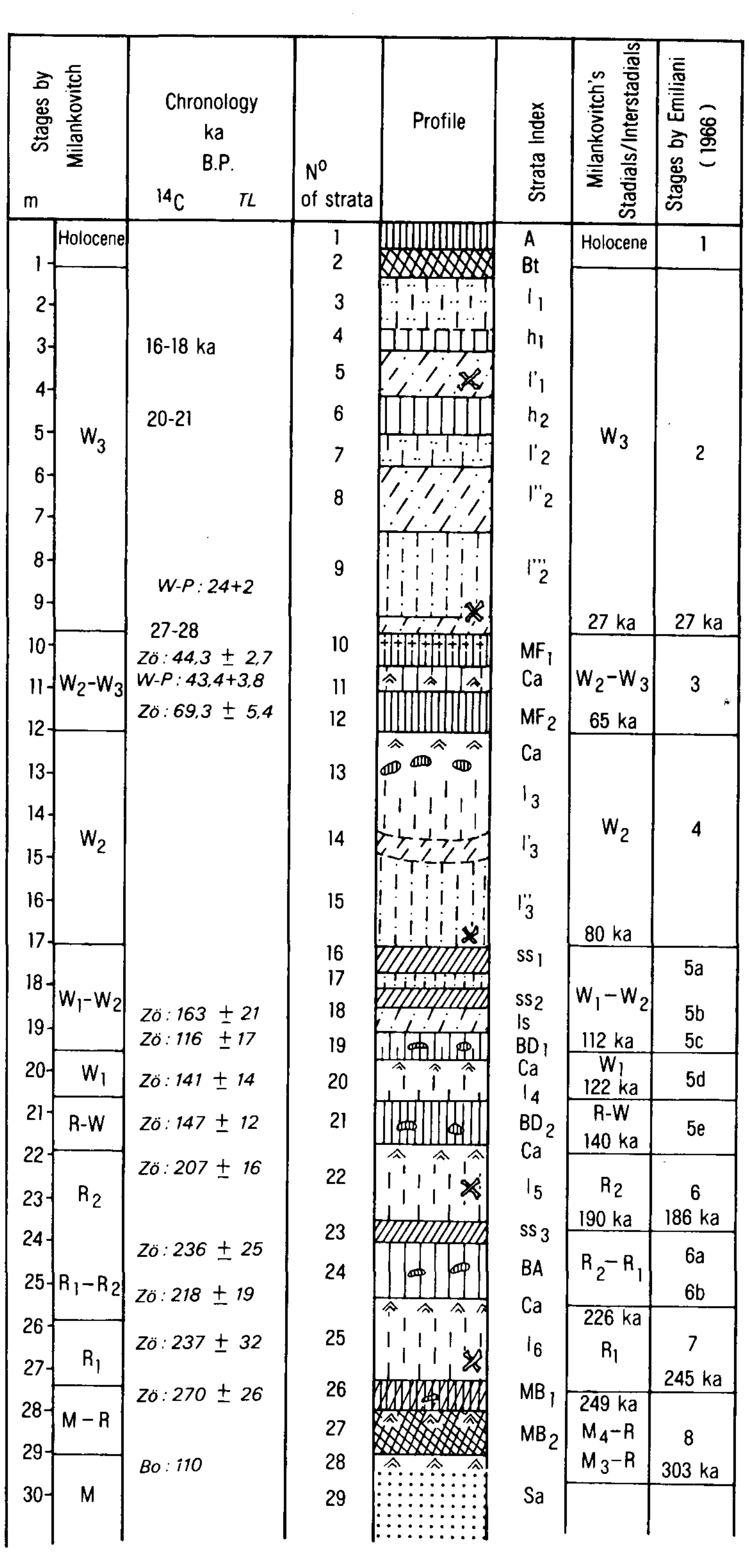

Fig. 12. Litho- and chronostratigraphy of the loess profile at Mende in Hungary (reconstruction of cyclical climatic changes on the basis of events in the loess-paleosol sequence, according to Pécsi, M.). ^{14}C data: Lab. Hannover; TL dating: Bo = BORSY *et al.*, 1979; W-P = WINTLE and PACKMAN, 1988; Zö = ZÖLLER, 1989-1991; Magnetostratigraphy: MÁRTON, 1979; PEVZNER, 1979-1990; many samples of the profile were investigated and only normal polarity was found. 1 = Chernozem, steppe dynamics, afforestation in the Late Holocene, cultivation, soil erosion, dell (dry valley) infilling; 2 = Brown forest soil (Bt); 3 = Sandy loess formation (l_1); 4 = Loessy humus (h_1), embryonic soil, taiga parkland with charcoal of *Picea, P. cembra*, end of dell infilling; 5 = Sandy loess (l'_1) with reindeer remnants, dell incision and infilling, permafrost; 6 = Loessy humus (h_2), embryonic (arctic) soil, taiga grove, charcoal; 7 = Sandy loess (l'_2); 8 = Dell loess (l''_2), dell incision and infilling with residual permafrost; 9 = Sandy loess and typical loess (l'''_2), complete skeleton of *Elephas primigenius*; 10 = Steppe soil (MF_1), cold-steppe taiga groves with much charcoal of *Larix, Picea* and *P. cembra*; 11 = Thin loess, strong carbonate accumulation under MF_1 soil, erosional hiatus; 12 = Better-developed grove steppe soil (MF_2) with charcoal (*Picea*). The paleosols MF_1 and MF_2 are not identical with the present-day soil on which the original association was *Prunetum tenellae* or *Acereto Tatarici -Quercetum*; 13 = Thick, triplicat loess horizon (l_3); 14 = Dell loess, slope loess (l'_3); dell incision and infilling; 15 = Sandy loess (l''_3), remnants of *Elephas Primigenius*; 16 = Semipedolite, soil sediment (ss_1); slope wash and solifluction; 17 = Sandy loess (l''_3); 18 = Semipedolite, soil sediment (ss_2), and solifluction loess (l_s); 19 = Steppe soil (BD_1), with *Betula, Pinus* and *Artemisia* pollen (URBAN, 1984); 20 = Sandy loess (l_4); 21 = Steppe soil (BD_2) with medium carbonate accumulation, predominant pollen are *Pinus, Betula* and *Artemisia* (URBAN, 1984); 22 = Loess (l'_5), remnant of *Elephas primigenius*, probably belonged to Riss 2 Glacial stage; 23= Soil sediment (ss_3), slope wash and steppe soil formation; 24 = Steppe soil formation (BA) with a strong carbonate accumulation horizon, predominant pollen are *Artemisia* > *Cerealia* typ (URBAN, 1984), warm temperate climate, moderately dry steppe condition; 25 = Loess (l_6), remnant of *Equus sp.*, probably Riss 1 Glacial stage; 26 = Steppe soil formation (Mende Base, MB_1) with many krotovinas, this part of the soil complex probably developed during a transitional steppe climate between mediterranean xerophile brown forest soil and loess steppe conditions; 27-28 = Brown forest soil (MB_2) with $CaCO_3$ nodules in the Bt horizon and very strong Cca horizon with big loess dolls (28), in MB_2 the predominant pollen are *Pinus* > *Picea, Chenopodiaceae* (URBAN, 1984), warm temperate climate with dry and wet seasons; the MB paleosol complex probably developed during the upper part of Mindel-Riss Interglacial stage; 29 = Proluvial sand, TL dating is a minimum age, however, underestimation is possible.

The TL ages of the paleosols BD_1 and BD_2 seem to be somewhat too high since these values — according to the different time-scales — indicate stadials instead of interstadials (see *Figs. 9-10*). A somewhat similar problem of calculation exists with the age determined by ZÖLLER for the paleosol MF_2. Both, by the SPECMAP time-scale and by the MILANKOVITCH time-scale, climate was cold between 59,000 and 71,000 years B.P. Consequently, these periods were less suitable for soil formation.

Russian Plain a forest soil developed, which shows distinct eluvial and illuvial horizons. In the Mezin key section the paleosol overlies the **Dnieper moraine** and it is covered by **Mikulino peat.**

During the younger pedogenetic *Krutitsa* phase a chernozem-like soil was formed *(Fig. 13).* It is regarded as the first interstadial of the Valdai Glaciation and it is correlated with the Brörup or Amersfoort Interstadial. The two soils of the Mezin complex formed in two warm phases are divided from each other by a thin loess horizon, which dates from a cold spell. This is called the **Sevsk or intra-Mezin loess horizon.**

The whole Mezin soil complex was deformed by permafrost during two phases of the so-called *Smolensk cryogene stage.* Within the interglacial soil the permafrost phase "A" is evidenced by minor earth- and ice-wedge pseudomorphs, while during phase "B" solifluction and cryoturbation deformed the entire Mezin soil complex, including the **Krutitsa soil**. Thus, these frost phenomena date from times after the Lower Valdai Interstadial.

The first major loess unit of the Valdai Glacial, the **Khotylevo loess** is, in fact, the parent material of the **Bryansk soil** *(Fig. 13).* It is considered to be an interstadial formation, dated by the radiocarbon method to 32,000 to 24,000 years B.P.[6] Pedologically, it is an arctic, cryogene, gleyic soil with predominance of *Betula* pollen, typical of the present-day permafrost zone. In the Desna Basin (Arapovichi key section) a tundra steppe fauna was found in this paleosol. The Bryansk paleosol is correlated with the **Kesselt paleosol** in Belgium, **Gleina** in Germany, **Stillfried B** in Austria, **PK I** in the Bohemian-Moravian Basin and **Mende F₁** in Hungary.

After its formation, during the so-called *Vladimir cryogene phase*, the Bryansk soil was transformed into a spotted skeletal soil with nonsorted circles. This was followed by the formation of the **Valdai loess** II (Desna), i.e., 3 to 4 m of typical loess.

The **Trubchevsk embryonic soil** or gleyed horizon, which formed in a warmer interval of the Late Glacial, ca. 17,000 years B.P., divides the units Valdai loess II and III (Altynovo) from one another. In the latter loess the sand fraction has a strikingly higher percentage than in loess II. The Valdai Glacial was closed by a very strong cryogene phase and in the loess profiles deep ice wedges and cast pseudomorphs are preserved. From this it may be concluded that a periglacial climate resembling the present environments of Eastern Siberia prevailed. On top of Valdai loess III the recent soil formed, and its maximum age is according to the radiocarbon method 9,000 to 8,000 years B.P.

6 The radiocarbon data allow conclusions for the formation of two different soils or the prolonged development of a polygenetic soil.

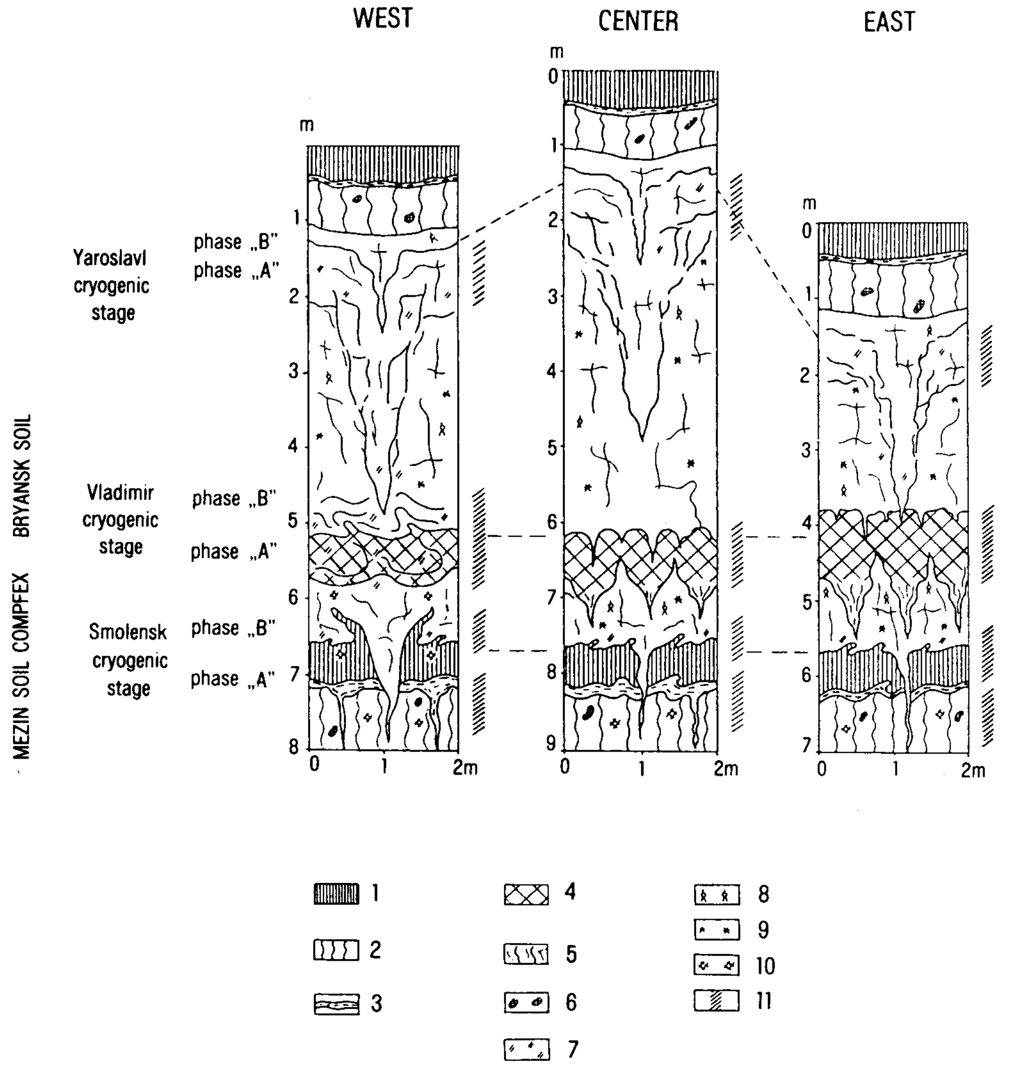

Fig. 13. Subdivision of the Last Glacial loess profiles in the Russian Plain (according to VELICHKO, 1990). 1 = Humus horizon of recent and paleosols; 2 = Illuvial horizon of recent and paleosols; 3 = Elluvial horizon of recent and paleosols; 4 = Weakly humified loam of paleosols; 5 = Loess; 6 = Krotovina; 7 = Gleyization; 8 = Carbonates; 9 = Trace of cement ice; 10 = Trace of icelens; 11 = Presumed ancient active layer of permafrost

In the Mezin soil complex and Valdai loess series — as described above — together 14 to 15 paleogeographic events, i.e., climatic oscillations, can be revealed.

5.6 In *Western Siberia* an extensive and thick (60-100 m) loess mantle covers the Ob Plateau. The loess bluff section along the Ob river from Barnaul to Novosibirsk has been studied in natural outcrops along many kilometres (SHAEVICH, 1984, 1987; DERGACHEVA and ZYKINA, 1988). The most typical Upper Pleistocene loess-paleosol sequence is found in the upper (8 to 12 m) parts of the Lozhok profile at Novosibirsk and Belovo profile at Barnaul *(Fig. 9)*. The Last Interglacial is represented by a double forest steppe soil, the **Berdsk soil complex** (br_1 and br_2), correlated in Siberia with the *Kazantsevo* Interglacial (VOLKOV and ZYKINA, 1984).

The lower (br_1) dark humous soil (A and B horizons) was affected by an intense cryogenic impact, it is divided by ice wedge and cast pseudomorphs. This is overlain by a veneer of loessy loam (suglinok) with krotovinas and fine calcareous precipitations. The upper paleosol (br_2) is similar to br_1, but thinner and poorly-developed in some profiles. It essentially represents the humus (A) horizon. The duplicate soil was deformed subsequently by permafrost phenomena and it was dissected by ice wedges. The older member (br_1) of the *Berdsk soil complex* is a forest steppe or *steppe soil, which was probably formed under a strongly continental, temperate climate.* It is assumed that the suglinok enclosed between the two soils developed in the cold stadial (Ermakovo I), the beginning phase of the *Zyryanka* Glacial, while the paleosol br_2 dates from the first interstadial (Bogdashkino) of early Last Glacial.

The overlying loamy loess, loess-like sandy loam (Tulin loess) is some metres thick, in some profiles a two-storeyed horizon. It may represent the Lower Zyryanka of the complex pleniglacial (Ermakovo).

The **Iskitim soil complex** (IS_1 and IS_2) consists, too, of two humic steppe soils (with A, B and C horizons), but they are obviously much more faintly developed than the Berdsk soils. Radiocarbon dating gave ages of 33,000 to 32,000 years B.P. for IS_1 and of 26,000 to 20,000 years B.P. for the IS_2 paleosol. The ^{14}C values for humus, however, are only minimum ages.

Consequently, the steppe soils IS_1 and IS_2 must have been formed during the Late (29,000 to 25,000 years B.P.) and the Early **Karginsky Interstadial** (Middle Zyryanka), (50,000 to 30,000 years B.P.), respectively.

The formation of **Sartan loess** (25,000 to 10,000 years B.P.) is placed into the Last Glacial maximum and that of the **Elcov** or **Bagan loesses** into the Late Glacial.

In spite of all the differences mentioned, the above described Siberian loess profiles show much lithostratigraphic resemblance to the sequences of the key sections in the central Russian Plain. During the Last Interglacial paleoclimate was essentially different here from that on the Ob Loess Plateau. On the latter meadow and chernozem steppe soils alternated in time with a cyclical development of suglinok and permafrost phenomena.

5.7 *The loesses of Central and Inner Asia* are distributed along the margins of the Asian orographic deserts, on pediments of high mountains and in basins in long belts stretching from the Loess Plateau of China to the Turkestan Basin. In this belt the so-called **warm loess** has formed in a great thickness and cryogenic phenomena are of subordinate role or they are entirely absent. During the Pleistocene arid, semiarid and continental conditions alternated with pluvial phases in this zone. This was interpreted by some authors such that loess had formed during the interglacials and paleosols in the glacials. Yet recently, another group of scientists proposes that loess formed in the glacials as elsewhere and paleosols are mostly associated with the interglacials (DODONOV, 1986; LAZARENKO, 1984; LIU, 1985). During phases of transitional climate, however, reworking of loess and of paleosols, sand intercalations and degradation of loess horizons may have also taken place.

5.7.1 In the high basins of *Tajikistan* the *Chashmanigar, Lakhuti* and *Kairubak* profiles, and in *Uzbekistan* the *Orkutsai* profile (DODONOV, 1984, 1986; LAZARENKO, 1984) show almost identical sequences.

Relying on information from these profiles and on other Tajik profiles, DODONOV refers to five paleosols and five loess layers during the Upper Pleistocene period *(Fig. 8)*. Based on paleomagnetic and TL data, he places the paleosol V into the Last Interglacial (Mikulino).

In the overlying **loess 5** the Blake event was found. The paleosol IV is associated with the upper part of the Mezin soil complex (Brörup, Amersfoort interstadials). The soil complexes IV and II, with the enclosed **loess** layers **4** and **3**, are held to belong to the middle part of the Last Glacial. The **loess** layers 4, 3 and **2** are correlated with the cold Isotope Stages 4, 3 and 2. The soil complex II is regarded to be of the same age as the Bryansk soil (30,000 to 26,000 years B.P.). DODONOV considers **paleosol I** and **loess 1** to be of Late Glacial age. The profile is terminated by the recent greyish cinnamon soil. In the Tajik Basin the paleosol I is a poorly-developed serozem, **paleosol V** is a well-developed red-brown forest soil, while the **paleosols II, III** and **IV** are brown-red cinnamon soils (brown-red kastanozems, i.e., chestnut soils).

5.7.2 From the *Loess Plateau of China* hardly any paleosol has been described from the Upper Pleistocene **Malan loess**, only the **paleosol**, called S_1, at 10 to 12 m below the surface, was referred to the Last Interglacial (LIU, 1985). The young sequence of the loess plateau, lying much higher than 1000 m a.s.l., is assumed to be incomplete in many places. Recently, paleosols are described from the 30 to 40 m Malan loess on the second terrace of one of the Hwang Ho tributaries in the vicinity of the town Linxia *(Fig. 9)*. At the bottom there is a three-storeyed, poorer developed soil complex (S_{1-b} to S_{1-d}), which is not older than 137,000 years B.P. according to TL investigations and the calculated rate of sedimentation (LI *et al.*, 1989).

The ages of the paleosols in the loess overlying the Beiyuan platform are calculated by LI *et al.* (1989) in a complex manner. The magnetic susceptibility curve for the profile was compared with the stages of isotope stratigraphy and fauna data and TL results were used *(Table 6)*.

Table 6. Correlation of paleosols of the Beiyuan profile with the oxygen isotope stages

Stages; ka B.P.	Beiyuan profile	^{18}O Isotope Stage
Holocene	S_0 Paleosol	1
Upper Glacial (30-10)	S_{m-1}	2/3
	S_{m-2}	3a
Middle Glacial (85-30)	S_{m-3}	3c
	S_{m-4}	4/5
Lower Glacial (110-85)	S_{1-a} S_{1-b}	5a
Last Interglacial (140-110)	S_{1-c}	5e
	S_{1-d}	5/6

Finally, the cyclical changes of the isotope temperature curve of the Vostok research station, Antarctica, valid for the last 160,000 years, were compared with the loess-palsolsequence of the Beiyuan profile. The loess horizons were correlated with the colder peaks and the paleosols with the relatively warmer peaks of the temperature curve (LI *et al.*, 1989). The conclusion was that regarding the Last Glacial chronology of loessy deposits the mentioned isotope temperature curve provides more reliable information than the comparison with the deep sea isotope stratigraphy curve or stages.

CHEN *et al.* (1991) proposes that the isotope temperature curve calculated from the Antarctic ice core should be the basic time-scale for the dating of young loess-paleosol sequences.

If the chronological subdivision suggested by LI *et al.* (1989) for Last Glacial loess-paleosol sequences could be confirmed by several profiles on the Loess Plateau, the climatic history of the Malan loess series (9 loess layers and 8 paleosols) would become more consistent with the subdivision of other Eurasian loess regions.

5.8 The most extensive loess region of *North-America* is situated in a broad west to east belt along the Missouri, Mississippi and Ohio valleys and to the south it forms a narrower zone almost reaching the delta in the Lower Mississippi valley. The huge north to south extension (more than one thousand kilometres) covers the area from the limits of the *Wisconsin* ice sheet (ca. 41° N) to the latitude of the Gulf of Mexico (30° N).

The recent results of young Pleistocene loess stratigraphy have been published and presented during field-trips by the AQA and the North-American Working Group of the INQUA Commission on Loess (FOLLMER, 1978, 1979; FOLLMER et al., 1986; McCRAW and AUTIN, 1989; McKAY and STYLES, 1986).

The Last Interglacial soil is the **Sangamon polygenetic paleosol** *(Fig. 9)*. In its vast distribution area this soil complex displays many varieties. It was formed before the Wisconsin sediments were accumulated on Illinois (Loveland) loess, till and occasionally on older fluvial deposits *(Fig. 9)*. The age of its stratotypes is 130,000-75,000 (?)-50,000 years B.P. Soil formation was prolonged and interrupted repeatedly since it was not always able to keep pace with the accumulation of mineral material. In fact, the storeyed soil complex in south Illinois has a relatively well-drained profile with red-brown to dark red Bt horizons (in some places with A, E and EB horizons, as well; FOLLMER, 1982).

In many places its formation was not restricted to the Sangamon Interglacial only, but — mainly to the south, along the Lower Mississippi valley — it began already in the Late Illinois (ca. 198,000 years B.P. — McKAY and FOLLMER, 1985) and continued to the Early Wisconsin.

The overlying **Roxana silt** — as a lithostratigraphic unit — is in fact not loess, but a sequence of slightly greyish brown embryonic soils and silts. In Illinois it is subdivided into four members (r_1 to r_4, see *Fig. 9)*.

Roxana silt as a whole gave ages of (76,000 or) 65,000 to 30,000 years B.P. and the beginning of *Wisconsin* Glacial is defined by its start.

The **Farmdale soil** is a leached greyish brown loam, partly developed in the upper part of the *Roxana* or **Robein silt** and occasionally in the **Peoria loess**. Its ^{14}C age is 28,000 to 25,000 years B.P. Chronostratigraphically, this interval is the *Farmdale* Interstadial.

The *Peoria loess* (corresponding to the *Woodford* Stadial — 25,000 to 8,000 years B.P.) is subdivided into five parts (zones 1 to 5) on the basis of clay mineral composition. In the uppermost zone (p-5) the dark flood-plain clay layer — **Jules paleosol** — was formed at about 16,000 years B.P. Below this soil thin clay and sand layers occur. They represent fluvial, flood-plain deposits (20,000 years B.P.).

In Illinois the stratigraphic units of Roxana silt and Peoria loess are contemporaneous with six soil horizons, among which the Farmdale paleosol is the most prominent, while the others are more faintly developed but show a soil structure. Therefore, it is assumed that the repeatedly occurring initial soil formation of various intensity can be associated with a not too rapid accumulation of silt and a somewhat warmer and moister climate of (inter)stadials. The number of loess, paleosol and sand layers during the Wisconsin Glaciation is about 12, not counting the horizons of the Sangamon soil complex locally including parts of the Wisconsin.

The present-day soil as a Holocene Interglacial formation is not the same in a given point as the Sangamon soil of the Last Interglacial. The time factor was of great importance during the formation of the latter.

The present-day prairie soil — on Peoria loess — is a marked, polygenetic formation. While the comparably polygenetic, storeyed Sangamon forest and meadow soil formation lasted for a long interval particularly in the Lower Mississippi valley, where some authors hold that it comprised the time of the Late Illinois, Sangamon, Eowisconsin and the Early Wisconsin. Others propose loess formation within the Early Wisconsin.

6. *Conclusions.*

6.1 For the identification of the *duration* and subdivisions of the *Last Interglacial/Glacial cycle* numerous attempts have been made. Although the different time-scales are recently being correlated with each other, using a paleogeographical reconstruction of young loess-paleosol sequences, the differences of time-scales, e.g., between the ^{18}O isotope stages (2 to 5) and the glaciated and ice-free stages by MILANKOVITCH, are evident. The most important discrepancy is found in the duration of the Last Interglacial (Riss/Würm or 5e) and its place in the time-scale.

On the MILANKOVITCH and BACSÁK time-scales the R/W Interglacial can be placed in the interval between ca. 140,000 to

120,000 years B.P.,[7] while the stage 5e is dated to 128,000 to 116,000 years B.P. *(Figs. 8 to 10).* The two time-scales, frequently applied for a loess-paleosol stratigraphy indicate a difference of ca. 20,000 years for the age of the Last Interglacial. If the longer interval is assumed for the R/W Interglacial, the formation of polygenetic soils would be more easily interpretable.

The difference between the duration and position on the time-scales of oxygen isotope stages 2 and 4 on the one hand and stadials W_2 and W_3 on the other does not seem to be more than 5,000 years. The prolonged interstadial (66,000 to 26,000 years B.P.) between two glacial stages gave sufficient time for the formation of double or storeyed soils, the younger of which (29,000 to 25,000 years B.P.) is of almost world-wide distribution.

The most widespread and thickest young loess layers date from stadial W_3 (i.e., oxygen isotope stage 2), between 24,000 and 12,000 years B.P. Under the prevailing glacial climate of this period cool-humid and cold-arid spells alternated. Until the beginning of the Holocene only two or three poorly-developed humous arctic soils formed in the periglacial loess zone.

6.2 *The paleoclimatic reconstruction of the Last Interglacial/Glacial cycle* is hindered by differences between various key sections (differing numbers of paleosol layers and loess horizons, paleocatena variation and hiatuses). Therefore, either a general reconstruction can only be given or it has to be based on a particular loess-paleosol profile which includes most of the stratigraphic units present in key sections.

In each case a fundamental task is to identify the Last Interglacial soil. Because of the few and uncertain absolute datings available, this is even now only possible by the use of comprehensive, indirect approaches.

With some restrictions we assume that in the selected type localities *(Fig. 9)* the Last Interglacial soil is identified.[8] With this in mind, the general statement can be made that in most of the key profiles between the Last Interglacial and the recent soils there are 5 or 6 (among them two or three poorly-developed) paleosols and 5 to 7 loess or sandy loess horizons. In the layers mentioned and between them — especially in the former periglacial zone — severe cold climate is indicated by *permafrost pseudomorphs* in 3 to 5 levels (*Fig. 9* — columns 3, 4 and 11*)*.

In addition, in some loess regions (such as in Eastern Central Europe, North America, the Columbia Plateau and the Tashkent loess in Central Asia) *buried dells* in two or three levels are observed in the Last Glacial loess layers. Dell development could take place in cool-moist climatic spells, simultaneously with the formation of embryonic soils.

The number of climatic phases which can be reconstructed from the loess-paleosol sequence mentioned and the enclosed phenomena (dells, cryoturbations, solifluction and erosion) of climate-indicating role is 16 to 20. These can be interpreted as having happened at least 16 to 20 changes in the climatic conditions, necessary for the development of the layers and phenomena mentioned, over the time span when the young loess-paleosol sequence was formed (130,000 to 10,000 years B.P.). These climatic phases of 2,000 to 10,000 years duration are partly composed of climatic oscillations of shorter duration and partly made up by higher-rank cycles of 20,000 to 40,000 years length (on three occasions stadial and interstadial and on one occasion interglacial paleoenvironments recurred, excluding the Holocene).

6.3 The sequence of some loess profiles may be *quasi-complete*, embracing a series from the Last Interglacial to our days. Still, according to their paleoenvironmental position, they show various dimensions in their sequences. For an approximative reconstruction of climatic changes — in our opinion — a particular sequence of a loess region has to be investigated, paying due attention to results obtained from similar profiles in the area under study *(Fig. 12).*

6.4 Within some loess regions young loess mantles may occur which formed exclusively during the Last Stadial, W_3 (24,000 to 12,000 years B.P.). In such loesses — which can attain thicknesses of 4 to 10 m — only embryonic soils, two or three humous loess horizons, cryoturbational phenomena and dells filled by sandy loess occur (MAROSI and SZILÁRD, 1988; PÉCSI, 1982). Loess profiles most suitable for the reconstruction of the Last Stadial climatic oscillations are usually expected to be found in the sections of dells of various size.

For a more exact characterization of the Last Stadial climatic fluctuations in loess profiles of the Great Hungarian Plain SZÖŐR *et al.* (1991) used a **malaco-thermometer method.** This method can be used for the determination of the climatic optimum and also of July mean temperature (JMT) by the existence of several species. Simultaneously an absolute dating was carried out in the Institute for Nuclear Physics, Debrecen using Gastropoda shells and applying the ^{14}C method *(Table 7).*

Table 7. The climatic fluctuation during the Last Stadial in the Great Hungarian Plain. Based on Gastropoda shell analyses at the brickyards of Debrecen and Lakitelek near Szolnok (constructed by P. SÜMEGI)

^{14}C age of the molluscs ka BP	Loess and paleosol sequence, existence of several species, and the July mean temperature (JMT)
5 – 10	JMT: up to 20 °C; *Cepaea vindobonensis* with climatic optimum about 7,000 years B.P. as a thermoxerotic phase with *Granaria frumentum, Pupilla triplicata, Helicopsis striata* dominance
10 – 12	JMT: 16-17 °C; *Succinea oblonga* dominance peak
12 – 14	Sandy loess; decrease of cryophylous taxa to be correlated with Bölling Interstadial
14 – 16	Loess, sandy loess; cool phase with JMT: 13-14 °C; last appearance of *Pupilla sterri*; dominance peak of *Columella columella, Vallonia tenuilabris*
16 – 18	Loess and humic loess; a short thermohygrotic phase (*Vestia turgida, Punctum pigmaeum*) followed by a massive cooling correlated with Gravettian sites
18 – 20	Loess; abrupt cooling down to JMT: 12 °C (*Columella columella, Vallonia tenuilabris*)
20 – 22	Loess phase and humic loess; phase of warming-up (*Pupilla muscorum*) to be correlated with the lower part of Last Pleniglacial loess series
22 – 24	Loess; cryoxerotic phase (*Vallonia tenuilabris*); JMT: 12-14 °C
26 – 30	Thermohygrotic phase to be correlated with paleosols MF_1, Stillfried B, Kesselt, with JMT: 17 °C

7 The isotope temperature curve for the Vostok station shows strong warming in a very similar period.

8 In the opinion of some of the authors, the Last Interglacial soil was formed over a longer time of warmer and moister climate and it is better developed than the present-day surface soils or other paleosols in young loess (FINK, 1974; GERASIMOV, 1973). The statement that the Last Interglacial soil is of the same nature as the recent soil in the given locality (BRONGER and HEINKELE, 1989) is an oversimplification of reality or founded on a misinterpretation of nomenclature.

DOMINANT GEOMORPHIC PROCESSES DURING THE MAXIMUM COOLING OF THE LAST GLACIATION

by I. I. Spasskaya

(Explanatory notes to map on page 53)

The map of dominant geomorphic processes was based on the concept of the climatic control over the exogenous relief-forming processes. Previously an idea had been developed for the present-day relief formation. It resulted in the elaboration of a system of climatic-geomorphic zones and of zonal **types of morphogenesis** (BÜDEL, 1977; WILHELMY, 1975; DEDKOV *et al.*, 1977, and others). Some of the works were illustrated by geomorphological maps, although small in scale and schematic, only.

Few attempts, however, were made to outline the total range of relief-forming processes which might have operated in a certain region and within a certain time period of the past. No attempt was made to present a detailed scheme of the spatial distribution pattern of types of processes in the past either on a global or on the subglobal scale. Thus the map presented here is the first approach which undoubtedly should be further developed and specified during future investigations. It was based on numerous regional and Quaternary studies which had been summarized partly in monographic volumes and atlases (GERASIMOV and VELICHKO, 1982; WRIGHT, 1983; VELICHKO, in press).

When compiling the map the author used the above mentioned concept about the *types of morphogenesis controlled by climate*. At the same time it was attempted to identify some elementary processes (as well as the resulting landforms and/or sediments) which were particularly typical of a specific type of morphogenesis and which could be used as diagnostics.

The *best known processes* are those having been induced by *ice sheets* and *alpine glaciers*. They have been discussed at length by several authors and need not be considered here. The second group which includes landforms and structures which indicate periglacial conditions (or at least the presence of perennially or seasonally frozen ground), is the most numerous and well defined. The landforms and structures are of interest, because they provide rather precise information on paleoenvironments. For example, cryoturbation or solifluction structures, aside from indicating permafrost (or deep seasonal freezing) also attest to relatively high humidity, the dimensions of ice wedges can be directly interpreted in terms of ground temperatures, etc.

Processes which occur *in arid climates* produce conspicuous landforms: fossil sand dunes, desert pavements and ventifacts, deflation hollows. All of them are usually well preserved long after the climate had changed.

Identification of *landforms* associated with *more humid climatic conditions* is less unambiguous. The evolution of the environment is usually recorded in slope deposits and fluvial terraces, which reflect alternating phases of accumulation and of erosion.

When mapping the paleogeomorphic processes special attention was paid to information about the above mentioned fossil landforms, and especially to their chronological characteristics. On the map they generally represent the "spheres of influence" of different patterns of environment with specific types of morphogenesis.

The most striking feature of the glacial maximum of the Late Pleistocene (20,000 to 18,000 years B.P.) is the wide distribution of glaciers, first of all of ice sheets which completely dominated the landscapes and essentially modelled their beds over vast areas in high and middle latitudes. The **glacial type of morphogenesis** is shown on the map within the limits of the Late Pleistocene ice sheets on plains, and also comprises mountain glacier complexes which had been formed by an intricate system of valley and transfluent glaciers, piedmont glaciers, ice caps, etc. The pronounced topographic gradients determined active glacial erosion and deposition which left noticeable traces in the present-day topography. Areas of separate small valley and cirque glaciers shown on the map are not included into the glacial morphogenesis type since the areas of the glaciers are relatively limited and the topography was controlled by other processes.

The **periglacial type of morphogenesis** was even more widespread than the glacial one. Periglacial processes dominated vast areas of the continents which now belong to the temperate zone. Their scope was mostly controlled by a cooler and drier climate compared to the present-day one. It should be noted that a *combination of cryogenic and eolian processes* is recorded in many regions within this zone. The importance of the eolian processes seems to have increased eastward and southward. The periglacial forms themselves also indicate some regional differences which resulted from differences in the continentality of the climate. The more "oceanic" climate of Western Europe was favourable for cryoturbations, involutions and solifluction, whereas the Russian Plain and Siberia preserved numerous traces of more arid and severe environments described by VELICHKO (1973) as "tundra-steppe". Morphogenesis within this zone was controlled predominantly by low winter temperatures, high continentality, relative aridity and mostly low intensity of rainfalls, as well as by the scarcity of vegetation. Under these conditions cryogenic processes were most important, although those requiring a high soil moisture had developed only locally. Frost fissure formation seems to have been prevalent. It created regular polygonal nets and preconditioned subsequent geomorphic processes. Under conditions of low rainfall intensity, surface runoff could not develop very active, though the rain water was probably able to create deep gullies on valley slopes insufficiently protected by vegetation.

The sheet and rill wash accounted for some short-distance transport of loose material at watersheds, for example of small windborne particles, which were not yet fixed by vegetation. Surface runoff increased locally due to the accumulation of snow transported by wind and its melting in the spring. Under extremely low winter temperatures the snow had a very loose micro-crystalline structure and could easily be carried away by wind and accumulated in sheltered places. Snow melting produced not only sheet wash but solifluction as well, which induced further development of linear hollows, dells, derasion valleys. PÉCSI (1969) introduced the term *derasion* to denote the joint effect of gelifluction, cryoturbation, pluvionivation, cryofraction and gravitational movements under periglacial conditions. This process facilitated a more intensive snow accumulation during the following winters which in turn increased denudation (an example of a positive feedback). This spatially differentiated degradation proceeded against the background of the deposition of eolian dust, the parent material of the loess mantle.

Arid type of morphogenesis, in its "purest" manifestation, was mainly restricted to the same regions as today (such as tropical deserts). Many arid landforms dating back to the time in concern, however, are described from periglacial regions and intermediate areas which produced wide varieties of landforms created under the joint influence of eolian processes, sheet and rill wash, piping and of other kinds of erosion and to a lesser extent by mass movements on slopes. In the plains such areas are assigned to **semiarid cold temperate types of morphogenesis.**

Morphogenesis of predominantly **humid types** (mild temperate, that of seasonally wet tropics and equatorial) and characterized by intensive weathering, fluvial processes and creep was very limited at that time.

The vertical zonation in the mountains presents certain difficulties for identification of the types of morphogenesis. In general, it is possible to distinguish "periglacial mountains" with pronounced cryogenic processes and with small local glaciers, from "temperate mountains" covered with forest, where fluvial and other noncryogenic slope processes predominated. In addition, extended areas were occupied by mountains with mixed semiarid and semihumid morphogenesis, where the local pattern of processes was controlled by topography. Arid mountains were less numerous, although they provide a rather spectacular complex of processes (desert pavement formation, thermal creep of debris, desiccation cracks, etc.).

The reconstruction of relief-forming processes during the maxumum cooling of the last glaciation is of special interest because it provides information on the state of environmental components at one of the "extremes" of the natural-climatic cycle. If similar reconstructions were available for the warmest phase of the cycle, they together would represent the whole range of fluctuations of the processes that may be of importance for the assessment of their future development.

VEGETATION DURING THE MAXIMUM COOLING OF THE LAST GLACIATION

by B. Frenzel, in charge of the Research Project Group "Terrestrial Paleoclimatology", Federal Republic of Germany
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(Explanatory notes to map on page 55)

General remarks. A wealth of information is available as to vegetation types in the Northern Hemisphere at about 18,000 years B.P. These indicators are pollen and macrofossil floras, mollusc, insect and vertebrate faunal remains, and distribution patterns of loesses, loess-like, and other eolian sediments. There are frequent problems with age determination of these lines of evidence. Thus, a time interval from about 21,000 to 18,000 years B.P. is used for the construction of the map discussed. Where ^{14}C datings were not available, age determination was carried out using geological, geomorphological, and stratigraphical methods.

Vegetation types. As can be seen from the map on page 55, forests were nearly completely lacking at about 18,000 years B.P. This even holds true for the present-day belt of tropical rain forests, which had retreated to the lower parts of tropical mountain systems with a fairly abundant moisture supply and to more sheltered positions (FLENLEY, 1979). The biological consequences of this disjunction and isolation of previous rain forests have not been thoroughly investigated to date. Rain forest was as a rule replaced by **tropical grasslands** and by **xerophytic shrub.** Simultaneously, various types of vegetation similar to the modern **subalpine forests** had developed in the now dry regions of southwestern North America (e.g., LANNER and DEVENDER, 1981; DEVENDER and BURGESS, 1985).

Nearly all the rest of the Americas and Eurasia, along with North Africa, were covered by various types of **steppe-like vegetation**, which in the arid regions was replaced by **desert shrub** and **semideserts**, in which facultative halophytes were a distinctive component. Tundra was nearly completely lacking, since the moisture available was not sufficient for its maintenance (FRENZEL, 1990).

Each of the widespread vegetation types, indicated in the map, was divided into various subtypes based on differences in available moisture and on parent rocks or soils (GERASIMOV and VELICHKO, 1982; FRENZEL, 1968, 1983). In general, the vegetation cover was an extremely sparse one, and the frequent storms caused deposition of thick loess layers, thus inhibiting the formation of clearly developed soil profiles. Full-glacial soil types belong to the group of serozems, skeletal soils with no identifiable horizons, which indicate extremely weak biological activity. As a consequence, the turnover of organic material was negligible. This would mean that the CO_2 input into the atmosphere from the continents was very limited, thus explaining the strongly reduced CO_2 content in air bubbles of the inland ice in Antarctica and Greenland dating from about 18,000 years B.P.

The problem of net primary bioproduction. Net primary bioproduction was extremely low, and due to presumably intense interannual fluctuations of the available moisture, the ecological conditions for browsing animals are suggested to have been unfavourable. This contributed to a southward migration of herbivores, at least in Northern Eurasia, followed by a similar migration of carnivores (FRENZEL, 1989). It seems that these rigorous climatic conditions initiated the extinction of certain species in northern Eurasia during the Late Pleistocene (*Ursus spelaeus, Hyaena spelaea, Ovibos moschatus, Leo spelaea,* etc.); *Table 8* informs about the net primary production of various vegetation types today and that estimated for full-glacial plant communities. The data for the latter were obtained from present vegetation types approximately equivalent to the past ones (FRENZEL, 1985).

Table 8. Annual terrestrial net primary bioproduction (tonnes per hectare) in the distribution areas of recent vegetation types

Present-day vegetation	t/ha	Full-glacial vegetation	t/ha
Tundra	2.0	Desert with extremely cold climates	0.75
Boreal forest	10.0	Semiarid steppe to semi-desert	6.0
Hygrophilous broad-leaved forest	16.0	Semiarid steppe	6.0
Grassland of temperate climates	10.0	Semidesert	1.0
Savanna	10.0	Semidesert	1.0
Tropical rain forest	30.0	Open grass-savanna	10.0

From *Table 8* and map on page 55 it may be concluded that the **terrestrial net bioproduction** was reduced by about 60 per cent in comparison with present-day natural conditions. Due to the extremely dry soils it is suggested that at that time the terrestrial CO_2 production may have been reduced by about 50 per cent, in comparison to present-day conditions. It is suggested that the mass of fodder available for browsing animals had shrunk on average by at least 30 to 40 per cent. Moreover, the presumably considerable interannual change in moisture supply must be duly taken into account (FRENZEL, 1985) since this meant an additional decrease in the general mean net bioproduction.

MAIN TYPES OF VEGETATION (ECOSYSTEMS) DURING THE MAXIMUM COOLING OF THE LAST GLACIATION

by V. P. Grichuk

(Explanatory notes to map on page 57)

The vegetation of extratropical Eurasia and of Northern Africa was reconstructed for the maximum phase of the Late Pleistocene glaciation — the time interval between 20,000 and 18,000 years B.P. This map represents the first attempt to present a large scale spatial reconstruction of vegetation during this time period. Previously published reconstructions were all fragmentary.

We assumed that the vegetation during the period of minimum heat supply would be most characteristic for the glaciation. This corresponds to the **transition from a cryohygrotic** (cold-humid) **to a cryoxerotic** (cold-dry) **phase** (GRICHUK, 1960). Paleobotanic, geologic, and radiocarbon data from the East European Plain show that this period also corresponds to the maximum advance of the Scandinavian ice sheet (about 20,000 years B.P.) and to the maximum ice advance on the Central Siberian Plateau (KIND, 1974).

Geologic and geomorphologic criteria are most important in selecting sites for paleobotanic studies. Strata deposited during the coldest stage, however, can be identified only paleobotanically. In practice this is possible in thick continuous sections only where pollen analyses reveal vegetational changes. Unfortunately, there are very few radiocarbon dates for paleobotanic sites of this interval. Because of this, we had to use sections with older or younger dates where geological materials indicated no unconformities and, thus, permit extrapolation to be made.

The vegetation map is based on a rather large number of pollen records and includes 26 sites from Western Europe, 106 from the USSR, 54 from Eastern and Southern Asia, 31 from North and South America and 22 from Africa *(Fig. 14)*. In reconstructing full-glacial vegetation, historical floristic data were particularly useful. They are rather numerous since the relict species from glacial floras are common in the present-day vegetation and have been studied by many botanists.

The basic map units employed in this reconstruction are vegetation types similar to those shown on maps of modern vegetation (e.g., „Fiziko-geograficheskij atlas mira", 1964). The following nine vegetation types were identified.

1. **Periglacial tundra vegetation** was spread over most of Western Europe north of the Alps and the Carpathians, in narrow bands along the ice sheet margin on the East European Plain, and in the Ural Mountains. Further east, it was present in northern Siberia and on Kamchatka. The same vegetation was characteristic of Beringia and Alaska. The main formations in western Siberia included subarctic meadows, low bush tundra and open birch woodlands on the plains, and montane bush tundra in the highlands. In Eastern Europe, the tundra and cryophytic steppe communities predominated in combination with open woodlands of larch and birch. In Siberia, polar deserts and montane tundra with some xerophytic steppe associations were widespread. The same vegetation was reconstructed for unglaciated areas of northern Greenland, Iceland, and Scandinavia (based on radiocarbon dated pollen diagrams).

2. **Periglacial steppe vegetation** was reconstructed for Western Europe within intermontane basins, over most of Eastern Europe, for the Western Siberian Lowland, and eastward in central Yakutia. Meadow steppes with larch, pine and birch forests predominated here.

3. **Boreal forests** were present in limited areas of southern and central Western Europe, in the Urals, and in some of the mountains in central and eastern Siberia. They also occurred in Manchuria, on Sakhalin, and in northern Japan, which at that time was connected with the continent. The main formations consisted of mixed (birch, pine, larch) and coniferous (spruce and fir) forests, with an admixture of oak and other broad-leaved species in the southern regions, and with *Tsuga* and *Cryptomeria* in Manchuria and Japan.

4. **Broad-leaved and mixed coniferous forests** (broad-leaved forests of nemoral type) consisting of *Quercus, Carpinus, Fagus*, etc., covered western and southern parts of the Iberian Peninsula, the Apennines, and most of the Balkan Peninsula. Both pollen and historical floristic data indicate their presence in most parts of Asia Minor, the Caucasus, the Zagros Mountains, and along the southern coast of the Caspian Sea. Paleobotanic and other data indicate that the same vegetation existed in the mountains of Soviet Central Asia, in the southern Urals, and in the Altai. Based on paleobotanic and historical floristic information, this type of vegetation is reconstructed also for central and eastern China, Korea, Manchuria, and central Japan. It consisted of coniferous/broad-leaved and broad-leaved forests with evergreen elements.

5. **Steppe and desert vegetation** spread over interior regions of the Middle East and of Central Asia. Desert steppe formations which prevailed here included halophytic communities. In northern Mongolia grass-herb and sagebrush formations dominated, while forests of larch, birch and pine grew on the mountain slopes.

6. **Subtropical tree and shrub communities** were not recorded anywhere in Europe. Isolated patches of such vegetation existed in northern Africa (formations with *Pinus halepensis, Cedrus atlantica, Quercus ilex,* and other evergreen species), in Asia Minor, and in the Middle East (pistachio woodlands with *Periploca* and other shrubs). Pamir-Himalayan formations consisted of juniper communities. Mixed coniferous/broad-leaved forests (with *Abies, Tsuga, Keteleria*, and other conifers) as well as montane evergreen forests with *Magnoliaceae, Lauraceae*, etc. were reconstructed for Eastern Asia.

7. **Tropical forests** were reconstructed from a combination of scanty paleobotanic data and historical floristic information. Tropical savannas (with *Acacia*) and open shrub communities presumably existed in northern Hindustan. Further east they were replaced by humid evergreen forests rich in species.

8. **Tropical desert steppe vegetation** was reconstructed only for northern Africa and for the Arabian Peninsula. It consisted

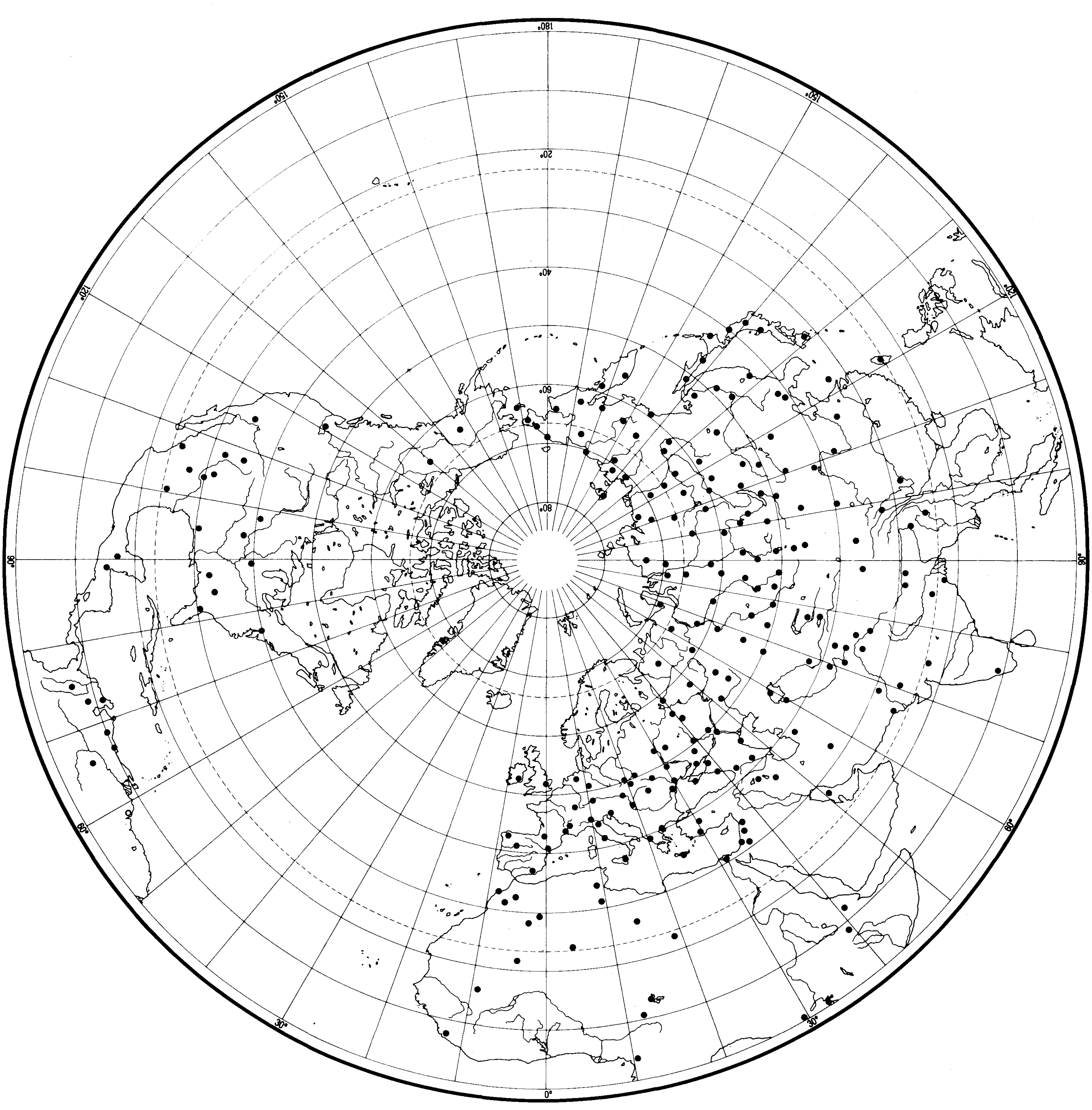

Fig. 14. Sites used for the construction of map on page 57

mostly of semiarid steppe like vegetation (including psammophytic formations with *Ephedra alata*) and also of sandy deserts with bushes (*Calligonum*).

9. **Alpine vegetation** of temperate and subtropical zones (alpine meadows and subalpine tree and bush communities) was confined to mountains south of 50° N. It was reconstructed from historical floristic data only.

10. **Vegetation of the exposed shelves** is shown on the map for vast areas which emerged after a considerable glacio-eustatic drop in ocean level had happened. Pollen data, although sparse, do provide evidence for a meadow-type vegetation with halophytic communities.

LANDSCAPE TYPES DURING THE LAST GLACIAL MAXIMUM

by A. A. Velichko and L. L. Isaeva

(Explanatory notes to map on page 59)

General remarks. The environments during the coldest stage of the Late Pleistocene are of particular interest because their comparison to the present-day ones permits us to estimate the range of changes in the natural systems within the framework of Pleistocene macrocycles (glaciation — interglaciation).

To compile such a map on a subglobal scale, it was necessary to gather data on all the natural components such as vegetation, fauna, glaciers, permafrost, relief-forming processes, oceanic biota and sea-ice, paleoclimates, etc. Materials included into the atlas consist both of these and of some other new data. Some difficulties are caused by a very uneven distribution pattern of reliable data in the Northern Hemisphere. Europe and North America are the best known, while studies of Asia are somewhat less in number (CLIMAP, 1976; WRIGHT, 1983). Rather detailed information, both cartographic and from the literature, was used for mapping these regions. Territories of Central America, Africa, and Southern Asia are less known, and, in some cases, general small scale schemes developed by BOWEN (1979) and VELICHKO (1980) and were used to fill in the gaps.

Oceanic environmental features have been recently reconstructed from microfossil data obtained from Atlantic and Pacific deep sea cores (BARASH, 1980). These data made it possible to determine the position of climatic zone boundaries in the oceans.

For some regions where data are either lacking or ambiguous, an interpolation was used which considered the evidence from adjacent territories for the general pattern of climatic change.

Since an independent paleoclimatic reconstruction was to be done for the entire hemisphere, new criteria were needed to subdivide the environments. No criteria used in small scale maps of modern landscapes were applicable in this respect. It did not make sense to use such terms like "polar regions", "temperate regions", etc., since environments similar to polar ones (though, by no means, identical to them) existed far beyond polar latitudes during the pleniglacial. Furthermore, at the time in concern, very rigorous environmental conditions dominated in the modern temperate zone. This periglacial tundra steppe in no way resembled the present-day landscapes of temperate latitudes.

Considering all these data, VELICHKO introduced another classification scheme for these landscapes. It is based on the distribution of glaciers, permafrost, and vegetation types both on the plains and in the mountains. Four main landscape-climatic zones were distinguished on land: **I — glacial zone, II — cryoboreal and boreal zone, III — warm temperate zone, IV — warm subtropical and tropical-equatorial zone**. The unglaciated oceans were subdivided into climatic zones according to the existing classification: polar and subpolar (I), temperate (II), subtropical (III) and tropical (IV). The resulting map is only a first approach to a synthetic landscape map, and, apparently, requires further work.

I. The **glacial landscape-climatic zone** occupied not only land but considerable areas of present oceans, too. They were characterized by the constant presence of ice on the surface: ice sheets on the plains of North America and Eurasia, various kinds of glaciers in the mountains, and pack ice on sea and ocean surfaces. During the last pleniglacial, about one fifth of all the land belonged to this region.

The Arctic Ocean and part of the Atlantic were covered by continuous pack ice. Opinions about the size of the ice sheets vary considerably. Some authors postulate that a single ice sheet covered northern Europe and northwestern Asia and that its center was located on the Arctic shelf (GROSVALD, 1983). They also believe that a single ice sheet, with its center over Hudson Bay, existed in North America. Recent studies, however, provide evidence for a rather limited distribution of ice sheets in the Northern Hemisphere during the Last Glacial stage. Advances of ice sheets from different centers could have been heterochronous and it makesdifficult to set limits for these glaciers. Zones of such heterochronous advances probably existed between the British and Scandinavian ice sheets, the Scandinavian and Novaya Zemlya ones, and also between the Laurentide and Cordilleran ice sheets. The map also shows the zones of convergence between separate ice domes as well as ice divides within the large ice sheets (VELICHKO *et al.*, 1989).

Glacial regions also included areas of mountain glaciers with reticular glaciers, valley and cirque glaciers, etc. Small ice sheets and ice caps existed in some low mountains. It is worth noting that small „islands" of glacial regions — isolated mountains capped with firn and ice — existed even in the low latitudes.

II. The area of the **cryoboreal and boreal zone** occupied a large part of Eurasia (up to 30° N) and far less extensive regions in North America. In all areas, they were distributed south of the glacial regions. The diagnostic features of these regions are the prevalence of low annual mean temperatures, the high degree of climatic continentality, and a general aridity. These regions encompassed such types of landscape like arctic steppes, steppes, tundra steppes and tundras, periglacial steppes, semi-deserts, and deserts. The forest zone, especially in the Eastern Hemisphere, was considerably reduced. Forests were almost completely absent except for open woodlands along river valleys. Somewhat more extensive woodlands were reported south of the ice margin in the Western Hemisphere (WRIGHT, 1983; VELICHKO, 1984).

Mountains were covered by block fields and shrub tundra in the north, farther south tundra or alpine meadows were associated with steppes and open woodlands on the lower slopes. A few refugia of dark coniferous and broad-leaved forests existed in the Urals, Tien Shan, Altai, and other mountains.

A typical feature of these landscapes was the wide distribution of permafrost with various patterned grounds, ice-wedge polygonal systems, pingos in lowlands, solifluction terraces, etc. The southern boundary of this region approximately coincided with the limit of deep seasonal freezing of the ground. In the oceans, the polar and subpolar climatic zones corresponded to these regions. During the cold seasons, most parts of the oceans were frozen. Based on the distribution pattern of fossil plankton organisms, the southern boundary was drawn along the +8 °C annual mean sea surface isotherm.

III. The **warm temperate zone** was of a limited distribution only, and occurred as separate oases between cryoboreal and warm subtropical regions. They existed on the northern coast of the Gulf of Mexico, on the northern and, occasionally, on the southern Mediterranean coasts, west of the Yellow Sea and the Sea of Japan, and, partially, on Honshu Island. Dark coniferous and broad-leaved forests were characteristic of these landscapes. They resembled the southern forests of the modern temperate zone. In winter months, the ground was mostly frozen in these regions. Since the surface was covered with rather dense vegetation (due to the sufficient supply of moisture) it was protected against wind or sheet erosion, and the fluvial processes (restricted to river valleys) were mainly responsible for relief modelling.

The oceanic temperate zone was also very narrow. Its southern boundary (as estimated from the distribution pattern of the ice-age planktonic microfossils) approximately coincided with the 13 to 15 °C isotherm.

IV. **Warm subtropical and tropical-equatorial zones** were most widespread in Africa. They occupied large areas of southern Asia, and of Central and South America. Up to 80 per cent of these regions were covered by deserts and savannas, semideserts, and, in the mountains, by stony semideserts. Tropical and equatorial deciduous and evergreen forests (considerably more open than at present) grew in southeastern Asia. In Africa and Asia, desert areas were much more extensive than today. Forests were markedly reduced on the plains of Africa and South America (GOUDIE, 1983; STREET-PERROTT and HARRISON, 1985). They shifted to the mountains and were replaced in the plains by savannas. It should be emphasized that within these regions, slope processes were more active in the mountains and in the valleys. The increased continentality of the climate and sparser vegetation cover resulted in the partial erosion of the weathered mantle and in the formation of stone lines in slope deposits. Over the oceans, these regions constituted a subtropical zone whose southern boundary followed approximately the 23 to 25 °C isotherms. Tropical and equatorial conditions were restricted to the central parts of the oceans.

Conclusions. Reconstructions of landscapes at the glacial maximum exhibit a general decrease in annual mean temperatures and an increase in aridity over the Northern Hemisphere. These climatic changes resulted in glacier and permafrost formation in high latitudes, in an expansion of deserts in the subtropics, and in a pronounced reduction of forests. Global sea levels were about 100 m below those of today. Some inland seas may have undergone transgression at that time. This possibility, however, remains to be confirmed.

The climatic and landscape reconstructions presented here do not corroborate previously held theories that glaciations in high latitudes corresponded to pluvials in low latitudes. On the contrary, all available data point to an increase in aridity in the middle and low latitudes during the glacial maxima.

EARLY MAN AND MAMMALS DURING THE UPPER PLENIGLACIAL

MAIN MAMMAL ASSEMBLAGES BETWEEN 24,000 AND 12,000 YR B.P.

by G. F. Baryshnikov and A. K. Markova

(Explanatory notes to map on page 61)

The map of the Late Valdai mammalian fauna is largely based on numerous data on fossil mammals, as well as on some present-day zoogeographical materials. The majority of the dated Pleistocene faunal remains come from Paleolithic sites. Some fossil mammals were found in alluvial, lacustrine, and surface deposits, as well as in loess and fossil soils. All the sites dated between 24,000 and 12,000 years B.P. which correspond to the coldest stage of the last ice age were considered. While compiling the map, about 600 locations with mammalian bones were analysed. About 90 per cent of these were Paleolithic sites assigned to different regional Paleolithic cultures. About one third of the sites, most of them from Europe and North America (the USA and Canada), have been radiocarbon dated. Since their geology and archaeology are the best known, these regions provided the most detailed evidence on the distribution pattern of mammalian communities during the Valdai. Mammalian sites are less numerous in Siberia, and information is relatively scarce from North Africa, the Middle East, India, Indochina, China, and Japan. Almost no data are available for Central Asia; South America and the West Indies are studied somewhat better.

The mammal assemblage areas are designated on the map by different colours and numbers. The boundaries were often drawn arbitrarily with present-day zoogeographical provinces taken into account. Broad zones of transitional fauna are shown for some regions.

The most diagnostic element of the Valdai mammal fauna was the so-called **Mammoth faunal assemblage (N°1)**. It was widespread over vast areas of Eurasia (within the present-day tundra, forest tundra, and taiga zones) and extended into North America. It developed under specific severe climatic conditions, widely spread during the Late Pleistocene Glaciation when periglacial tundra steppes and steppes occupied large areas of northern Eurasia and forests were considerably reduced in area. Highly productive open landscapes permitted the taxa in the mammoth assemblage (such as *Mammuthus, Coelodonta, Saiga, Ovibos, Bison priscus, Equus, Panthera spelaea, Dicrostonyx, Lemmus, Marmota*) to expand their range considerably. Woodland species mostly inhabited high mountains and foothills, but were found in riparian areas as well (GROMOV, 1948; VANGENGEIM, 1977; VELICHKO, 1961; VERESHCHAGIN and BARYSHNIKOV, 1980; MARKOVA, 1982, 1984; SHER, 1971; KURTÉN, 1969; KOWALSKI, 1961).

In Eurasia the southern boundary of this assemblage reached 45° N and coincided, approximately, with the southern limit of mammoth. 50,000 to 40,000 years ago mammoths expanded its range southward into Crimea and into the northern Caucasus. Later, however, they retreated to the north. A similar distributive rearrangement has also been recorded in Eastern Asia. 30,000 years ago mammoths lived in the south up to the Hwang Ho, penetrating even farther south; they also inhabited Hokkaido Island. No mammoth remains younger than 18,000 years B.P., however, have been reported from China to Japan.

The mammoth faunal assemblage was not homogeneous. Two varieties can be distinguished within Eurasia: the **arctic (1a)** (with *Ovibos, Lemmus, Dicrostonyx*), and the **boreal** (*Allactaga, Eolagurus, Lagurus*) **(1b) subassemblages**.The boreal subassemblage was characterized by *Equus latipes*, the presence of *Alticola, Cuon, Uncia, Moschus* and *Capra* in South Siberian mountains and the distribution of *Microtus maximowiczii, Microtus fortis, Nyctereutes, Charronia, Cervus nippon* and *Naemorhedus* in Manchuria and in the south of the Ussurian region.

The area of the latter became smaller in a westward direction and is completely disappeared in Western Europe. Some regional features can also be noted including an increase in woodland species in Europe, the presence of Eastern Asiatic species in Manchuria, and in the Primorje territory (STUART, 1982; BONIFAY, 1969; BOSINSKI, 1969; CHALINE, 1972; DEPLECH, 1983; BUTZER, 1964). Such nearctic species as *Casterodies, Ondatra, Mammuth* and *Camelops* were found in Alaska (HARRINGTON, 1978; KURTÉN and ANDERSON, 1980).

The **Mediterranean mammal assemblage (N°7)** included inhabitants of xeric shrub and montane forests. *Rupicapra* and *Chionomys* are most typical for this assemblage. *Hystrix, Apodemus, Ursus arctos, Meles, Equus hydruntinus, Sus, Cervus elaphus, Dama, Bos primigenius, Capra* and *Ovis* also lived there. Woodland boreal taxa invaded these areas from the north (*Ursus spelaeus, Capreolus, Cervus*). Some Ethiopian species (*Genetta*) reached the Iberian Peninsula from the south. *Herpestes, Lynx pardinus* and endemic taxa (*Capra pyrenaica*) were distributed in the same place.

Marmota, Clethrionomys and *Capra ibex* inhabited the Apennine Peninsula; *Mesocricetus newtoni, Dolomys, Panthera pardus* and *Lynx lynx* — the Balkan Peninsula; *Myotragus* — the Balearic Islands; *Palaeoloxodon antiquus* — the islands of Greek archipelago; *Dryomys, Mesocricetus raddei* and endemic types *Prometheomys* and *Capra caucasica* — the Caucasus. *Dryomys, Hystrix, Nannospalax, Mesocricetus auratus, Microtus guentheri, Canis lupaster, Vormela, Herpestes, Hemibos, Gazella* lived in Palestine in the second half of the last ice age (ARAMBOURG, 1962).

Four faunal assemblages can be distinguished within the broad zone of steppes and deserts of the Old World, although there are only a few concrete data about this area. The **Pontian-Kazakhstan assemblage (N°2)** is typical of steppe species, many of which were also distributed outside its borders (such as *Saiga, Lagurus, Eolagurus, Mustela eversmanni*). Some extinct (*Megaloceros giganteus*) and boreal (*Rangifer tarandus*) taxa were also present. Such steppe species like *Spermaphilus, Marmota, Allactaga, Alactagulus, Spalax, Cricetus, Vulpes corsac* and *Vormela* lived here. *Cervus, Megaloceros, Equus hydruntinus* lived in the Crimea (GROMOV, 1948; MARKOVA, 1982; TATARINOV, 1966).

The **Sahara-Arabian (N°8), Iranian-Turkestanian (N°6),** and **Mongolian (N°3) assemblages** are known only from some sites. They can, however, be reconstructed on the basis of some specific features of modern desert fauna (which is, undoubtedly, very ancient). The Sahara-Arabian mammal assemblage included desert species (*Alcelaphus, Fennecus*). Ethiopian mammals (antelopes *Oryx* and *Addax, Diceros, Sincerus*) were more important here than in the present-day fauna of the same time. The Sahara-Arabian assemblage also included *Hystrix, Hyaena, Herpestes, Caracal,* odd-hoofed mammals (*Equus hemionus, Equus burchelli*), hyraxes (*Procavia*), even-hoofed mammals (*Dama, Bos primigenius, Gazella dorcas, Ammotragus*), and rodents besides *Hystrix (Sciurus, Gerbillurus, Nesokia*).

The Iranian-Turkestanian mammal assemblage included mesic species (*Bos primigenius*), thermophilous forms from Southern Asia (*Hystrix, Panthera tigris*), as well as endemics (*Meriones, Rhombomys, Spermophilopsis*). *Diplomesodon, Ellobius fuscocapillus, Equus hemionus, Sus, Cervus, Gazella subgutturosa, Capra aegagrus* and *Ovis* also inhabited these areas.

The Mongolian assemblage consisted of steppic and desert taxa — *Salpingotus, Gazella gutturosa, Equus przewalskii, Camelus,* and included some now extinct forms (*Spirocerus*). *Myospalax* and *Lasiopodomys brandti* are typical of this assemblage.

The modern zoogeographical map of Africa suggests that three mammal assemblages occupied its tropical areas. The **Sudanese savanna assemblage (N°19)** included *Diceros, Giraffa, Equus burcelli, Loxodonta africana, Orycteropus, Phacochoerus, Alcelaphus, Damalisus, Syncerus* and *Acinonyx*. **Congolese equatorial woodland assemblage (N°20)** was reconstructed by taking into account the antiquity and African forests fauna abundance of species. *Pan, Gorilla, Tragelaphus, Boocercus, Choeropsis* and *Okapia* were mostly typical of this assemblage.

The mountain **Abyssinian mammal assemblage (N°18)** was distinguished on the basis of modern fauna uniqueness. It included such montane species as *Capra nubiana, Equus (Hippotigris) zebra* and *Simia*. The primates were represented by *Papio hamadryas* and *Theropitecus*.

The distribution of the assemblages within the African tropical zone (N°18, N°19, N°20) is shown on our map provisionally.

Very few reliable data are available on the Valdai fauna of Central Asian highlands. Two assemblages can be distinguished here: the Tien Shan-Hindukush and Tibetian ones. Their species composition and distribution were defined on the basis of modern zoogeographical features. The **Tien Shan-Hindukush assemblage (N°4)** included both montane-woodland Alpine species (*Uncia, Cuon, Ovis ammon*) and those of rocky habitats (*Capra sibirica, Capra falconeri, Alticola* and *Neodon*). Marmots, brown bears, cervids, and roe deer also inhabited the regions while *Selenarctos* was found further south.

The mammals of the **Tibetian mountain-desert assemblage (N°5)** were adapted to severe climatic conditions like those of the tundra steppe. This accounts for the wide distribution of some ungulates in the Tibetian mountain-desert assemblage during the cold stages of the Pleistocene. *Poephagus,* for example, lived as far north as Yakutia, and was a member of the mammoth assemblage. *Ochotona, Vulpes ferrilata, Ursus pruinosus, Equus kiang, Pantholops, Ovis ammon* and *Pseudois* also were present in this assemblage.

The **Japanese-Chinese mammal assemblage (N°9)** included open landscape species (*Coelodonta, Sinomegaceros, Bison priscus, Bos primigenius*), boreal forest taxa (*Palaeoloxodon*), and Chinese forest mammals (*Lasiopodomys, Charronia, Cuon, Panthera tigris, Moschus, Cervus nippon, Nyctereutes*). The northern boundary of this assemblage was not stationary and, during the Late Valdai, migrated northward (KAMEI, 1981).

The **Indo-Chinese mammal assemblage (N°10)** of subtropical and mountain tropical forests existed to the south of the Yangtze river. It included some species of the ancient Indo-Malayan fauna which exist in this area today: *Ailuropoda, Selenarctos thibetanus, Neofelis nebulosa, Tapirus, Muntjacus* and *Rusa*. Primates were represented by *Pygathrix* and *Pongo*, rodents by *Hystrix* and *Hapalomys*. Some inhabitants of these areas (*Stegodon* and *Pongo*) are now extinct (HAN and HU, 1985).

The **South Himalayan forest steppe subtropical assemblage (N°11)** (*Ailurus-Budorcas*) occupied the southern slopes and foothills of the Himalayas. This complex included South Asian forest forms (*Macaca, Axis, Rusa, Cuon, Panthera tigris* and *Rhinoceros*), mountain taxa (*Eupetàurus, Budorcas*) and Central Asian steppe forms (*Equus, Bos*). It also included very few extinct species (*Stegodon*).

The composition of the **Indian savanna assemblage (N°12)** underwent a slight change through time. It consisted of some Indian endemics (*Melursus, Boselaphus, Antilopa cervicapra*), Southern Asiatic species (*Hystrix, Elephas, Bos namadicus, Bubalus* and *Rusa)* together with some extinct species of ungulates and proboscideans (*Stegodon, Hippopotamus*) (DASSARMA and BISWAS, 1977).

Forest faunas, which were rich in species, though they are not well known, characterized the tropical areas of the southeasternmost Asia and the islands of the Malayan archipelago. Their location and supposed species composition were defined on the basis of modern faunal data. The natural habitat of the **Malayan equatorial woodland mammal assemblage (N°21)** comprised the Malacca Peninsula and the Big Zond Islands which became a part of the Asian continent due to the lowering of the World Ocean's level during the second half of the Late Pleistocene when the shelf zone became dry. The forest mammals such as *Cynocephalus, Tarsius, Pongo, Cuon, Panthera tigris, Elephas, Rhinoceros, Viverra, Paradoxurus, Tragulus, Cervus* and *Rattus* exist in this area today. Stegodons and hyenas which lived on the island Java become extinct now. The Malayan fauna, poor in species, which occupied the southern Philippine Islands, can be considered as a particuiar **Mindanao insular mammal assemblage (N°22)**. *Podogymnura* and *Urogale* were endemic for this assemblage. Flying lemurs, tarsiers and palm civets also inhabited here. The fauna of the Luzon Island had got an even more clearly expressed insular character with high endemic features. The **Luzon insular mammal assemblage (N°23)** is characterized by the absence of many widely distributed species of primates, carnivores and hoofed animals. Some endemic gnawing mammals such as *Rhynchomys* and *Capromys* were found in it. Both mammal assemblages just mentioned are distinguished by the modern fauna originality and their insular isolation.

The very original **Sulawesian transitional assemblage (N°24)** included not only endemic placental mammals of Asian origin (*Babyrousa, Anoa*) but also Australian marsupials (*Phalanger*).

In North America, the mammoth assemblage area included Alaska. The wooly rhinoceros, however, did not live there. Almost all of the Canadian territory was covered by the ice sheet and the Late Pleistocene mammals have only been reported from the southernmost regions. Five mammal assemblages were identified from east to west in the United States. At that time their distribution pattern was, just as it is today, longitudinal (LUNDELIUS *et al.*, 1983; KURTÉN and ANDERSON, 1980).

The **Appalachian forest assemblage (N°13)** consisted of woodland and boreal species (*Mammut, Mammuthus jeffersoni, Tayassu, Rangifer, Sciurus, Casteroides, Synaptomys, Phenacomys, Bison* and the extinct *Platygonus)*. In the north, near the ice margin, they were augmented by collared lemmings (*Dicrostonyx*) and wooly mammoths (*Mammuthus primigenius*). Further south this assemblage was replaced by the more thermophilous **Floridan subtropical forest assemblage (N°14)** which is characterized by *Tapirus, Neofiber, Sciurus* as well as *Glyptoterium, Tamias, Ochrotomys, Neochoerus, Hydrochoerus, Mammut, Arctodus pristimus, Tremarctos* and *Mammuthus floridanus* (GUILDAY, 1971).

Steppe and semidesert species inhabited the arid regions of the middle and western parts of North America. Three mammal assemblages can be distinguished there: the **Central North American steppe assemblage (N°15)** of the Great Plains (including *Antilocapra, Geomys, Glossotherium, Platygonus, Spermophilus, Arctodus, Mammuthus jeffersoni, Equus, Camelops, Bison, Symbos, Rangifer* and *Ovibos* in the North, *Dasypys* and *Tapirus* in the South), the montane one of the **Rocky Mountains (N°16)** (with *Oreamnos, Navachoceros, Nothrotheriops, Ochotona, Marmota, Microtus montanus, Camelops, Ovis canadensis, Eucerathrium, Lagurus* and *Rangifer* in the north, *Capromeryx* in the south), and the **Californian xeric mountain woodlands** one **(N°17).** The latter occupied California, the Californian Peninsula and some adjoining islands. It consisted of *Aplodonta, Sciurus griseus, Oreamnos, Mammut* and, in the south, also of *Tapirus, Bassariscus, Camelops. Euceratherium, Nothrotherium* and *Peromyscus* were present in this assemblage, too.The pigmy mammoth form was discovered in the Channel Islands.

The boundaries between the Rocky Mountains and the Californian assemblages were not distinct. Some periglacial taxa (*Rangifer tarandus, Ovibos* and *Lagurus lagurus*) existed in the northern parts of these assemblages.

The **Central American transitional assemblage (N°25)** lived further to the south than the xeric ones described above. Its natural habitat occupied the midcontinental isthmus which was the zoogeographical filter in the Great American faunal exchange during the Pleistocene. Some nearctic species (*Sorex, Bassariscus, Orthogeomys, Canis, Urocyon, Arctodus, Smilodon, Tapirus* and *Odocoileus*) penetrated here from the north, as did some representatives of the neotropical fauna (*Didelphis, Glyptodon, Megatherium, Nothrotheriops, Dasyprocta* and *Ceboidea*) from the south.

The fauna of the Greater Antilles had an original character. The **Antillian insular assemblage (N°26**) included few mammals but the large flightless birds partly substituted for them biocoenologically. This assemblage was characterized by its insular isolation and extended to Cuba, Haiti, Puerto Rico and Jamaica. *Solenodon* and *Nesophontes* amongst the insectivores, *Ateles* among the primates, *Geocapromys, Plagiodonta* and *Isolobodon* from the rodents and *Acratocnus* from the edentates were present here. A poor fauna was found in Jamaica (endemic, *Clidomys*) and in the Bahama Islands (*Geocapromys*).

South America did not have the family representatives of *Elephantidae* and *Mammutidae.* Its Pleistocene fauna is not well studied yet, and its zoogeographical characterization is thus preliminary in nature only. The **Orinocian savanna assemblage (N°27)** (the llanos of Venezuela) and the **South American tropical forest assemblage (N°28)** (the Pacific Ocean coast and the Amazon Basin) occupied the flat areas of the northern part of the South American continent. The savanna mammal group included *Didelphis* from the marsupials, *Propraopus, Pampatherium, Glyptodon, Megatherium, Eremotherium* and *Mylodon* from the edentates, *Neochoerus* from the rodents, *Arctodus* and *Conepatus* from the carnivores, *Windhausenia* from the Litopterna, *Toxodon* from the Notoungulata, *Stegomastodon* from the proboscideans, and *Amerhippus* from the perrissodactyles.

The **Andes montane mammal assemblage (N°29)** extended into the Colombian mountain areas. Its basic natural habitat was located in the Andes to the south of the equator. *Tremarctos, Cuvieronius, Dusicyon, Protocyon* and *Nasuella* from the carnivores; *Amerhippus* from the odd-toed ungulates, *Palaeolama* and *Odocoileus* from the artiodactyles lived here. *Haplomastodon* existed in the mountain valleys.

The distribution pattern of the mammal faunal assemblages within the tropical and subtropical Northern Hemisphere during the Late Valdai (Late Würm, Late Wisconsin), in general closely resembled the present-day ones. The mammal associations of the woodland landscapes were especially stable. The high and middle latitudes were occupied by a specific mammoth assemblage which had a holarctic distribution. The ecological structure of this assemblage slightly resembles that of present-day African savannas and has no modern analogues among the assemblages of the present boreal regions. Such paradoxical paleoecological variant suggests the development of the tundra-steppe associations with the structural elements of the arctic cryogenic savanna in the north of Eurasia and in Beringia.

In composing this map we used original data obtained by ourselves as well as information published by E. ANDERSON, J. ALTUNG, C. ARAMBOURG, F. BACHMAYER, S. BISWAS, D. A. M. BATE, M. F. BONIFAY, V. I. BIBIKOVA, H. BREUIL, K. W. BUTZER, B .G. CARETTO, J. B. CAMPBELL, J. CHALINE, C. S. CHURCHER, J. CLUTTON-BROCK, D. C. DASSARMA, F. DEPLECH, E. DUPONT, G. ESTEVEZ, W. W. FERGUSON, V. FERRANT, K. K. FLEROV, M. FRIANT, D. A. E. GARROD, V. E. GARRUT, G. GIACOBINI, R. W. GRAHAM, W. F. GRIMES, A. N. GORING-MORRIS, F. I. GROMOV, U. M. GROMOV, J. GUILDAY, G. GUERIN, C. R. HARRINGTON, HAN DEFEN, J. A. HOLMAN, HUANG WANPO, HU CHUNHUG, A. HEINTZ, D. JÁNOSSY, H.-D. KAHLKE, T. KAMEI, T. KORMOS, K. KOWALSKI, M. KRETZOI, D. KUKAVINA, B. KURTÉN, I. E. KUZMINA, K. LUCHTERHAND, E. L. LUNDELIUS, M. MALEZ, M. MINATO, A. NADACHOWSKY, H. OBERMAIER, I. G. PIDOPLICHKO, V. POPOV, A. M. RADMILLI, G. RADULESCU, C. A. REPENNING, A. RUST, P. SAMSON, K. SKOTT, A. V. SHER, D. W. STEADMAN, G. STORCH, A. L. STUART, E. TCHERNOV, E. TERZEA, W. TOPACHEVSKI, N. K. VERESHCHAGIN, I. VÖRÖS, S. D. WEBB, G. ZBYSZEWSKI and others.

The authors are indebted to professors J. CHALINE, H.-D. KAHLKE and K. KOWALSKI for their comments on the first draft of this map and also to A. V. SHER for his valuable assistance in the selection of the references.

HUMAN OCCUPATION OF THE OLD WORLD DURING THE LAST GLACIATION

by T. Madeyska

(Explanatory notes to map on page 63)

In the present state of prehistoric knowledge the reconstructed picture of human settlement during the upper part of the Last Glaciation in the Northern Hemisphere (Upper Würm in the Alps, Upper Pleniglacial of the Vistulian in Western and Central Europe, Upper Valdai in the Russian Plain, Sartan in Siberia) cannot be objective or complete. The main reasons for this situation are: an equivocal state of research in different regions and the complexity of the problem of chronological and stratigraphic correlation.

Regarding the correlation of sites from different paleogeographic regions, radiocarbon dates must serve as the main criteria, in spite of their well known inherent problems. In addition to this, local geological stratigraphy, mainly climatostratigraphy, was very useful as a control or as a base for correlations on a regional scale. For a correlation on an interregional scale this criterion is less reliable because of the asynchroneity of the maximum extent of ice sheets as well as that of the fluctuations of humidity in different regions of the world.

In these brief notes it is impossible to cite each source used. The most important references of information used for the selection of European sites include papers by: I. BARANDIARAN, J. BÁRTA, M. BITIRI, C. CACHO, M. CARCIUMARU, J. CEDRA, V. CHIRICA, H. DELPORTE, R. DESBROUSSE, V. DOBOSI, M. ESCALON de FONTON, P. GAMBASSINI, P. HAESAERTS, J. HAHN, J. K. KOZLOWSKI, H. LAVILLE, M. OLIVA, M. OTTE, A. PALMA di CESNOLA, F. B. de QUIROS, J. P. RIGAUD, J. ZILHAO, published among others in: IX Congress UISPP Coll. II, XV, 1976; L'Aurignacien et le Gravettien, 1980; Aurignacien et Gravettien en Europe, 1982; The Pleistocene Perspectives, 1986; a synthesis by J.K. KOZLOWSKI and S.K. KOZLOWSKI, 1979; La Préhistoire Française, 1976; BORISKOVSKIJ, 1984; numerous reports in "Radiocarbon" and publications by J. ALTANA, G. BARTOLOMEI, I. K. IVANOVA, N. D. PRASLOV, F. PÉREZ, J. RENAULT-MISKOVSKY, A. LEROI-GOURHAN, J. SIMMONS, M. TOOLEY and others.

Information about North African and Near Eastern sites was obtained from the works of H. ALIMEN, O. BAR-YOSEF, P. BIBERSON, K. W. BUTZER, G. CAMPS, J. D. CLARK, G. DELIBRIAS, W. R. FARRAND, K. FLANNERY, D. GARROD, F. A. HASSAN, J. HEINZELIN, R.W. HEY, F. HOLE, A. E. MARKS, C. B. M. McBURNEY, J. ROCHE, A. RONEN, R. SAÏD, R. SCHILD, R. SOLECKI, F. WENDORF, published in different journals and in books: FAURE and WILLIAMS, 1980; MAHANEY, 1981; IX Congress UISPP Coll. II, III, 1976; Paleolit Blizhnego i Srednego Vostoka, 1978.

Data on Asia came from publications by: D. P. AGRAWAL, I. S. AIGNER, S. N. ASTAKHOV, N. N. DROZDOV, M. D. DZHURAKULOV, GAI PEI, F. IKAWA-SMITH, V. JAYASWAL, J. P. KNOLUSHKIN, V. E. LARICHEV, S. A. LAUKHIN, S. A. NESMEYANOV, A. ONO, J. N. PAL, V. A. RANOV, collected in books: IX Congress UISPP Coll. VII; Paleolit Srednej i Vostochnoj Azii, 1980; BORISKOVSKIJ, 1984; BRYAN, 1978; XI INQUA Congress Abstracts, 1982; The Pleistocene Perspectives, 1986; monograph by TSEJTLIN, 1979; Recent progress of Quaternary research in Japan, 1981; Quaternary geology and environment of China, 1982; as well as from publications by: Z. A. ABRAMOVA, T. AKAZAWA, A. P. DEREVJANKO, M. M. GERASIMOV, Yu. A. MOCHANOV, A. P. OKLADNIKOV, C. V. RUDENKO, T. E. G. REINOLDS, M. SHACKLEY, published in various journals.

The **Upper Paleolithic settlement** of the European northern latitudes was limited by low temperatures and by the proximity of the ice sheets. These factors resulted in the shifting of the northern limit of the inhabited land southwards during the cold substages. The minimum distance of the sites from the ice sheet margin was about 400 to 600 km.

In lower latitudes, in North Africa, in the Near East, and in Central Asia, aridity was the main climatic factor limiting human settlement. Fluctuations in humidity resulted from the modification of the jet streams and from many local changes in the movement of the air masses. These fluctuations overlapped global temperature changes associated with solar radiation. The net result was a considerable areal and temporal differentiation in humidity.

The youngest **Aterian assemblages** in the Sahara are dated to about 40,000 years B.P. Subsequently and during the following hyperarid period between 20,000 and 12,000 years B.P., the Sahara was unoccupied. The aridity of this period is documented at several Aterian localities in the Sahara and in southern Egypt and was expressed in the deflation of archaeological horizons and in the polishing of artifacts by wind.

During this arid period human settlement concentrated along the Atlantic and Mediterranean coasts, as well as in the Nile valley. The **Iberomaurisian assemblages** along the Mediterranean coast of Morocco and Algeria date from between 23,000 and 13,000 years B.P., while they are slightly younger along the Atlantic coast.

In the Nile valley of Lower Nubia, Upper Paleolithic sites are associated with the youngest part of the **Masmas formation,** dated to approximately 25,000 to 18,000 years B.P. This formation represents overbank deposits which record a period of vigorous floods, more extensive than recent ones, which occurred in Egypt during a period of arid climate. The floods resulted from an increase of summer precipitation along the eastern coasts of Central Africa. In the Nile valley, between Luxor and Idfu, the **Masmas-Ballana formation** consists of flood silts with eolian sands and even containing dune sand in its top part. This confirms the correlation of this formation with a period of hyperarid conditions in the desert. Two different **archaeological complexes** occur in this restricted geographical area: the **Idfuan** and the **Fakhurian**. In Sudanese Nubia the corresponding **Dibeira Jer formation** contains the **Khormusan** and the **Halfan.** These industries date back between 25,000 and 15,000 years B.P. Although some of the dates are problematic, it is clear that human settlement in the Nile valley continued during the whole period corresponding to the upper part of the Last Glaciation in Europe. The coexistence of several different industries here indicates that the Nile valley acted as a refuge area for groups withdrawing from the desert.

In the Near East, the Upper Paleolithic evolved from the local **Moustérian** through several transitional industries. The general term "Levantine Aurignacian" was recently replaced by local cultural designations. **Aurignacian-type cultures** lasted from 40,000 to 20,000 years B.P., and were later followed by the **Kebara** and, from about 16,000 years B.P., by the **Geometric Kebaran.** An almost continuous occupation is documented in the caves of the Zagros Mountains, where the transition from the **Baradostian** to the **Zarzian culture** dates from about 20,000 years B.P. The environment during the entire period was quite favourable for human occupation in the entire Near East macroregion, with the exception of the desert. Environmental differences in particular regions did bring about local translocations of human groups, e.g., from mountains to lowlands or from lake basins to caves.

The arid lowlands of Central Asia between the Caspian and Aral seas and the Balkhash Lake were probably devoid of human settlement during the last glacial time. Sites date from a period before or after the glaciation there.

Human occupation of the mountains of Central Asia seems to have been sparse but continuous from the Middle Pleistocene to the Holocene, persisting all through the Late Pleistocene climatic fluctuations.

Human settlement in Japan was more extensive. The chronology of the **Japanese Paleolithic** is based mainly on fission track and radiocarbon datings. The ecological and climatostratigraphic data are sparse because of the rarity of organic matter at the sites. The Upper Paleolithic of Honshu dates from between 30,000 and 12,000 years B.P. The oldest settlement on Hokkaido dates to 17,000 years B.P. The continuous but differentiated Paleolithic history of Japan was very much influenced by sea level fluctuations. The occupation of Japan from continental Asia could have occurred during the time when the Japanese islands were a part of the mainland. During the subsequent periods of lower sea level migration of early man in the opposite direction was possible. This is suggested by cultural connections observed between Japan, Korea and the Far East.

While evidence on the Upper Paleolithic settlement of Korea, China, and the Far East is not abundant, Siberian remains are more numerous. Recent research in Siberia indicates a regional diversity in the Upper Paleolithic. Increasing aridity in Central Asia, which occurred simultaneously with the Sartan Glaciation, probably resulted in the shifting of human settlement northwards along the main Siberian river valleys. The continental climate of Siberia during the time in concern brought about the development of a vast tundra-steppe zone with a rich biomass. This permitted Paleolithic groups to occupy regions beyond 60° N. This settlement did shift for a short time southwards during the coldest Gydan Stadial.

HUMAN OCCUPATION OF THE AMERICAS BETWEEN 24,000 AND 15,000 YR B.P.

by O. Soffer

(Explanatory notes to map on page 63)

General remarks. While the occupation of North America by hunter-gatherer groups using bifacially worked projectile points is widely accepted by 12,000 years B.P., data on earlier human presence on the continent has generated a great deal of debate amongst the specialists (BRYAN, 1986; CARLISLE, 1988, DINCAUSE, 1984; IRVING, 1985; OWEN, 1984; SHUTLER, 1983). The range of opinions on when the continent was colonized include (a those who accept human presence before 40,000 years B.P., (b those who see colonization between some 30,000 and 15,000 years B.P., (c those who hold that no unequivocal data exist for human presence before the well documented Clovis period of some 12,000 years ago. DINCAUSE (1984) documents over 25 sites and/or localities in North America for which pre-Clovis occupation has been claimed. None of these have won full acceptance as unequivocal Late Pleistocene sites because numerous problems exist with their dating as well as with the context or nature of the cultural remains found.

The following sites figure most prominently in the literature for the 24,000 to 15,000 years B.P. time period:

1. **Old Crow Basin** in the Yukon Territory, Canada, has yielded a large number of flaked, fractured, and cut pieces made of mammoth bone and ivory which some researchers consider as modified by human hand and as remains of bone tool kits. These pieces were all recovered in redeposited fluviatile contexts. Radiocarbon dates obtained on bone apatite have yielded dates between 37,000 and 23,000 years B.P. Recent redating of the famous flesher produced a young Holocene date of 1,350 years B.P. Current research at Old Crow Flats is concentrating on examining the taphonomy of these items to ascertain the extent of natural vs. cultural modification present.

2. **Bluefish Caves** (I, II, III), have been recently discovered and investigated preliminarily in the Keele Range, Yukon Territory, Canada. They contain loess deposits with faunal remains, retouched lithic pieces made of non-local materials, as well as some polished bone pieces which may have resulted from human use and modification. Twelve radiocarbon dates, eleven of which came from bone collagen and one from charcoal range from 23,500 to ca. 10,000 years B.P. The full acceptance of pre-Clovis occupation here awaits further research and publications.

3. **Channel Islands.** Finds made at these islands off the California coast (USA), purportedly include hearths in association with burnt bones of Pleistocene fauna. Remains discovered on Santa Rosa Island include dwarf mammoths. These finds have been dated to a period between 37,000 and 11,000 years B.P. The faunal remains, together with some stone tools, all come from redeposited alluvial contexts. Burned areas previously interpreted as remains of hearths on these islands probably resulted from natural root and stump fires.

4. **Laguna Beach.** Road construction carried on in 1931 uncovered two highly mineralized human cranial bones in this southern California (USA) location. Subsequent research indicated that the bones were washed down from higher lying strata which dated between 8,000 and 9,000 years B.P. The cranial bones, after travelling extensively through Europe and Africa, in the late 1960s were radiocarbon dated on collagen between 14,800 and 17,150 years B.P. Recent dating of amino-acid fractions placed their age at 5,100 years B.P. BADA used these acontextual remains and the ^{14}C date obtained for them for his calibration of the amino-acid racemization dating scale subsequently employed to assign chronometric dates to other hominid finds made both in California and elsewhere in North America. The disconformity between the dates on the cranial remains themselves and their purported context have caused many scholars to reject the Late Pleistocene assignment for these remains.

5. **Wilson Butte Cave** in southern Idaho (USA) contains a series of stratified deposits. The lowest levels, stratum C, contained a few stone artifacts, bones of camel, horse, and sloth, as well as remains of microfauna which ^{14}C dated to 14,500 years B.P. At present the association between the artifacts and the dated microfauna remains unclear and problematic.

6, 7. **Dutton and Selby Sites,** located on the eastern Colorado high plains (USA), contain extensive fossil beds. They were found in lacustrine unconformities below the Late Wisconsin soil horizon and stratigraphically dated to about 15,000 years B.P. The beds contained flaked, cut, and split bones of such Late Pleistocene taxa as mammoth, horse, bison, camel, sloth, and peccary. At the early stages of study, some researchers suggested that the bones found at these two sites represented remains of bone tool inventories. More recent taphonomic research conducted at Dutton and Selby indicates that natural factors were most probably responsible for the damage observed on the bones.

8. **Meadowcroft Rockshelter** is located in southwestern Pennsylvania (USA) in a narrow, steep-sided valley near the city of Pittsburgh. It is a complex stratified site with human occupation evident as late as the eighteenth century. The lowest levels, containing faunal remains, a few small sized pieces of worked stone, and some hearths have been assigned to human occupation at about 16,000 years B.P. and possibly as early as 19,000 years B.P. The Pleistocene dates obtained for the layers in the lower stratum at Meadowcroft range between about 19,000 and 11,300 years B.P. and stand in contrast with the strictly Holocene nature of the floral and faunal remains found in the same levels at the site.

9. **Tlapacoya,** located on the ancient shores of lake Chalco, at the eastern outskirts of Mexico City, Mexico, consists of some 18 localities, One of these, Tlapacoya I, contained remains of three hearths and a few obsidian and andesite tools and flakes along with the bones of bears and deer. A sequence of ^{14}C dates ranging from 24,000 to 15,000 years B.P. have been obtained from this and other localities at Tlapacoya, but their association with the cultural remains is problematic. The resolution of this issue must await future complete publication of work done at Tlapacoya.

10. **Valsequillo** (Hueyatlaco) river gravels, located southeast of the city of Puebla, Mexico, constitute a locality with extensive series of fossiliferous gravels. The lower gravels contain strata which, at various locations, have yielded flaked bones of

such extinct Pleistocene taxa as horse, camel, and elephant, as well as stone artifacts and flakes. Radiocarbon dating done on *in situ* freshwater shells in the lower strata produced two dates of 22,000 and 9,000 years B.P. Other dating methods employed (U/Th series, fission track and tephrachronology) gave age assignments for the lower gravels in excess of 200,000 years. At present, questions exist about the dating of the gravels, correlation of them from one location to the other, about the association of extinct fauna and stone tools, as well as about the artifact status of some of the lithic pieces discovered.

11. **El Bosque** in Nicaragua has yielded remains of bones of such Pleistocene species as sloth, mastodon, and horse, as well as flaked pieces of jasper. Radiocarbon dates obtained from the bones and from carbonate nodules range from 18,000 to over 35,000 years B.P. The flaked pieces of jasper are seen by some researchers as artifacts while other scholars seriously dispute their man-made status and association.

12. **Taima-Taima.** The base of an old spring deposit in northern Venezuela yielded a number of mastodon skeletons including one of a juvenile individual found with a segment of an El Jobo point, a distinctive nearly cylindrical willow leaf point, lodged in the cavity of the mastodon's right innominate. Another flaked stone artifact was found nearby. A series of radiocarbon done on the bones as well as soil, lignite, and wood found near the mastodon gave a wide range of dates with the oldest around 13,000± 2,000 years B.P. Since the sedimentary history and post-depositional processes at Taima-Taima still remain unclear, and because only the midsection of the point was found with the mastodon, the question of the association of the tools with the fossil fauna awaits further investigation.

Conclusions. Using these data, as well as other more tenuous claims not discussed here, for human occupation of North America before 12,000 years B.P., it is difficult to come to a firm conclusion. What is clear today is the pervasive human presence across the continent after 12,000 years B.P. and a possibility of such presence, albeit spatially discontinuous and sparse, during the 24,000 to 15,000 years B.P. discussed in this volume. The best data for such pre-Clovis presence comes from surprisingly few sites: Meadowcroft Rockshelter in North America and Monte Verde, dating to ca. 13,000 years B.P., at the southern end of South America, in Chile. These sites, all pre-dating Clovis age occupation, together with Bluefish Caves and the Nenana complex of sites in Alaska (ADOVASIO, in press), point to the probability that pre-Clovis groups in the New World, if present there on a permanent basis, were both very few in number and sparsely distributed across the continent. This stands in sharp contrast to the density and distribution of Clovis and other Paleoindian sites dating after 13,000 years B.P. — a period when the continent was successfully occupied from the Pacific to the Atlantic coasts.

HOLOCENE CLIMATIC OPTIMUM

CLIMATES AT ABOUT 7,000 TO 6,500 YR B.P.

by B. Frenzel, in charge of the Research Project Group "Terrestrial Paleoclimatology", Federal Republic of Germany
Co-authors: B. Becker, H.-J. Beug, K. Brunnacker, D. Busche, P. Frankenberg, P. Fritz, M. A. Geyh, H. Grotzfeld, H. Hagedorn, K. Hecker, J. Hövermann, A. Kessler, W. v. Königswald, K. Krumsiek, W. Lauer, H. Mensching, H. Moser, K.-O. Münnich, Chr. Sonntag, R. Vinken, W. Westendorf)

(Explanatory notes to maps on pages 67, 71, 75 and 79)

General remarks and methods used. It is the aim of these investigations to analyse those climatic conditions, which prevailed on the Northern Hemisphere shortly before the beginning of a strong neolithic human impact on vegetation. It is well known that the neolithic revolution commenced in various regions of Eurasia at different times, with the earliest impact in the Middle East and along the river Nile. Yet to the north of the mountain belt which includes the Pyrenees, the Alps, the southern Carpathians, the Caucasus, and the Himalayas, this event of utmost paleoecological importance only started at about 6,000 to 6,500 ^{14}C years B.P. As to the calibration of ^{14}C years against the dendrochronological time-scale, see Radiocarbon, **28**, 2B, 1987.

In the above indicated maps the climatic conditions at about 6,500 to 7,000 ^{14}C years B.P. are shown, whereas reconstructions by Soviet experts concentrate on a younger time interval (6,000 to 5,500 ^{14}C years B.P.), when neolithic man had already strongly influenced vegetation over vast territories of Western, Central, and Eastern Europe, in South Asia, in China, and presumably in Tibet, as well. Moreover the interval 7,000 to 6,500 years B.P. refers to the climatic optimum, whereas the interval 6,000 to 5,500 years B.P. falls into the transition to a subsequent period of cooling. As to the methods used for reconstructing climates, see explanation to map on page 15. Again it must be stressed that paleoclimatic data derived only from fossil material were used, not those obtained from climate models. The sites investigated are given in *Figure 15.*

In general, the deviations of climatic parameters for Middle Holocene time from present-day conditions are not significant. Thus, distribution patterns of entire plant communities are not as suitable for reconstruction of Middle Holocene climates as distribution patterns of the most sensitive plant and animal taxa. These sensitive taxa generally occurred only in suboceanic to oceanic climates. Thus the starting point for the reconstructions is not equally reliable for the various regions of the Northern Hemisphere. On the other hand, changes in the distribution pattern of permafrost at about 6,500 years B.P. relative to today can be used to some extent in Central Siberia.

Winter temperatures (map on page 67). As can be seen from this map, nearly everywhere on the continents of the Northern Hemisphere the mean air temperatures were about 1.5 °C higher than today. The sea surface temperatures of the Earth's oceans have not yet been analysed either by CLIMAP or by other research groups.

In the international literature, there are references to distinct local deviations from the general picture. This is primarily caused by different methods used to obtain proxy data of the past. Most of the data evaluated conform to the common pattern, which shows only minor deviations of winter temperatures at about 6,500 B.P. from present-day conditions, so these more uniform and small values were chosen to provide information about the minimum temperature anomalies. Nevertheless it is obvious that the northern part of Eastern Europe had experienced much more intensive warming than other regions of the Eurasian continent. This suggests that the milder winters were caused by a more pronounced influx of moist air masses from the north than today, which in turn should have influenced precipitation in Eastern Europe. This problem will be discussed in the explanations to map on page 79.

The positive deviations of winter temperatures, in comparison to the present in tropical Africa or South America, were very similar to those at the mid-latitudes. Consequently, relatively warm winters within the cool temperate and boreal belts of the Northern Hemisphere, shown in our map, did not result from an uncritical use of climatic indicators, rather they refer to a general trend.

At least in Europe and to some extent in North America, no positive deviations of Middle Holocene winter temperatures were found for the more oceanic regions. The validity of this conclusion, however, is doubtful, since under these presently mild climates a mid-Holocene phase of increased winter temperatures is difficult to verify due to methodological problems.

Summer temperatures (map on page 71). With the exception of subpolar regions, mean air temperatures in July seem to have been higher nearly everywhere than today by about 1.5 °C at most. This holds true for tropical climates as well as for the temperate and boreal ones. Yet the subpolar belt had obviously experienced an even more pronounced warming with 2 to 3 °C positive deviations for the summers.

Thus, the same situation existed during the Holocene climatic optimum as predicted in some models for the near future. It was mentioned that the same had held true for the Last Interglacial (see map on page 19), and perhaps also during the penultimate interglacial. Thus a general tendency in the intensity of relative warming for the continents during interglacial periods seems to be evident, though the extent of this warming varied from interglacial to interglacial.

Approximately the same positive deviations of summer climates relative to present-day conditions (see our map), also prevailed during the European medieval climatic optimum, at about 1100 to 1300 A.D. (KLIMANOV, 1987; ELOVICHEVA and BOGDEL, 1987; FRENZEL *et al.*, 1989). This demonstrates that only minor changes in summer temperatures have occurred in Europe during most of the Holocene, as shown by the relatively weak changes in the extension of Holocene glaciers in the Alps

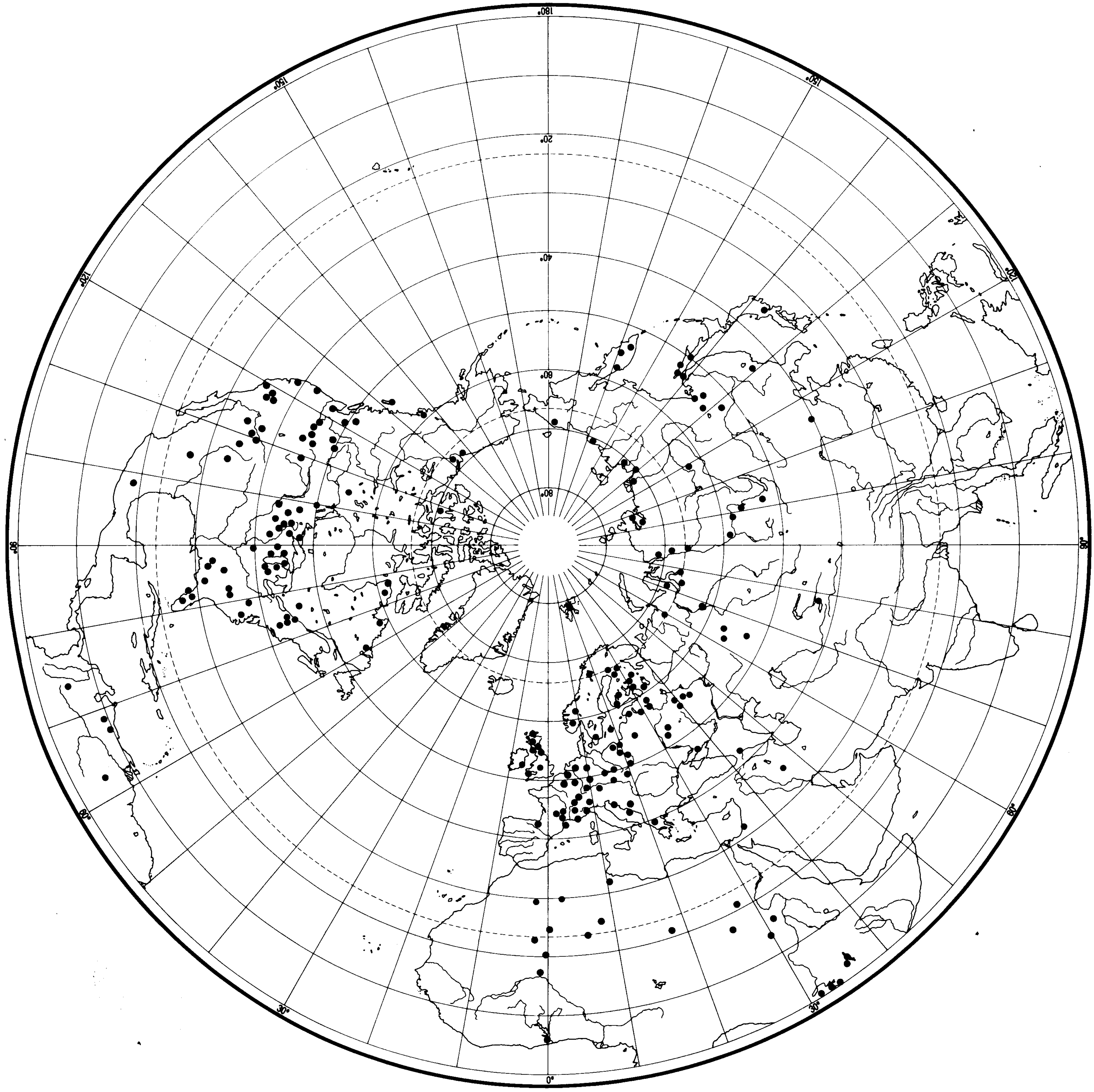

Fig. 15. Sites used for the construction of maps on pages 67, 71, 75 and 79

and in other mountain systems of the Northern Hemisphere (PATZELT, 1977; BORTENSCHLAGER, 1977; HOLZHAUSER, 1984; WETTER, 1987; MAISCH, 1982; VUAGNEUX, 1983; HANNSS and WEGMÜLLER, 1976). Yet, there existed exceptions from this general pattern, particularly in the subpolar regions, as stated earlier.

As shown in map on page 67, winter temperatures were appreciably higher at about 6,500 years B.P. in Eastern Europe than in other parts of Europe. This was explained by more frequent invasions of humid air masses from the north during winter. On the other hand, from map on page 69 it becomes clear that the same did not occur during the summer season. This leads to the suggestion that summer climate was not controlled by moist air masses coming from the White Sea. Thus, the suggestion that winter precipitation in Eastern Europe had increased around 6,500 years B.P. can be corroborated.

Annual mean air temperatures (map on page 75). For the construction of this map indicators of climate different from those applied for the previous two maps on Holocene climates were used. This approach was already mentioned in the explanations to the maps of annual mean air temperatures of the Last Interglacial optimum (map on page 23), of the time of inland ice formation (map on page 33), and of the glacial maximum of the Last Glaciation (map on page 43). Moreover, the former distribution pattern of permafrost in Siberia was also involved in the reconstruction of Middle Holocene annual mean air temperatures.

As shown in map 75, mean annual air temperatures were about 1.5 °C higher than today nearly everywhere on the Northern Hemisphere, between the equator and the north-boreal regions. From this general pattern, a deviation existed only within the subpolar belt, also shown in maps on pages 67 and 71, since in the subpolar region annual mean air temperatures were somewhat higher than those farther south, showing positive deviations up to about 2 °C, with maximal values in the northern part of Eastern Europe.

There did exist some local exceptions from this general picture, which can be attributed either to the different methods used for reconstructing mean annual air temperatures of the past, or to the specific distribution patterns of certain plant taxa. For example, to the north of the mouth of river Amur, either unexpectedly strong advances of oak forests had occurred at the time in concern, or in these areas the present-day distribution of this genus is controlled predominantly by human impact.

Annual mean precipitation (map on page 79). Indicators of climate used for the compilation of this map are former distribution patterns of various plant and animal taxa, of plant communities, lake sediments, evidence of lake-level oscillations, stable isotopes in groundwaters and distribution patterns of certain soil types.

There is a recurrent assumption in present-day discussion about the future development of climate that an increase in air temperatures of about 2 to 4 °C will be accompanied by a growing aridity in present-day semiarid regions of the world. As was already discussed in the explanatory text to map on page 27, this did not occur during the optimum of the Last Interglacial, when the moisture supply in recent semiarid and arid regions in most cases was more abundant than nowadays. In this context it is extremely interesting to learn from map on page 79 that the same also happened during the Holocene climatic optimum. There exists a wealth of information about climates with increased humidity all over the Sahara, in the Mediterranean region and in the steppe areas of Eastern Europe.

There was a remarkable increase in moisture available in Eastern Europe. It happened in the same regions that experienced a marked rise in winter temperatures at that time. From these facts it may be deduced that the amount of winter precipitation must have been appreciably higher than presently. On the other hand, a conclusion can be drawn from the map showing the deviations of summer temperatures from today's values at about 6,500 years B.P., that higher summer precipitation was very unlikely then. Since there was a significant increase in available moisture and winter temperatures had ameliorated considerably in Eastern Europe, it is suggested that extensive snowfalls during the winter season supplied substantial moisture, and contributed to increased groundwater resources in this region.

Just the opposite happened during Middle Holocene times in the prairie regions of North America and in Central Asia. By comparing map on page 27 with the one on page 79 it becomes evident that during the Holocene climatic optimum both these regions received substantially less moisture than today. During the Last Interglacial, however, the climatic optimum in Central Asia was more humid than at present. This is an indication that circulation patterns during the two interglacial periods differed somewhat. Yet in general, an increase in temperatures, even at levels expected in the near future, would not necessarily be accompanied by drier climates in present-day semiarid and arid regions. Just the opposite can be predicted in general, though with regional exceptions.

CLIMATES BETWEEN 6,000 AND 5,500 YR B.P.

by A. A. Velichko, V. A. Klimanov and I. I. Borzenkova

(Explanatory notes to maps on pages 65, 69, 73 and 77)

General remarks. A large quantity of empirical data is now available (e.g., pollen diagrams, paleontological and archaeological remains dated by radiocarbon, lake-level records, etc.). They permit to try a **reconstruction of past climates** on a global scale. At present, a series of methods exists, which enables us to estimate quantitative climatic parameters (first of all, ground-level air temperatures and annual precipitation) on the basis of qualitative geological data. Paleobiological methods, which rely on species composition of fossil flora or microfauna at the appropriate levels in dated geological sections or deep sea cores, are the most widely applied ones. This group of methods includes: (1) the **paleofloristic method**, introduced by SZAFER and IVERSEN and developed by GRICHUK (see the chapter on the climates of the Last Interglacial optimum), (2) the **zonal method** (elaborated by SAVINA and KHOTINSKY, 1984) based on the reconstruction of zonal types of past vegetation and identification of their modern analogues (the climatic indices of the latter are extrapolated into the past), (3) the **information statistical method** based on correlation of **pollen spectra** with climatic conditions (KLIMANOV, 1976 and others). A number of different techniques was developed by WEBB and BRYSON (1972) and others. Almost all the approaches were applied to reconstructions of air temperatures and precipitation for the time period between 6,000 and 5,500 years B.P. The accuracy of the results in practically all the methods appeared to be very similar — about ± 0.5 to 1 °C for temperatures and ± 25 mm for annual precipitation.

According to the data obtained, a rather high heat supply was characteristic for the interval between 8,500 and 5,500 B.P., i.e. within the entire Atlantic period. In some regions the peaks of the heat supply occurred in the late Atlantic (7,000 to 5,500 B.P.), in other regions their chronological position was somewhat different.

The Holocene climatic optimum is defined by the highest mean global temperatures. The latter were calculated to have been 1° C above the present-day values between 6,000 and 5,500 and thus this interval was accepted as representing the Holocene optimum.

Main sources of data. For the natural-climatic reconstructions radiocarbon-dated pollen diagrams were used, including those published and summarized by WEBB and BRYSON (1972), WRIGHT, Jr. (1976), DELCOURT and DELCOURT (1977) and WATTS (1980). Reconstructions for northern Africa and adjacent territories were mostly based on the archaeological materials, pollen diagrams, and data on lake-level fluctuations summarized by NICHOLSON and FLOHN (1980), NICOLE (1988) and others. For Western Europe, data from MÖRNER (1980), GUIOT (1987) and others were used. Altogether, more than 400 sources were involved in the process of map compilation *(Fig. 16)*.

Quantitative characteristics were estimated applying the zonal method developed by SAVINA and KHOTINSKY (1984). Although these evaluations are less accurate than those calculated using the information statistical technique or climagrams, the zonal method gives reliable information on the shifts in zonal boundaries. In addition to those data, some estimates of temperatures and precipitation for Western Europe and North America were considered, calculated by other authors using various transfer functions. The territory of the USSR was reconstructed according to KLIMANOV (1978) for the European part and MURATOVA and SUETOVA (1983) for Siberia. To this some new material for the northwestern part of the country, northern Siberia, and the Far East was added. Air temperatures over oceans were calculated from planktonic microfossils via factor analysis for the North Atlantic, and from other biological data for the Pacific.

Paleobotanic data supplemented by information on lake level changes in closed basins, archaeological, and other proxy data which are especially important for arid regions where palynological evidences are scanty or completely absent, were used to reconstruct the precipitation values.

Winter temperatures (map on page 65). Analysis of winter (January) and summer (July) temperature maps of the Late Atlantic warming (6,000 to 5,000 years B.P.) revealed significant deviations from present-day values in high latitudes, and showed that winter temperature differences were more significant than those of the summer temperatures. The position of the summer isotherms was nearly latitudinal; on the other hand, the pattern of winter temperatures was more complex. Three large regions are outlined where the temperature was 3 to 4 °C higher than today. One of them, centred in northern Canada and Greenland, covered a considerable area of the North Atlantic. The other two were located in the Asiatic part of the USSR, one in the northeast (with its centre near Yakutsk) and the other in the Kazakhstan-Caspian region.

In Western Europe and in the central Russian Plain, winter warming did not exceed 2 or 3 °C, compared with present-day temperatures. Similar deviations were reported from Canada and Alaska. In the USA and southern Europe (including Spain, Italy, Hungary, Bulgaria and Turkey) the warming was comparatively low, and did not exceed 1 °C.

Summer temperatures (map on page 69). The deviations in July temperatures were largest (up to 4 °C) in high latitudes north of 65° N, decreased in middle latitudes to 1 or 2 °C, and were negative further to the south. Temperature lowering was most pronounced in central Sahara, Central Asia, and in the monsoon regions of India and Indonesia. This was probably due to higher humidity in these regions than observed today.

Annual mean temperatures (map on page 73). Using the map of annual mean temperature deviations from the present-day values, the following latitudinal mean deviations were calculated for the Northern Hemisphere during the Holocene climatic optimum *(Table 9)*.

Table 9. Latitudinal mean deviations of the annual mean temperatures during the Holocene climatic optimum (6,000 to 5,500 yr B.P.) from present-day values

Latitudes	90°	80°	70°	60°	50°	40°	30°	20°	10°	0°
Δ Tyr [°C]	4.5	6.1	4.3	2.5	1.2		1.1	0.8	-1	-1

Taking the surface areas of the latitudinal belts into consideration , the annual mean global temperature deviation from the present-day conditions was calculated for the Northern Hemisphere. It amounted to about 0.6 to 0.7 °C in comparison to the period between 1881 and 1959 (Fiziko-geograficheskij atlas mira, 1964).

Annual mean precipitation (map on page 77). The reconstructed values correspond to a global warming of about 1 °C. North of 60° N the annual precipitation was approximately 50 to 75 mm higher than at present. Further south, over most parts of Central and Western Europe, in Southern Eastern Europe, and in western Siberia, precipitation decreased by about the same value. Most significant changes in precipitation occurred in present-day dry subtropical (monsoon) regions of Asia, Africa and America, where increases of 200 to 400 mm per year were estimated in comparison to 50 mm per year precipitation observed at present. A significant rise in the moisture supply was also recorded in some areas of Soviet Central Asia (Ustyurt Plateau, Mangishlak Peninsula, Kyzyl Kum desert) which are arid today. Here many archaeological sites date from 9,000 to 5,000 years B.P., indicating that this area was inhabited during the Holocene climatic optimum (MAMEDOV, 1982; VARUSHCHENKO, 1984). In North America, south of 55° N, precipitation was considerably less than at present (by 200 to 250 mm per year). Here the whole warm climatic phase of the Holocene was characterized by extremely low lake levels (Great Basin) and by the expansion of prairies into the Great Lakes region.

A comparison of maps on past moisture conditions within different zones of the Northern Hemisphere throughout the Late Glacial and the Holocene (BORZENKOVA, 1980; BORZENKOVA and ZUBAKOV, 1984) indicates a certain relationship between global temperature changes and the distribution pattern of precipi-

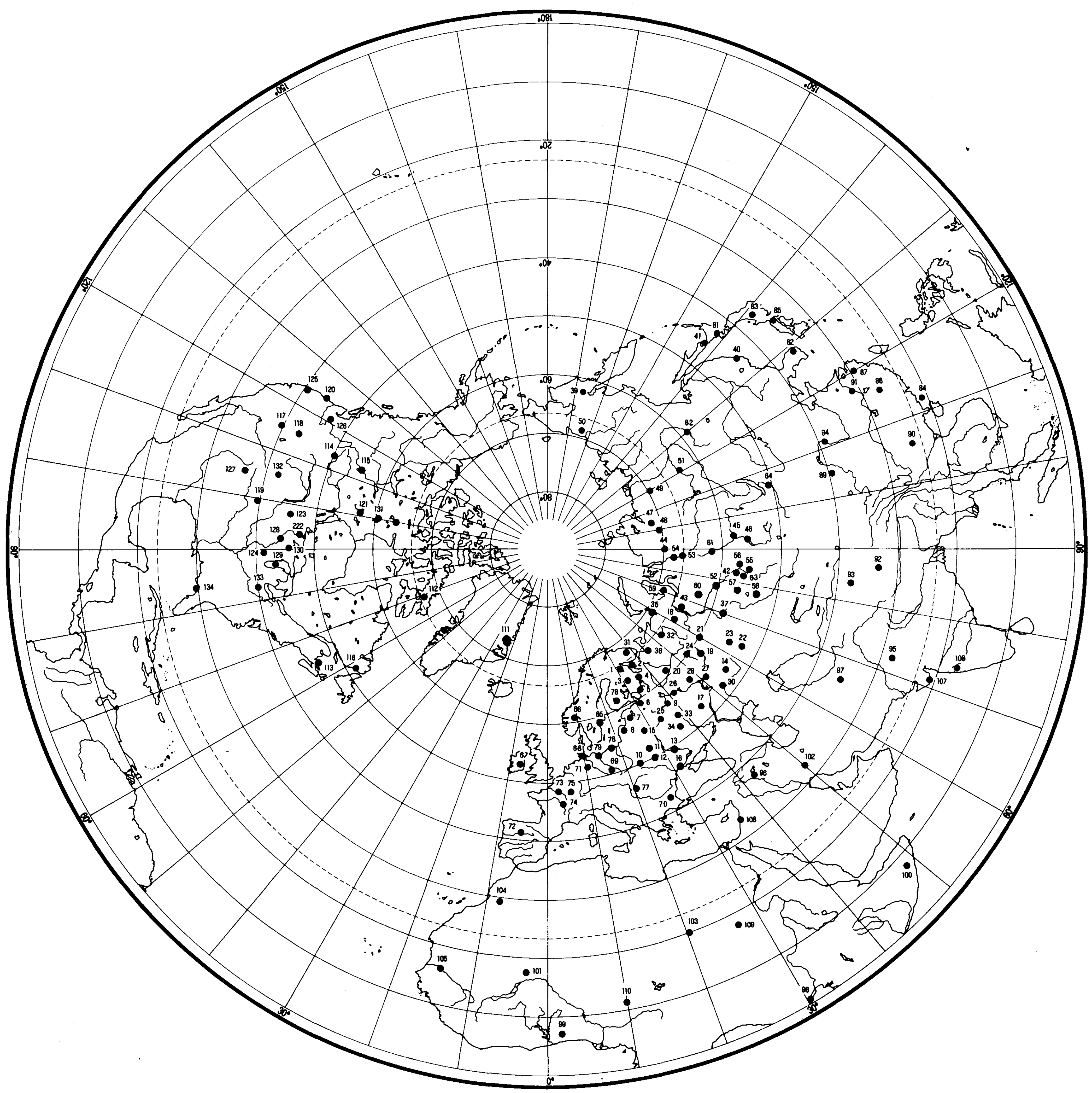

Fig. 16. Sites used for the construction of maps on pages 65, 69, 73 and 77

tation. The global rises in temperature was accompanied by an increase in precipitation in tropical (10° to 30° N) and northern (north of 60° N) regions. At the same time, a certain decrease in the moisture supply is observed in the zone between 50° and 30° N. This interrelationship between the global thermal regions and humidity, however, is far from simple, and it is controlled primarily by macro- and microclimatic conditions. The fact that the poleward temperature gradient would decrease during global warmings, and that this may have resulted in the southward shift of the circulation boundaries of various climatic zones, is the most likely explanation for this temperature — precipitation correlation. These mechanisms probably caused a corresponding migration of the **atmospheric centers** and changes in their strength.

Fig. 16. (Continuation) Deviations from present-day values: ΔT_I = January mean temperature; DT_{VII} = July mean tempereture; ΔT_Y = annual mean temperature; ΔP = annual precipitation

No	ΔT_I	ΔT_{VII}	ΔT_Y	ΔP_{mm}	No	ΔT_I	ΔT_{VII}	ΔT_Y	ΔP_{mm}	No	ΔT_I	ΔT_{VII}	ΔT_Y	ΔP_{mm}
1	2+3	3	2+3	25	46	2	0+0.5	1.5	0+25	91	-0.5	2	0	50
2	2+3	2+3	2+3	0+25	47	2	3+4	3	100	92	-1.5	1.5	-0.5	300
3	2	2	2	+0	48	2	3+4	3	100	93	-1.5	2	0	250
4	2	2	2	+0	49	2	1+2	2	75	94	-1.5	1.5	-0.5	250
5	1.5+2	1+2	1.5	0	50	1	1+2	1.5	25	95	-	-	-	100-200
6	1+2	1+2	1+2	0	51	2	1	1.5	25+50	96	-	-	-	+
7	1+2	1+2	1.5	-0	52	2	2	2	25+50	97	-	-	-	+
8	1+2	1+2	1.5	-0	53	1+2	2+3	2.5	50+75	98	-	-	-	+
9	1+2	1+2	1.5	0±25	54	1.5+2	2+3	2+3	50+75	99	-	-	-	+
10	1	1	1	+0	55	1+2	1	1+1.5	0	100	-	-	-	+
11	1	1	1	-0	56	1+2	0+0.5	1+1.5	0	101	-	-	-	+
12	1	1	1	0	57	1	1	1	-25+-50	102	-	-	-	+
13	0.5	0.5	1	0	58	0.5+1	0.5+1	1	0	103	-	-	-	150+350
14	2+3	2	2.5	+0	59	2	4+5	3+4	100+125	104	-	-	-	200+400
15	1+2	1+2	1.5	0	60	2+2,5	2	2	25+50	105	-	-	-	+
16	0+1	0	0.5	25+50	61	2+3	0+1	1+2	0+25	106	-	-	-	+
17	3	0+1	2	0+25	62	1+2	0.5+1	1.5	25+50	107	-	-	-	+
18	2+3	4	3.5	75+100	63	3	0+0.5	2	0	108	-	-	-	+
19	2	2	2	0±25	64	2	2+3	2.5	25+50	109	-	-	-	+
20	2	2	2	25+50	65	-	2+2.5	-	+	110	-	-	-	+
21	3+5	1+1.5	2+3	75	66	-	2+3	-	-	111	+	+	+	+
22	3+5	0+1	2+3	100	67	+	+	+	+	112	+	+	+	+
23	3	0.5	2	50+100	68	+	+	+	+	113	+	+	+	+
24	3+5	0+0.5	2+3	100+1500	69	+	+	1+2	+	114	+	+	1.9	-
25	2	1+2	1.5+2	±25	70	+	+	+	+	115	-	-	-	<0
26	2+3	2	2	-50	71	+	+	1	-	116	+	+	+	+
27	3	1+2	2	0	72	+	+	+	-	117	+	+	+	-
28	2+3	1+2	2	-25	73	+	+	+	+	118	+	+	+	<0
29	1+2	0.5	1.5	0	74	+	+	+	+	119	+	+	+	<0
30	3+4	0+1	2.5	50+75	75	0+4	0+4	2	+	120	-	1+2	-	-
31	2+3	3+4	3	75	76	+	+	+	-	121	-	3	-	-
32	3+4	4	3.5	100	77	+	+	+	-	123	-	-	-	<0
33	1+2	1+2	1+2	-0	78	+	+	+	+	124	-	0.5	-	-100-150
34	1+2	1	1.5	-25	79	-	-	-	+	125	+	+	+	-
35	3	4	3.5	100	80	-	1+2	-	-	126	+	+	+	<0
36	2+3	3	3	100	81	+	+	2+3	+	127	+	+	+	<0
37	2	1	1.5	50	82	+	+	2+3	+	128	-	1	-	-60
38	2	4+5	3	100+125	83	+	+	2	-	129	-	0.6	-	-25
39	1	1+2	1.5	25	84	-	-	1+3	-	130	-	0.9	-	<0
40	2	2+2.5	2	25	85	+	+	2	+	131	-	3	-	-
41	2	2	2	25	86	+	+	+	+	132	0	2	-	<0
42	1	0+1	1	0	87	+	+	2	+	133	0	1+2	-	+
43	1+2	2+4	3	50+75	88	+	+	3+4	-	134	0	0+1	-	+
44	2	3+4	3	50+100	89	-0.5	1.5	0	150	135	-	2+3	-	-
45	2+3	0+1	2	25	90	-1.5	0.5	-0.5-1	350					

ANNUAL MEAN RUNOFF DURING THE LAST INTERGLACIAL AND HOLOCENE CLIMATIC OPTIMA

by A. V. Belyaev and A. G. Georgiadi

Recent paleogeographical reconstructions of natural zonality and climatic characteristics for different regions of the Northern Hemisphere have provided prerequisites for regional quantitative reconstructions of river discharges (VELICHKO *et al.*, 1984; KLIMANOV, 1982; KHOTINSKY, 1978 and others).

A zonal principle widely used in geographical-hydrological studies by M.I. Lvovich's school has been applied as the basis for reconstructing the **spatial pattern of runoff** for the Northern Hemisphere in the past, with the involvement of paleofloristic data and of climatic indicators. In accordance with the above principle, this reconstruction is based on dependencies between annual river discharges and climatic characteristics, and can be established for distinct natural zones. Such dependencies reflect the average conditions most typical of a given zone. Previously the zonal principle had been used for evaluating the discharge of the Volga river in the past (VELICHKO *et al.*, 1988).

The reconstruction was accomplished on the basis of a system of relationships established for the major natural zones of the Northern Hemisphere as follows:

$$E_{max} = F(T) \;.\,.\,.\,.\,.\,.\,.\; (1)$$
$$K_R = F(K_{Emax}) \;.\,.\,.\,.\;(2)$$

where: E_{max} — annual potential evapotranspiration, mm; T — annual mean air temperature, °C; $K_R = R/P$, coefficient of the annual runoff, R — annual runoff, mm; P — annual precipitation, mm; $K_{Emax} = E_{max}/P$ — index of aridity.

The choice of such a system of relationships was needed to take into account two major climatic factors, **heat and moisture supply**, whose impact on river discharge are of opposite character. For the establishment of the equations (1) and (2) **hydroclimatic data** were used from the maps of the water balance elements from the monograph "The World's water balance and water resources of the Earth" (1974).

The map of vegetation zones of the Mikulino Interglacial optimum compiled by GRICHUK (1982) and the map of vegetation zones of the Holocene optimum for the USSR territory by KHOTINSKY (1978) were also involved in the reconstructions. Paleoclimatic data, part of which had been published earlier (VELICHKO et al.,1984a; KLIMANOV, 1982) were put at the disposal of the authors.

Figures 17 and *18* are maps of **deviations of the annual mean runoff** from present-day values (ΔR, mm) for the Last Interglacial optimum (about 125,000 years B.P.) and for the Holocene optimum (6,000 to 5,000 years B.P.), respectively.

In general it is assumed that positive deviations of runoff were caused by a strong influence of **increased precipitation**, surpassing the effect of the **rise in air temperatures**. Negative deviations were mainly caused by considerably higher air temperatures as compared to present values. In areas of substantial decline, the runoff was affected by both factors in the same direction, i.e., there was a rise in air temperature and a simultaneous decrease in atmospheric precipitation.

During the optimum of the Mikulino Interglacial the spatial pattern of runoff in Europe differed significantly from the present.

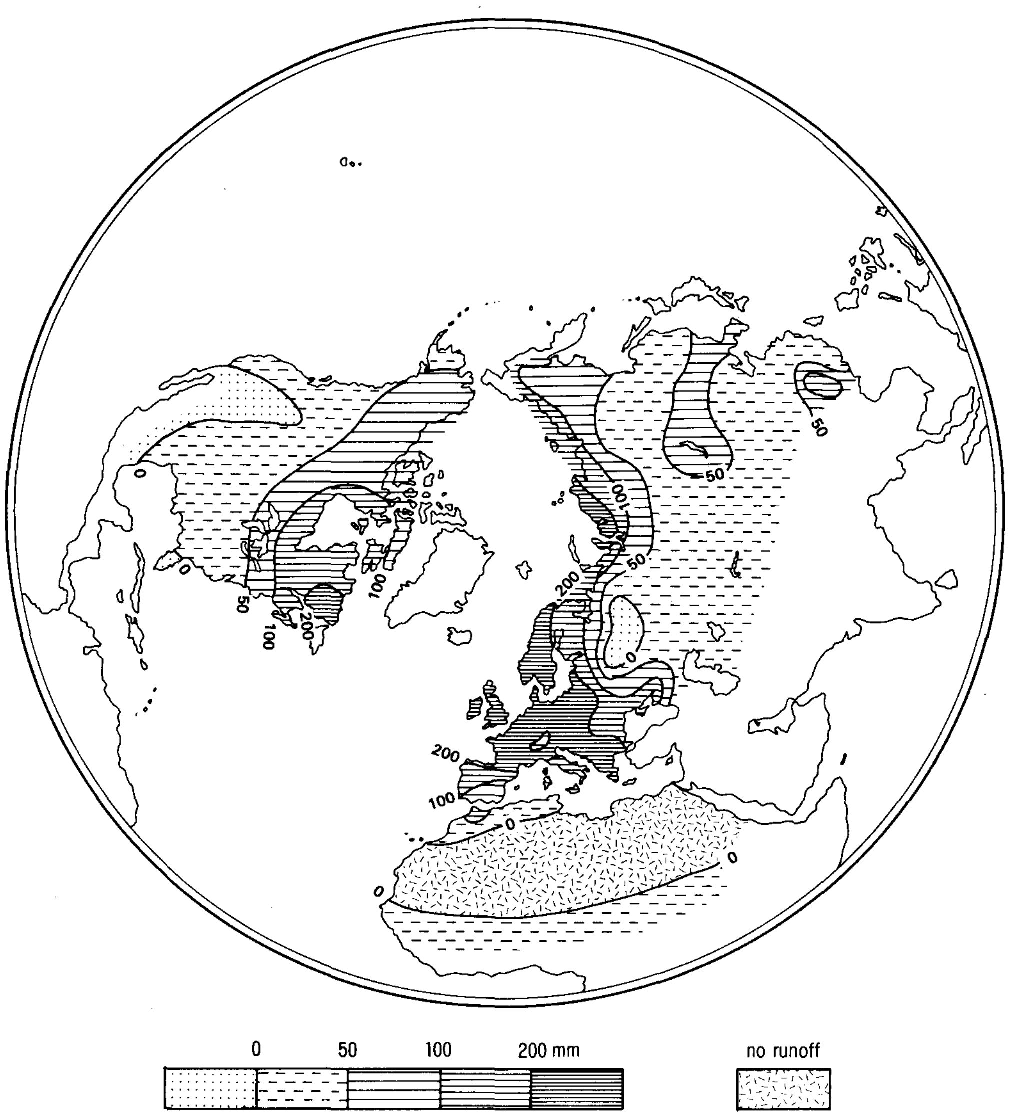

Fig. 17. Annual mean runoff during the Last Intergalacial climatic optimum (deviations from present-day values)

Maximum positive deviations (reaching 100 to 200 mm and more) were characteristic for nearly all of Europe west of 30° E. At the same time, in most areas of the Russian Plain the positive deviations from the present values did not exceed 50 mm. An extensive territory with relatively small negative deviations (up to 50 mm) occupied then approximately the same position as the one of maximum (over 50 mm) decrease of runoff later in the Holocene optimum. This decrease was observed despite the precipitation having been higher by 100 mm than at present, due to the rise in air temperatures by 3 to 4 °C compared to present-day values. In other regions, where the air temperatures were also higher, the increase in precipitation resulted in positive deviations.

During the Holocene optimum the runoff in a wide belt between 47° and 62° N was less than at present (not more than by 50 mm). Within this belt, in the Russian Plain (zone of central taiga and mixed forests) there existed an extensive area with a greater negative deviation (over 50 mm). The runoff pattern in the southern part of Europe (zone of steppe and sclerophylous forests) was similar to the present-day one showing only slightly higher values. A significantly positive deviation, amounting to 50 to 100 mm was found in a narrow belt of the present-day forest tundra and tundra.

During the Last Interglacial optimum the runoff in the overwhelming part of Asia was higher than at present. The maximum deviations (100 to 200 mm and more) were found north of 67° to

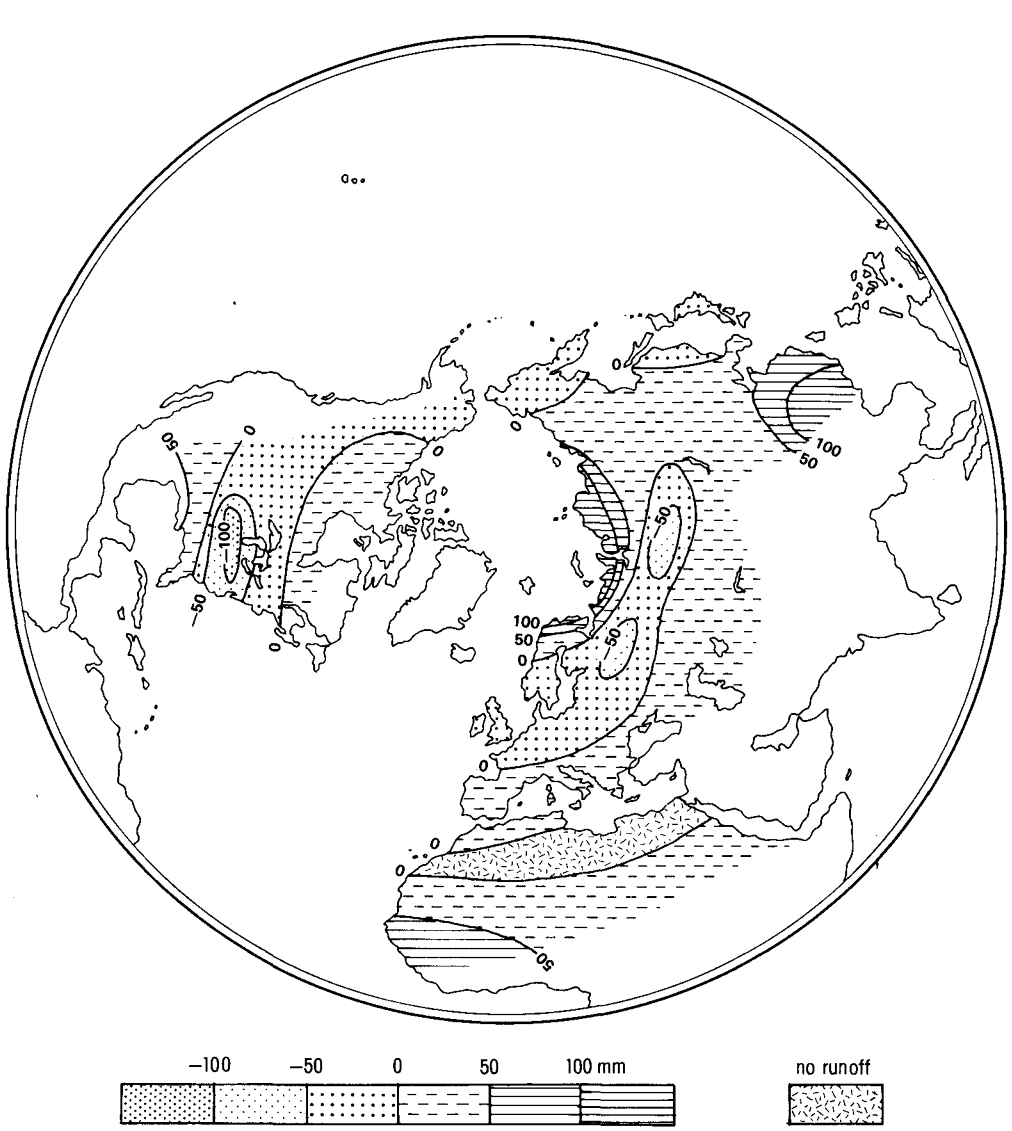

Fig 18. Annual mean runoff during the Holocene climatic optimum (deviations from present-day values)

70° N and east of 160° to 170° E, as well as in a limited in size area of humid mixed forests of East Asia. Over vast territories of the continent, occupying a 2 to 7 degree wide sublatitudinal belt in the north and in the eastern regions of the temperate and subtropical belts, the runoff showed higher values than at present by more than 50 mm. The observed large positive deviations were induced by a greater impact of the increase in precipitation compared to the effect of the rise in temperatures. For the major part of Asia positive deviations were insignificant and in many areas the values of runoff practically did not differ from their present-day analogues.

At the Holocene climatic optimum in the taiga zone of West and East Siberia, covering a 10 degree wide belt stretching from west to east along 50° N, the runoff showed lower values than at present (by 50 mm and more). The dimensions of this area are comparable with an analogous one which existed simultaneously in the Russian Plain, and resulted from a similar negative impact of temperature and precipitation. A minor decrease of the runoff was also observed along the northeastern margin of the Asiatic continent (east of 150° to 160° E) and in the area adjoining the Sea of Japan. The highest values of positive deviations (by 50 to 100 mm and more), induced by a more abundant precipitation at the Holocene optimum, are observed in the belt along the coasts of the Arctic Ocean and in the zones of forest steppe, broad-leaved forests, and humid mixed forests of East Asia (south of 40° N). For the rest of the temperate belt of Asia, runoff values of at the Holocene optimum were close to the present-day, surpassing them only slightly.

In North America, deviations of runoff during the Last Interglacial optimum, compared to the present, showed a pattern close to sublatitudinal distribution with the greatest positive values (100 to 200 mm and higher) in the northeast of Canada. Minor negative deviations were characteristic of the Midwest USA and Central America.

During the Holocene climatic optimum the runoff deviations in North America showed an analogous pattern with those of the temperate belt in Europe. At the same time, within a wide belt (roughly between 35° and 50° N) the runoff was lower than at present. Maximum negative deviations (50 mm and more) occupy an extensive area in the southeast of the temperate belt (zone of mixed and broad-leaved forests, forest steppe) and in the east of the subtropical belt (zone of humid mixed forests). To the north and south of this vast territory, the annual mean runoff was higher but positive deviations did not exceed 25 to 50 mm there.

At the Last Interglacial optimum in Africa the area of zero runoff practically coincided with the present-day one. Only in the north of the continent, within a narrow belt adjoining the Mediterranean Sea, positive deviations of runoff reached 25 to 50 mm and more.

During the Holocene optimum the area of zero runoff was very limited, although the increase over the greater part of the Sahara was rather small. Only south of 15° N and in the north of the continent became positive deviations considerable, surpassing 50 mm and more.

The above results are the first attempt at the evaluation of the annual mean runoff in the Northern Hemisphere for the past epochs. The maps compiled may also be considered as scenarios of a macroscale structure of the changes in discharge of the watercourses over the Northern Hemisphere that might be expected under a global warming by 2 °C (similar to the Last Interglacial climatic optimum) and by 1 °C (as at the Holocene optimum).

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INDEX

Numbers refer to pages of the explanatory text
Abbreviations: i/gl = Last Interglacial climatic optimum; i/st = interstadial; max. cooling = maximum cooling; gl. max. = glacial maximum (last pleniglacial); hol = Holocene climatic optimum

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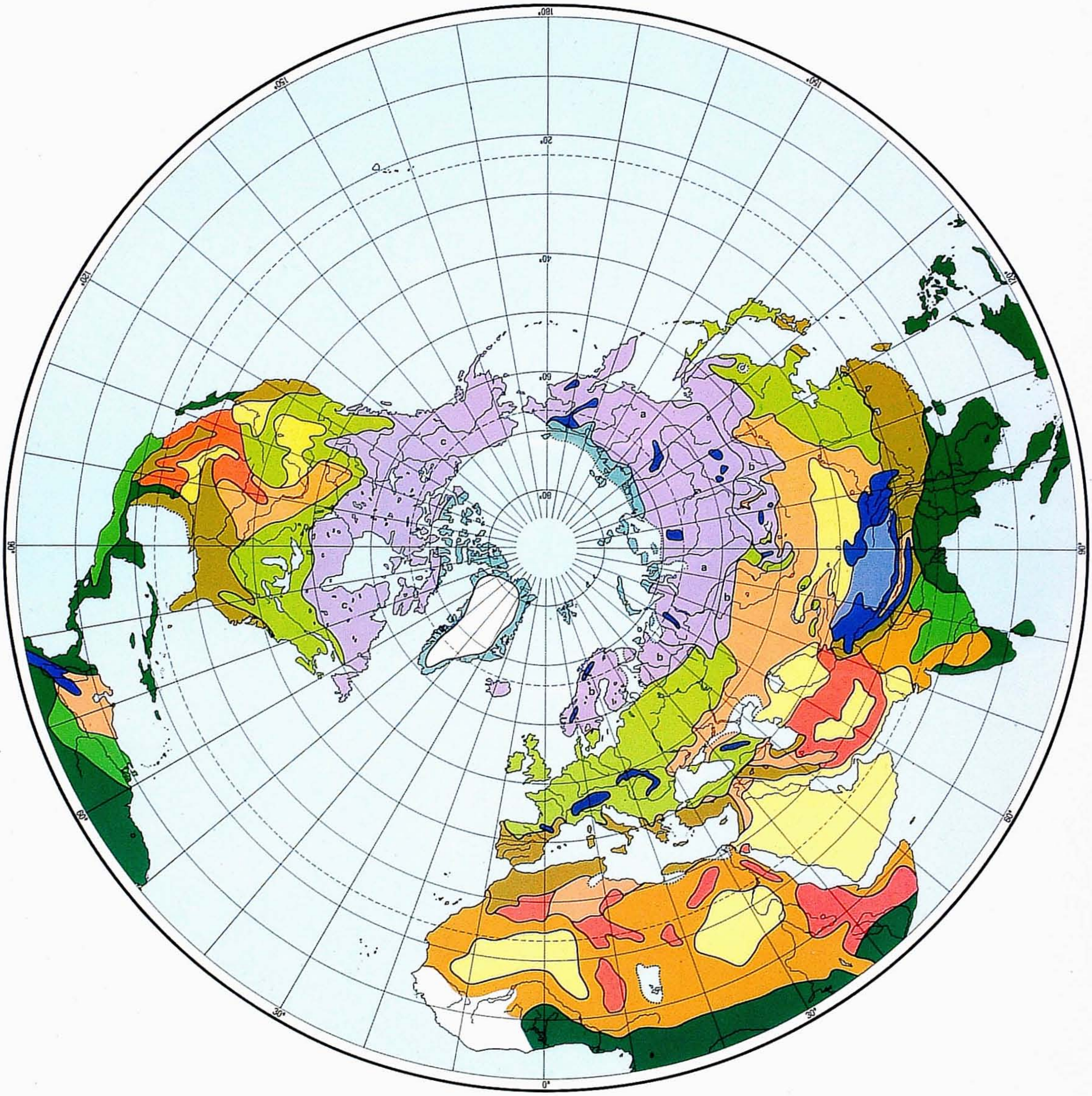
VEGETATION

(about 120,000 yr B.P.)

by V.P. Grichuk



- Herbs and bush, moss and bush formations of tundra type
- Forest formations of boreal type:
  - a Coniferous and coniferous-birch formations with patches of tundra associations
  - b Coniferous formations with broad-leaved species
  - c Coniferous formations with patches of tundra associations in the North, birch and broad-leaved associations in the South
- Species-rich broad-leaved formations in plains and mountain formations of nemoral type (broad-leaved species with conifers)
- Species-rich broad-leaved formations with evergreen and conifer trees
- Subxerophilous bush-arboreal formations of broad-leaved, deciduous and evergreen species with conifers, and formations of mountain xerophytes
- Subxerophilous and xerophilous herb-grass and Artemisia-grass formations of steppe type
- Xerophilous Artemisia-grass and bush formations of desert type on low-lying plains and in low mountains
- High mountain xerophilous vegetation of desert type
- Savanna and tropical open forests
- Seasonally wet evergreen and deciduous formations
- Constantly wet evergreen formations rich in species
- High mountain vegetation of tundra, alpine and andean types
- Coastline deviating from present-day position

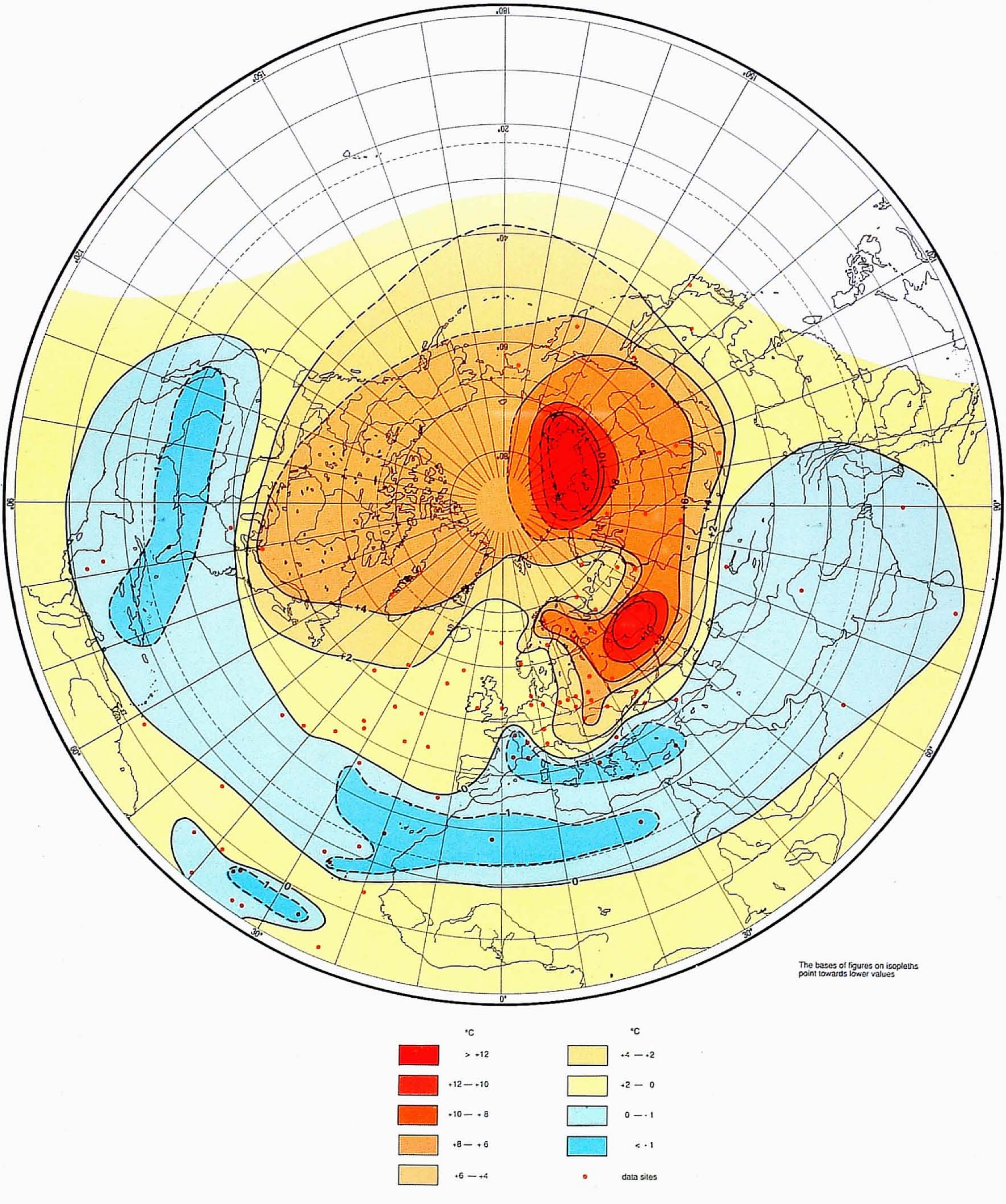
LAST INTERGLACIAL

(about 120,000 yr B.P.)

JANUARY MEAN TEMPERATURE

Deviations from present-day values

by A.A. Velichko, V.P. Grichuk, E.E. Gurtovaya,
E.M. Zelikson, O.K. Borisova, M.S. Barash



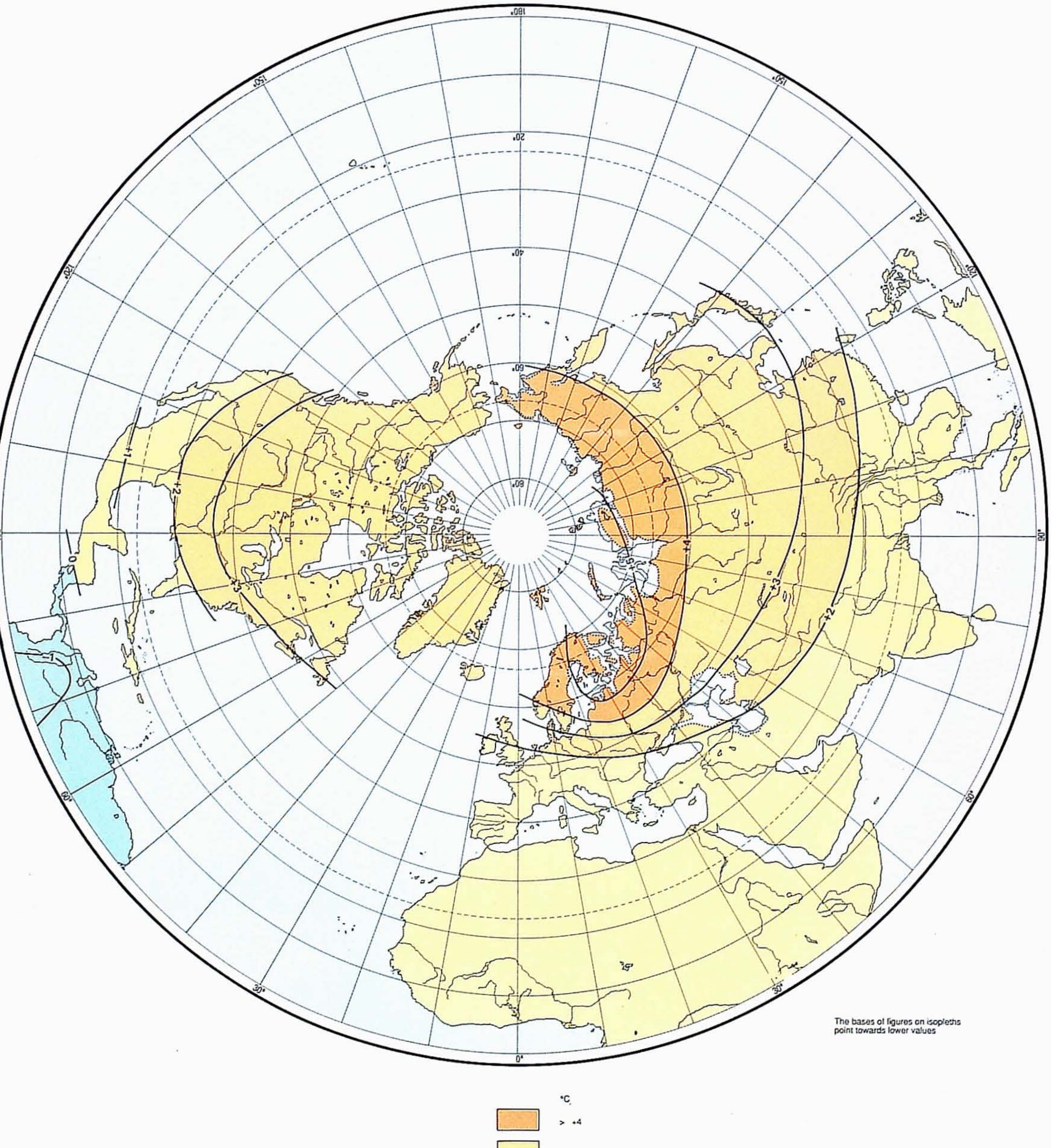
LAST INTERGLACIAL

(about 120,000 yr B.P.)

FEBRUARY MEAN TEMPERATURE

Minimal deviations from present-day values

by B. Frenzel



°C

> +4

+4 — + 2

+2 — 0

< 0

Coastline deviating from present-day position

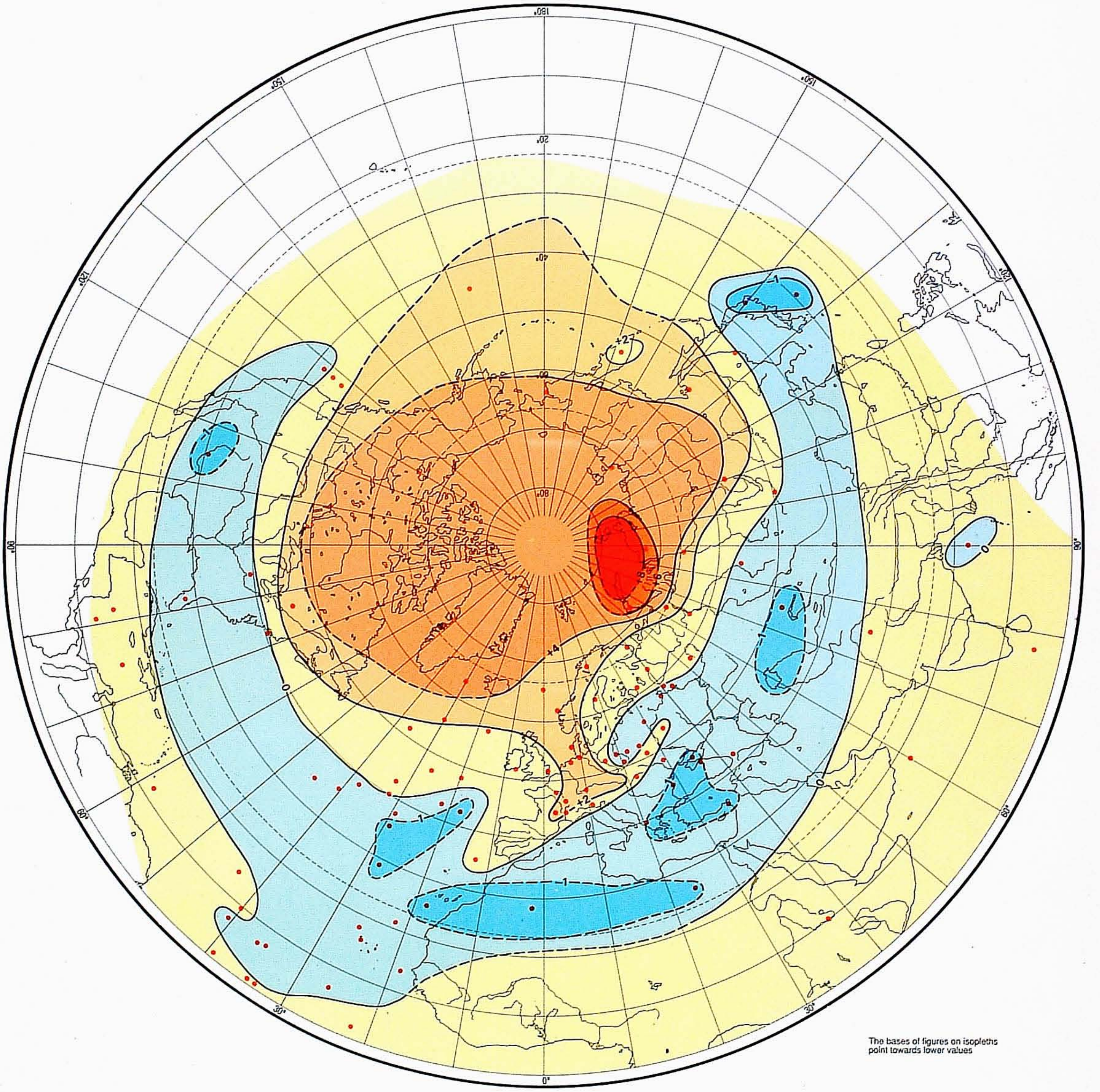
LAST INTERGLACIAL

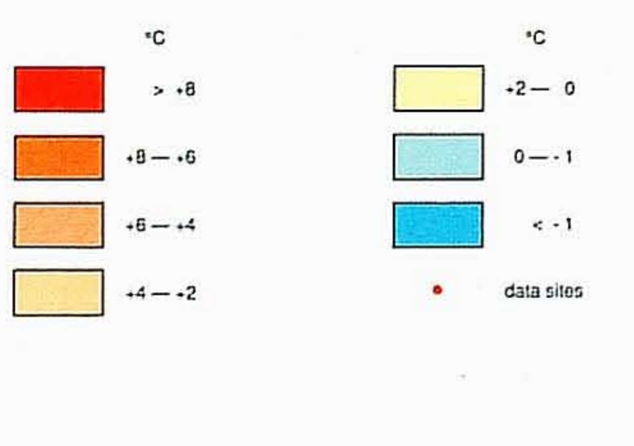
(about 120,000 yr B.P.)

JULY MEAN TEMPERATURE

Deviations from present-day values

by A.A. Velichko, V.P. Grichuk, E.E. Gurtovaya,
E.M. Zelikson, O.K. Borisova, M.S. Barash





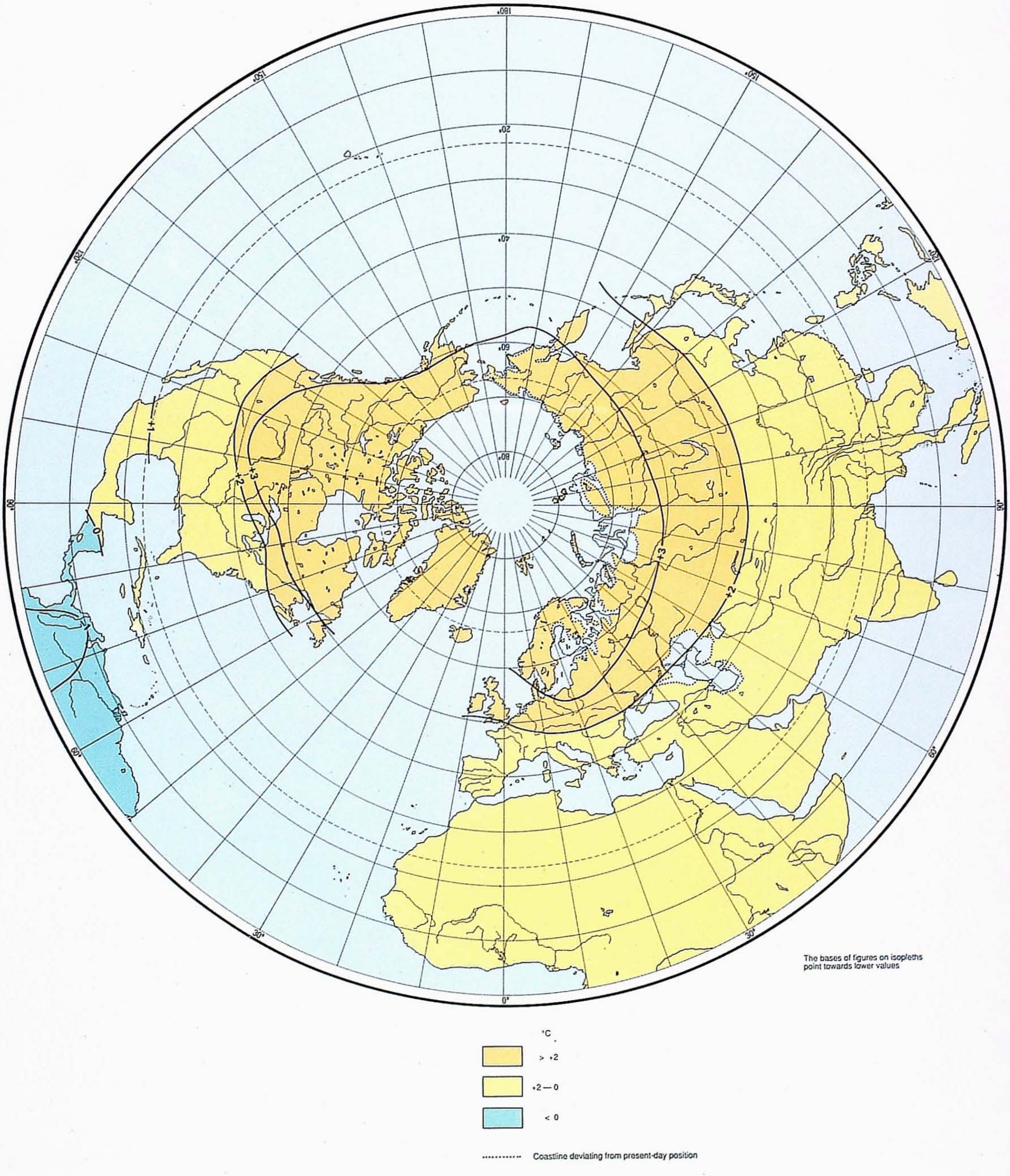
LAST INTERGLACIAL

(about 120,000 yr B.P.)

AUGUST MEAN TEMPERATURE

Minimal deviations from present-day values

by B. Frenzel



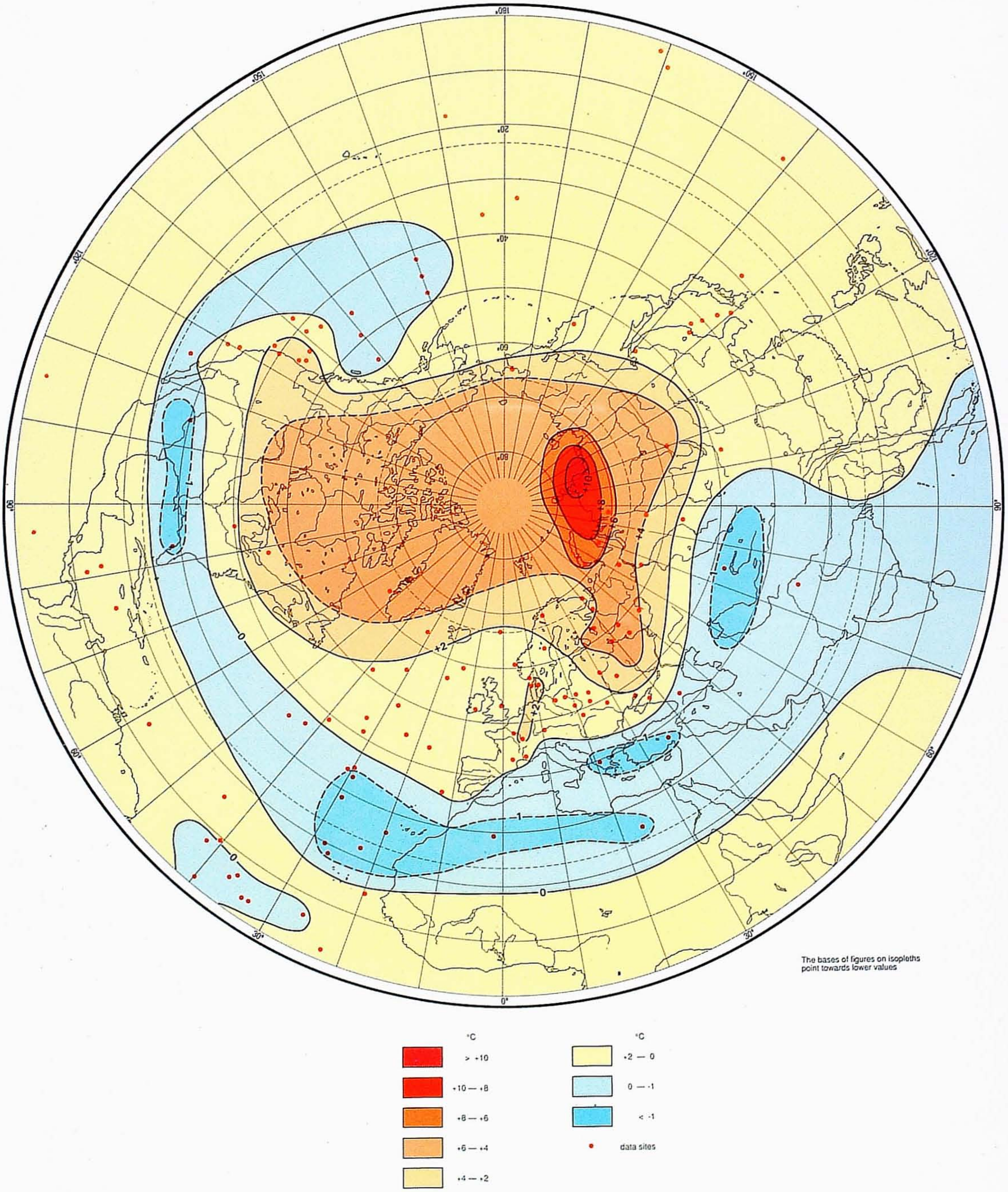
LAST INTERGLACIAL

(about 120,000 yr B.P.)

ANNUAL MEAN TEMPERATURE

Deviations from present-day values

by A.A. Velichko, V.P. Grichuk, E.E. Gurtovaya,
E.M. Zelikson, O.K. Borisova, M.S. Barash



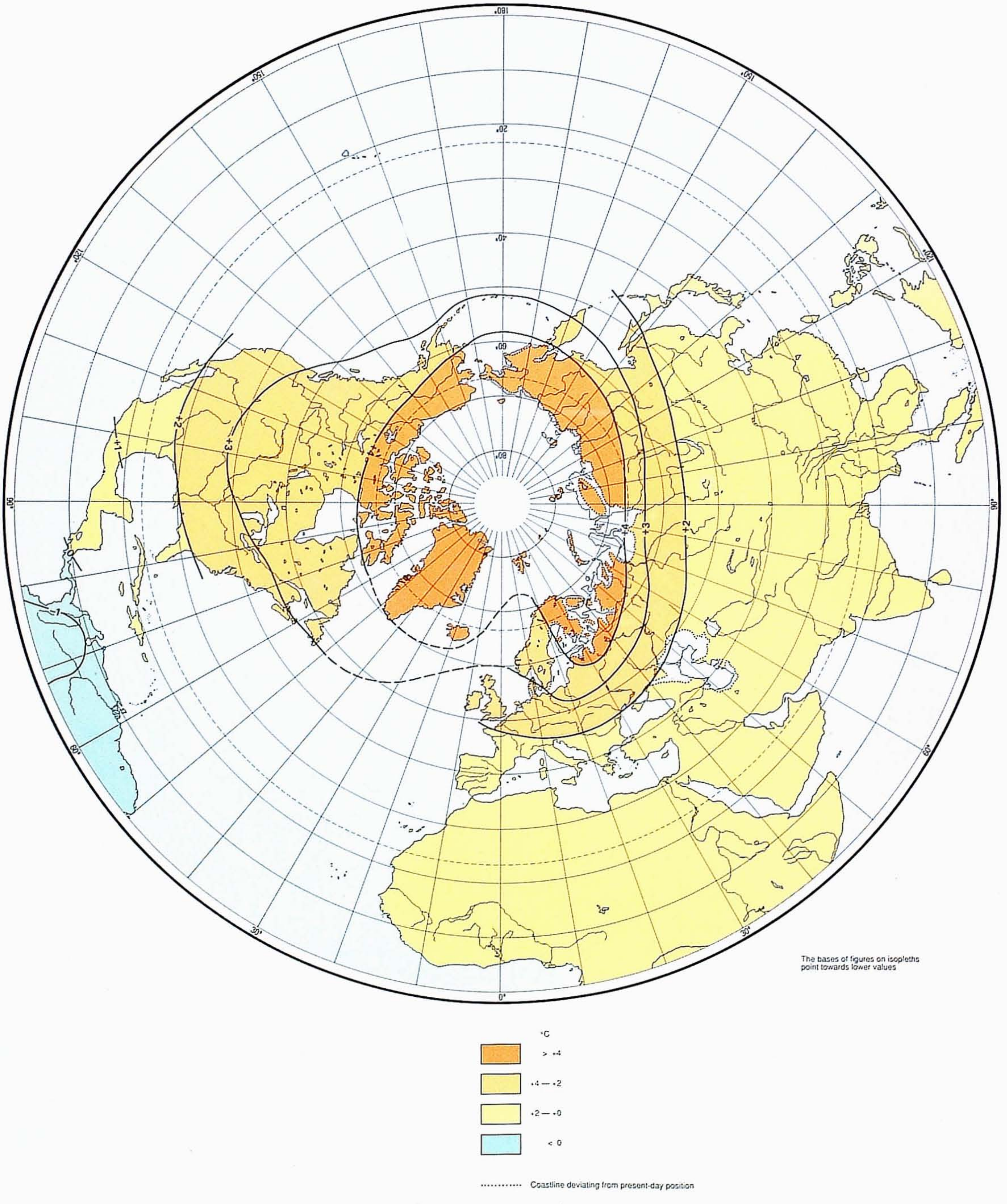
LAST INTERGLACIAL

(about 120,000 yr B.P.)

ANNUAL MEAN TEMPERATURE

Minimal deviations from present-day values

by B. Frenzel



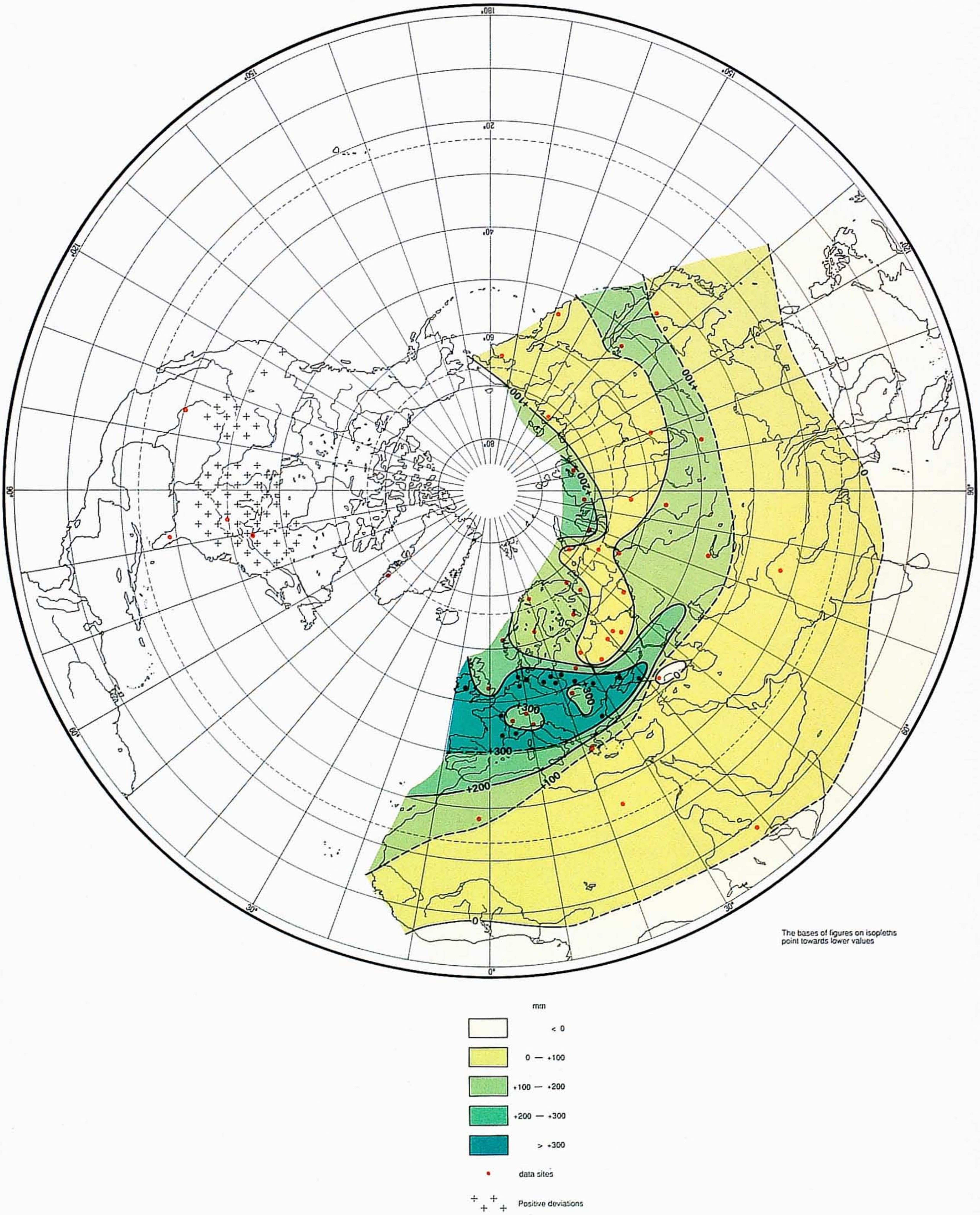
LAST INTERGLACIAL

(about 120,000 yr B.P.)

ANNUAL PRECIPITATION

Deviations from present-day values

by A.A. Velichko, V.P. Grichuk, E.E. Gurtovaya,
E.M. Zelikson, O.K. Borisova, M.S. Barash



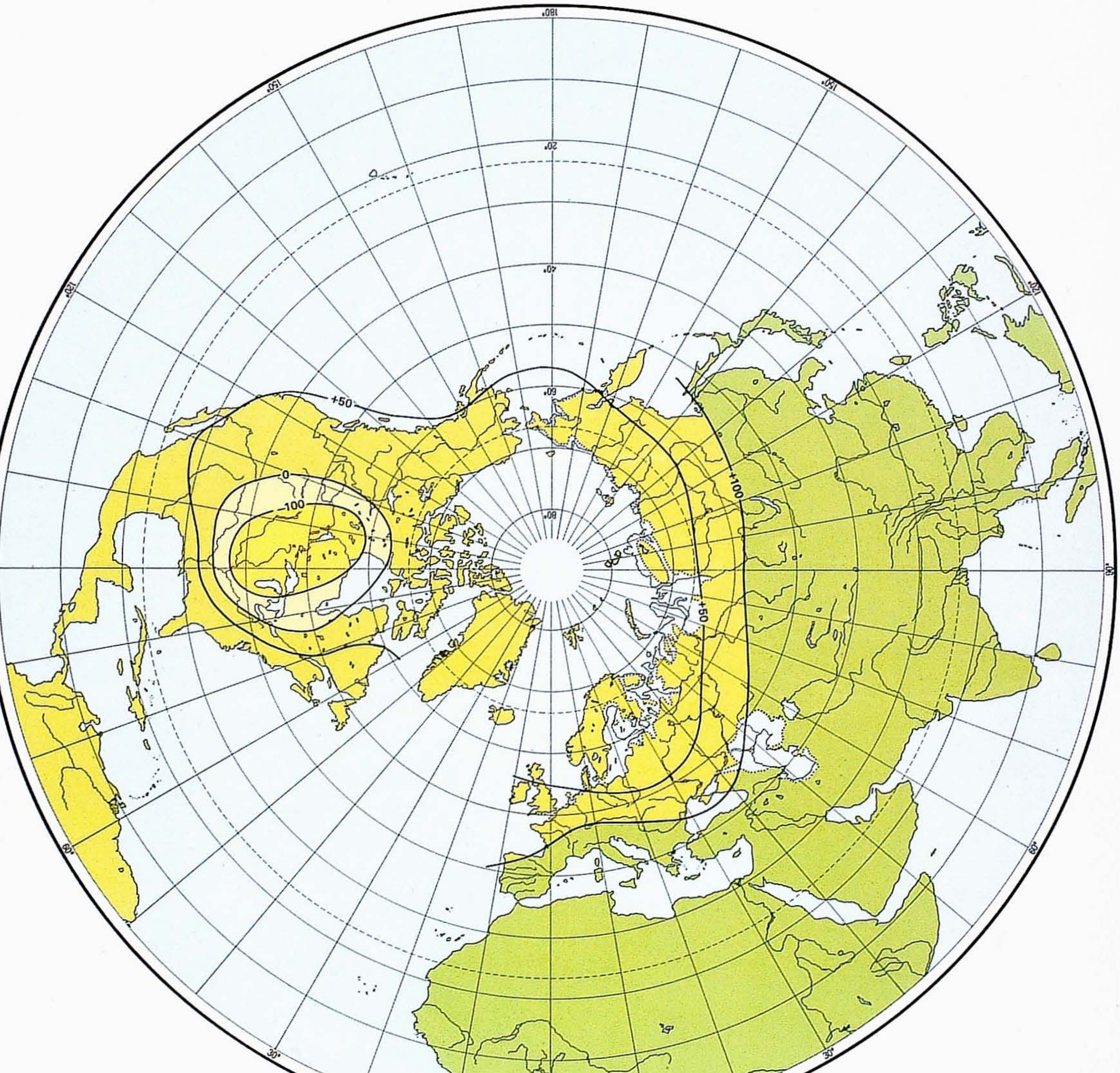
LAST INTERGLACIAL

(about 120,000 yr B.P.)

ANNUAL PRECIPITATION

Minimal deviations from present-day values

by B. Frenzel



The bases of figures on isopleths point towards lower values

mm

< -100

-100 — 0

0 — +100

> +100

.... Coastline deviating from present-day position

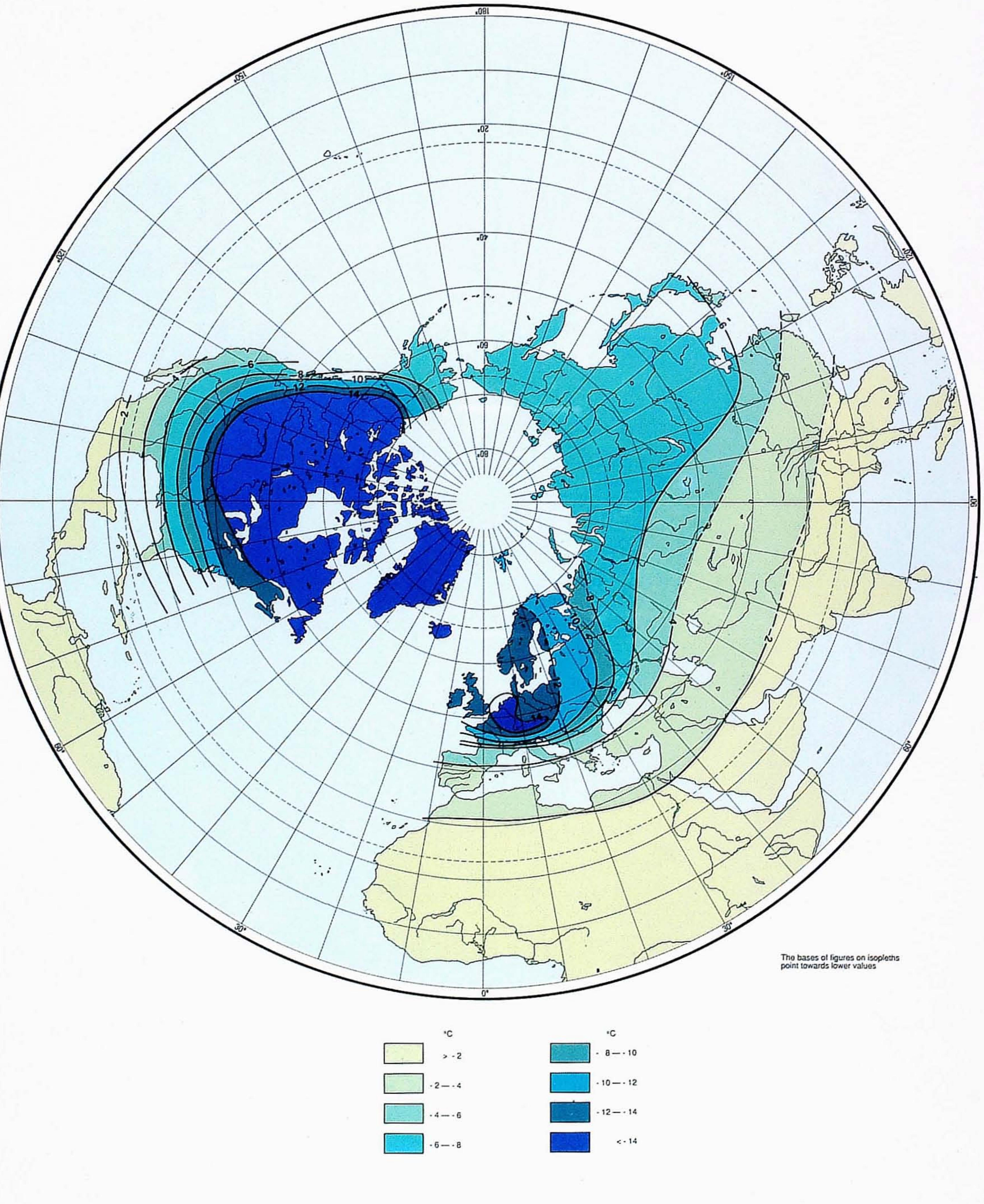
INTERSTADIAL OF THE
LAST GLACIATION

(about 35,000 to 25,000 yr B.P.)

FEBRUARY MEAN TEMPERATURE

Minimal deviations from present-day values

by B. Frenzel



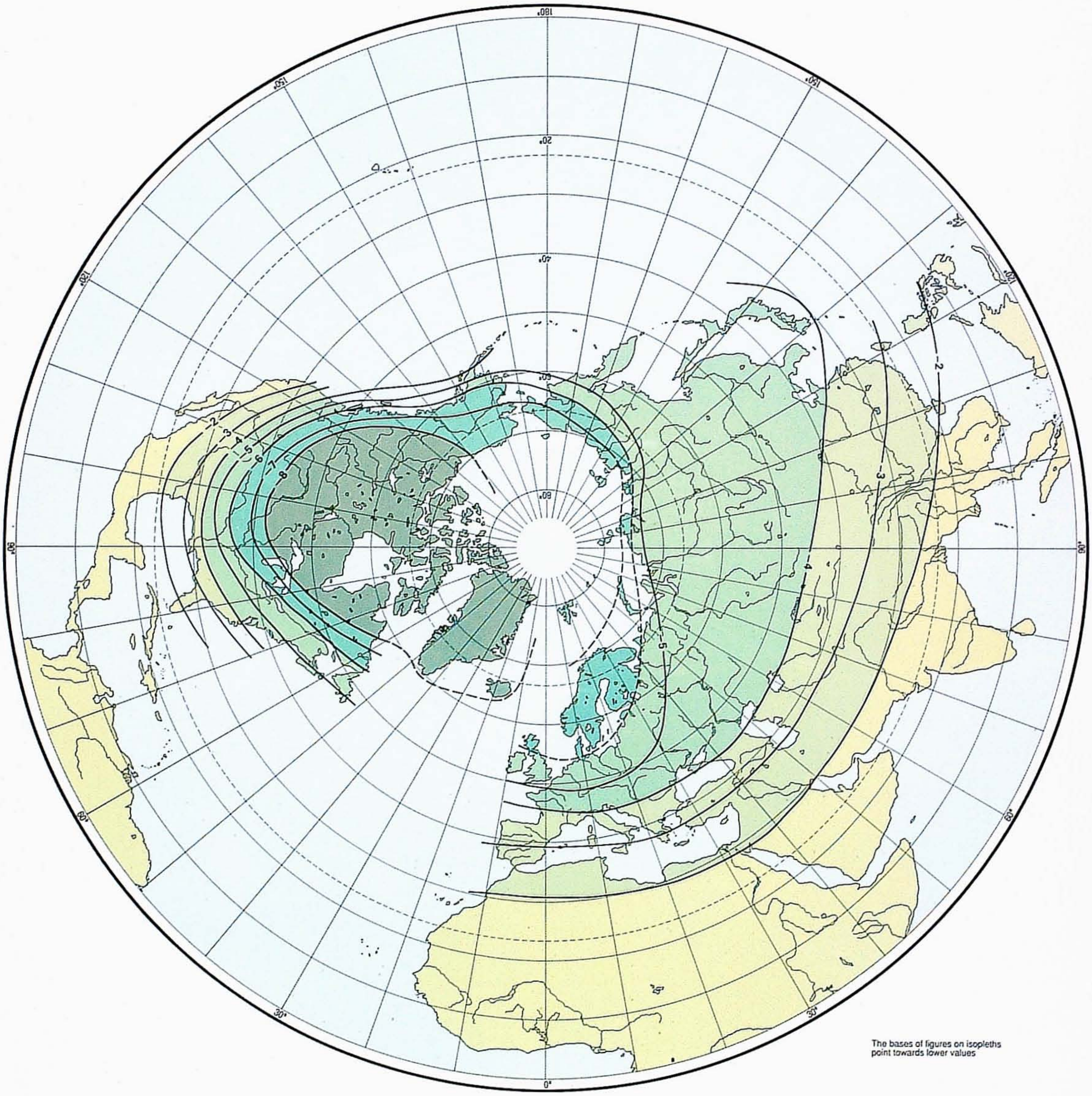
INTERSTADIAL OF THE LAST GLACIATION

(about 35,000 to 25,000 yr B.P.)

AUGUST MEAN TEMPERATURE

Minimal deviations from present-day values

by B. Frenzel



°C

	> -2
	-2 — -4
	-4 — -6
	-6 — -8
	< -8

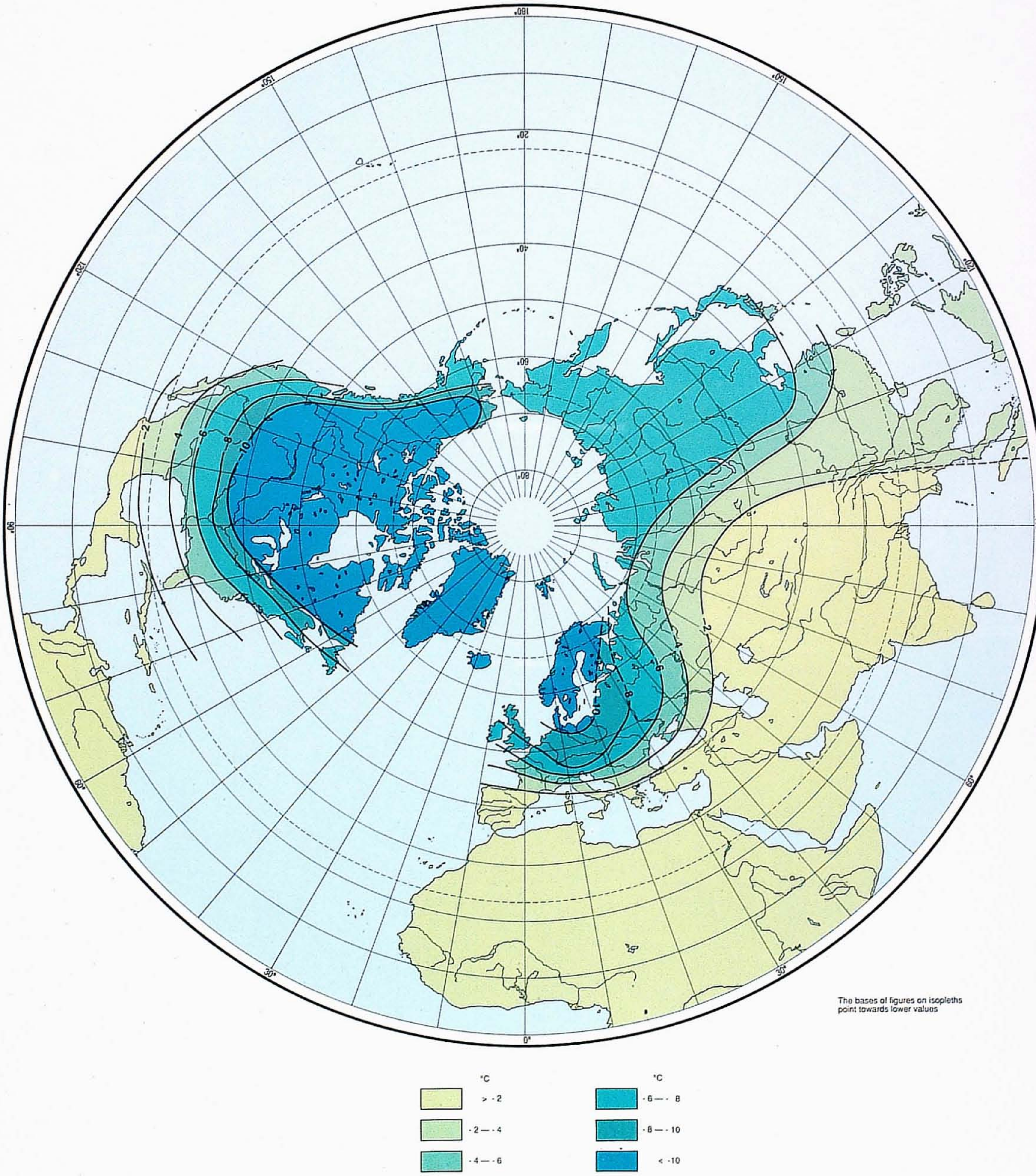
INTERSTADIAL OF THE
LAST GLACIATION

(about 35,000 to 25,000 yr B.P.)

ANNUAL MEAN TEMPERATURE

Minimal deviations from present-day values

by B. Frenzel



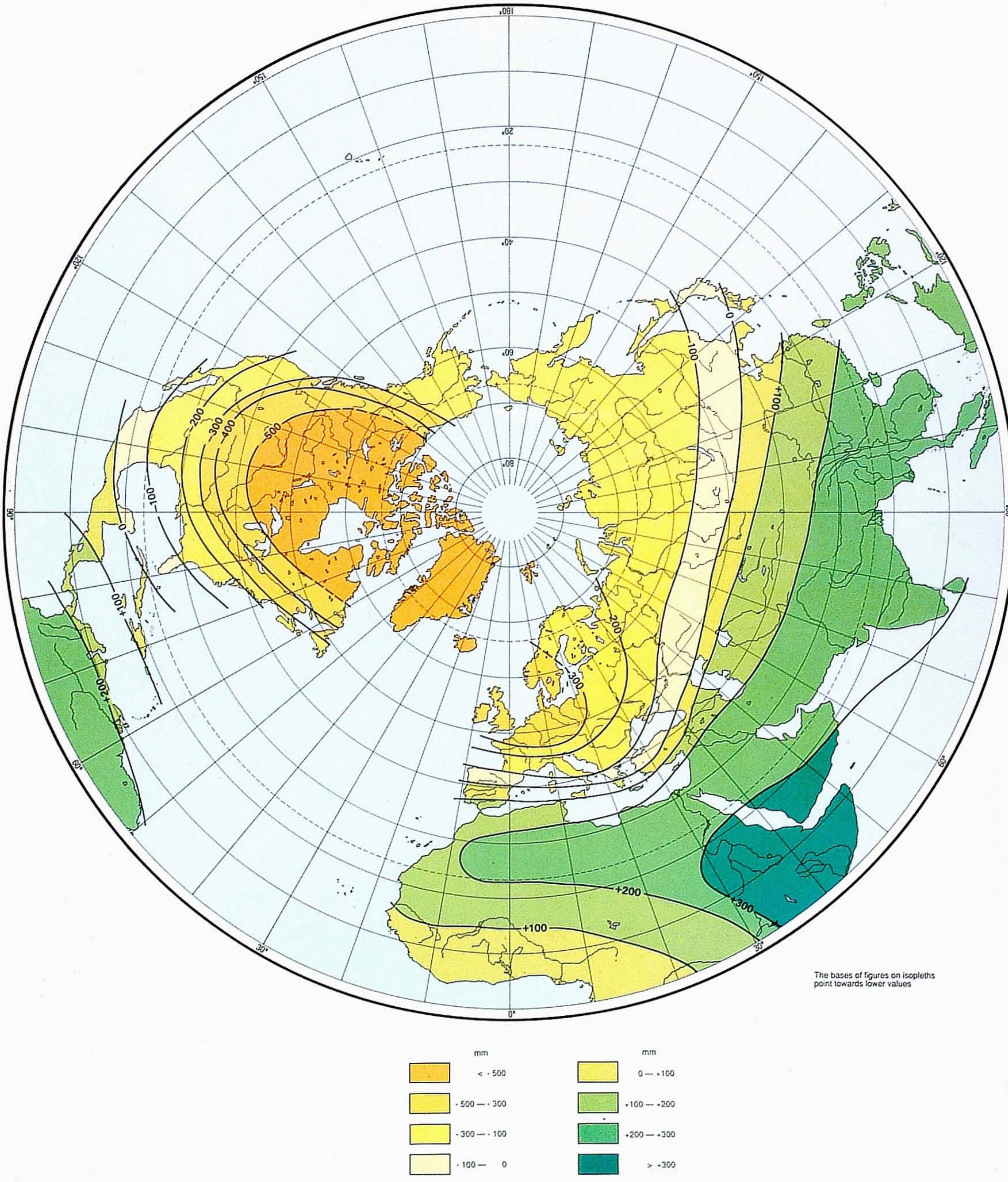
INTERSTADIAL OF THE LAST GLACIATION

(about 35,000 to 25,000 yr B.P.)

ANNUAL PRECIPITATION

Minimal deviations from present-day values

by B. Frenzel



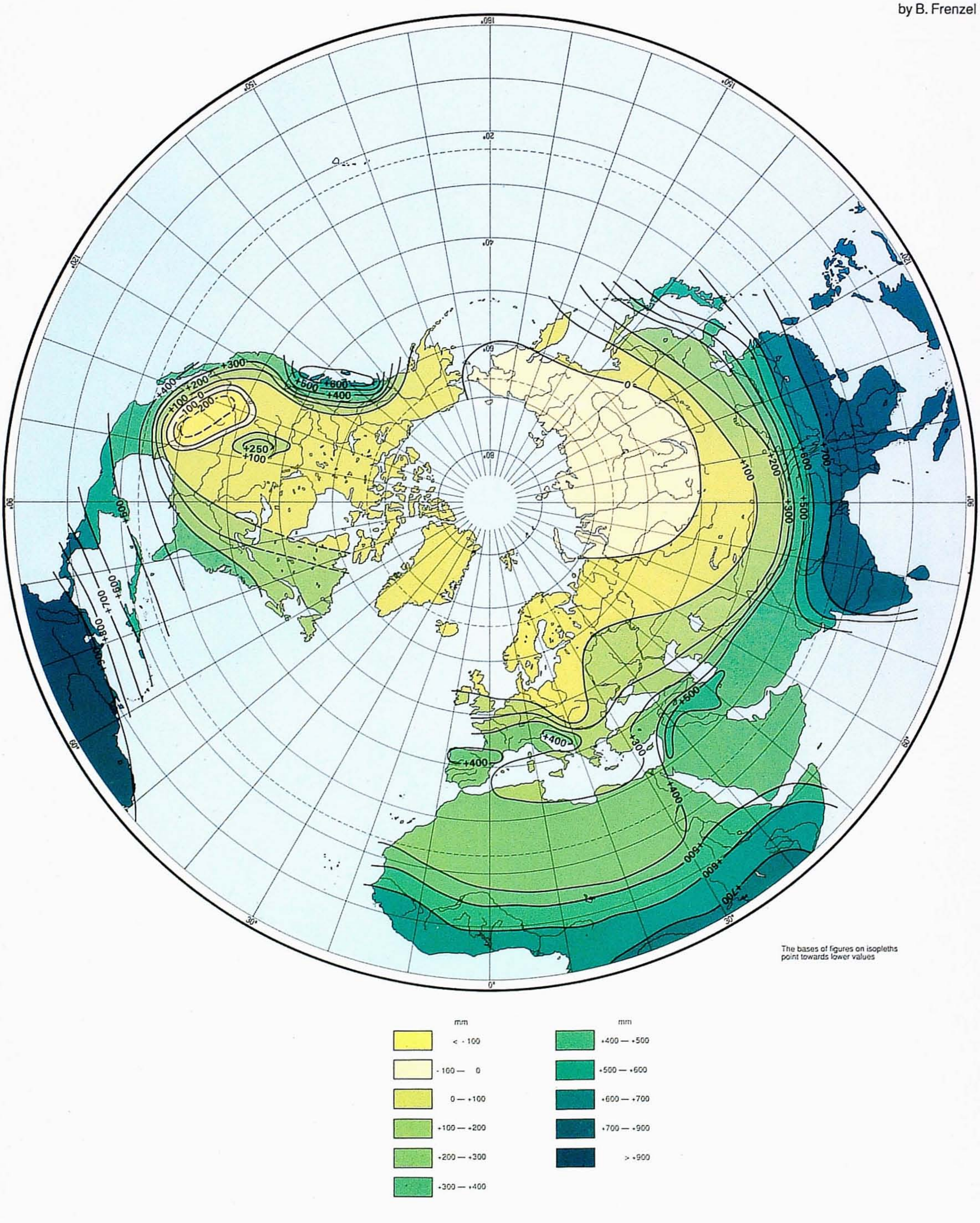
INTERSTADIAL OF THE LAST GLACIATION

(about 35,000 to 25,000 yr B.P.)

ANNUAL PRECIPITATION

Deviations from the pleniglacial values (20,000 to 18,000 yr B.P.)

by B. Frenzel



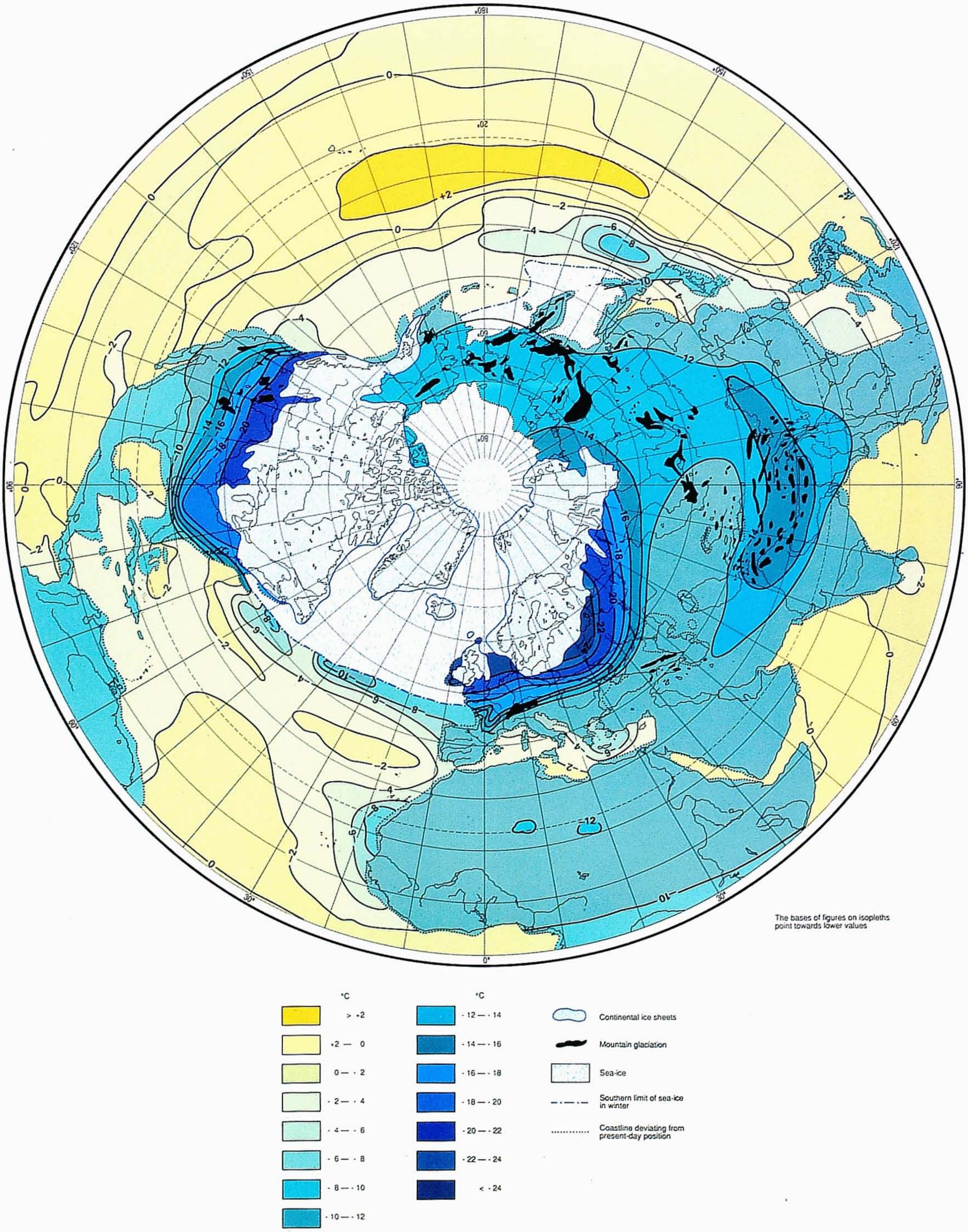
MAXIMUM COOLING OF THE LAST GLACIATION

(about 20,000 to 18,000 yr B.P.)

FEBRUARY MEAN TEMPERATURE

Minimal deviations from present-day values

by B. Frenzel



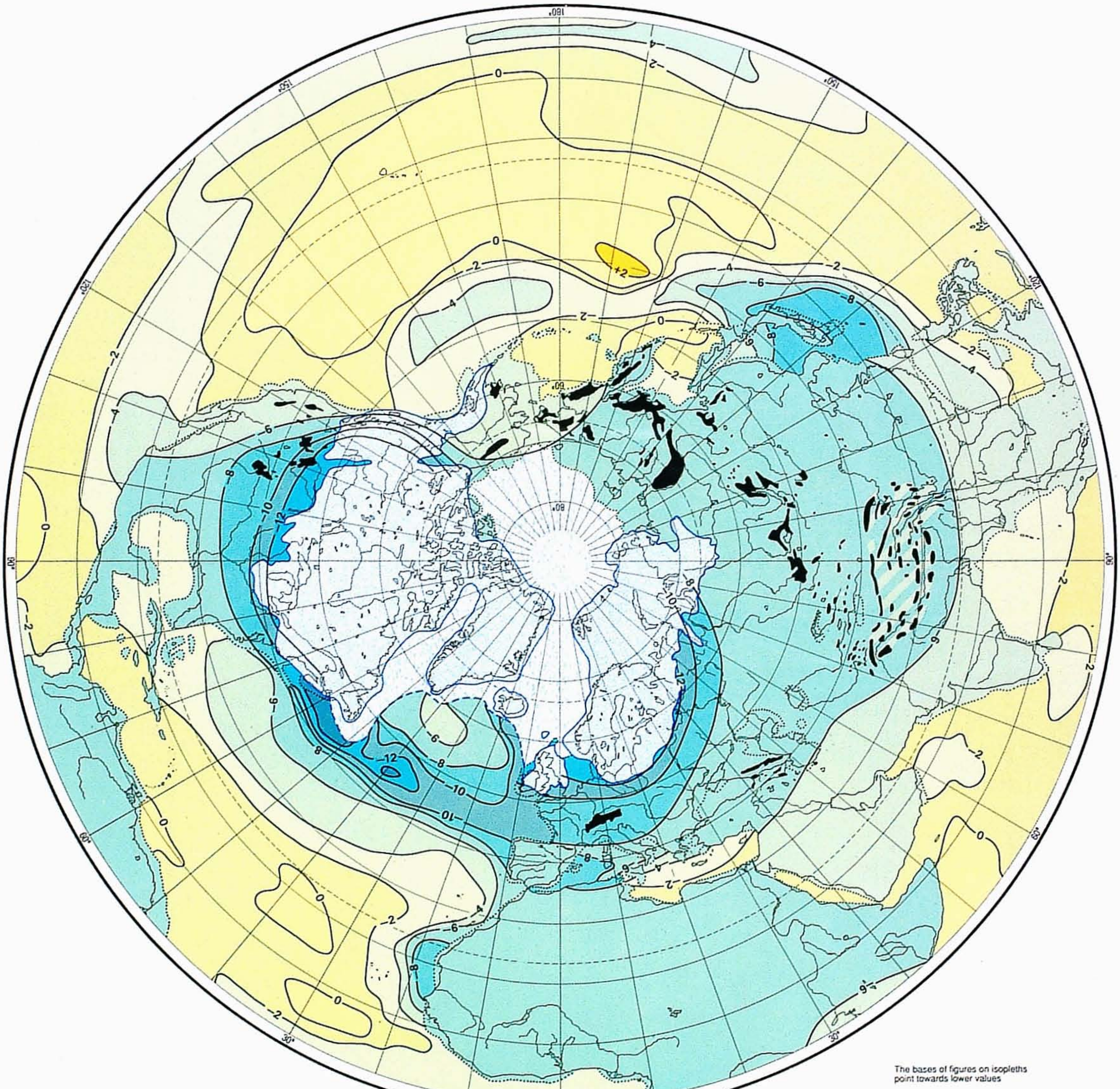
MAXIMUM COOLING OF THE LAST GLACIATION

(about 20,000 to 18,000 yr B.P.)

AUGUST MEAN TEMPERATURE

Minimal deviations from present-day values

by B. Frenzel



The bases of figures on isopleths point towards lower values

°C		°C			
(colour)	> +2	(colour)	- 6 — - 8	(symbol)	Continental ice sheets
(colour)	+2 — 0	(colour)	- 8 — - 10	(symbol)	Mountain glaciation
(colour)	0 — - 2	(colour)	- 10 — - 12	(symbol)	Sea-ice
(colour)	- 2 — - 4	(colour)	- 12 — - 14	- - - - -	Southern limit of sea-ice in summer
(colour)	- 4 — - 6	(colour)	< - 14	....	Coastline deviating from present-day position

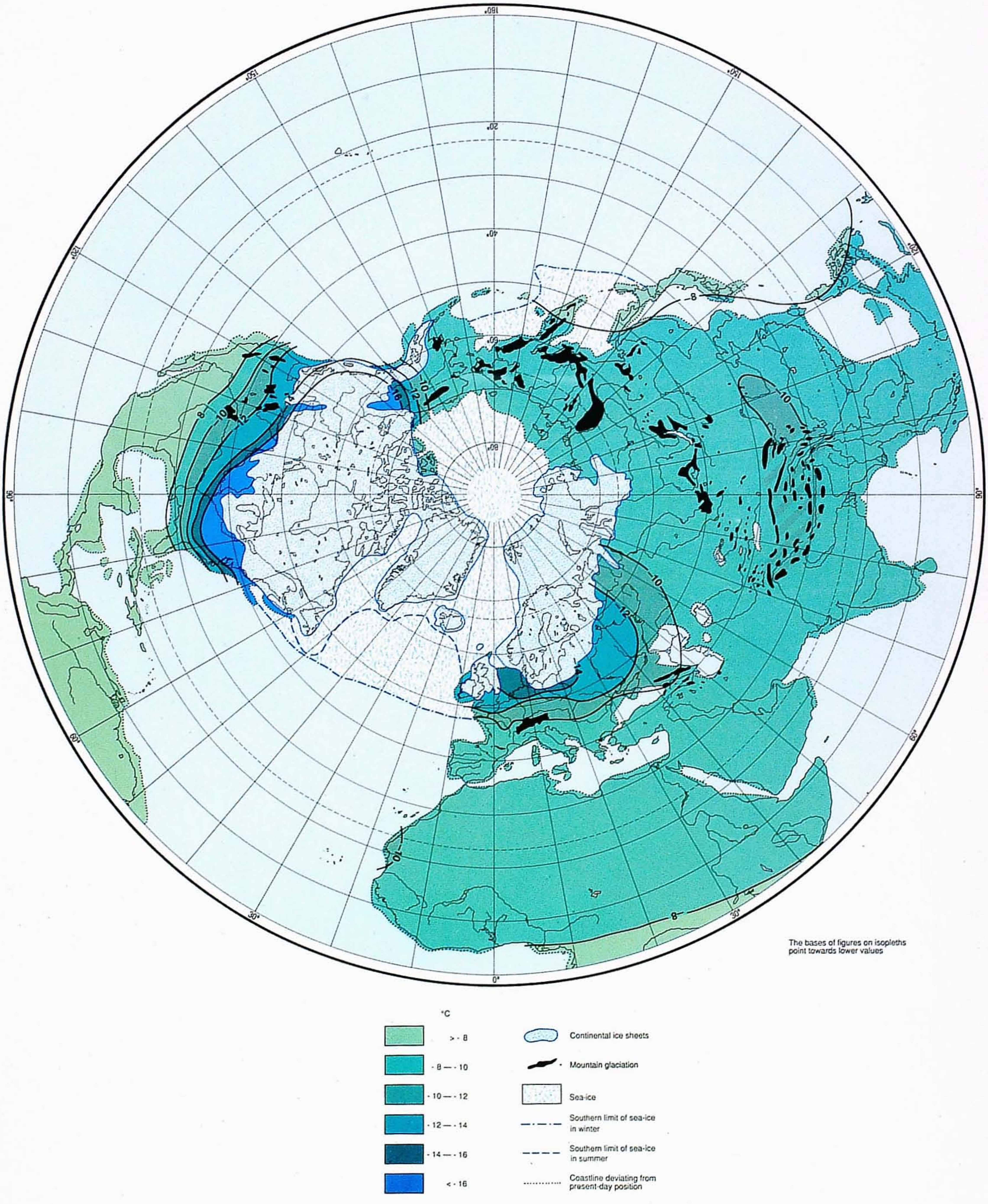
MAXIMUM COOLING OF THE LAST GLACIATION

(about 20,000 to 18,000 yr B.P.)

ANNUAL MEAN TEMPERATURE

Minimal deviations from present-day values

by B. Frenzel



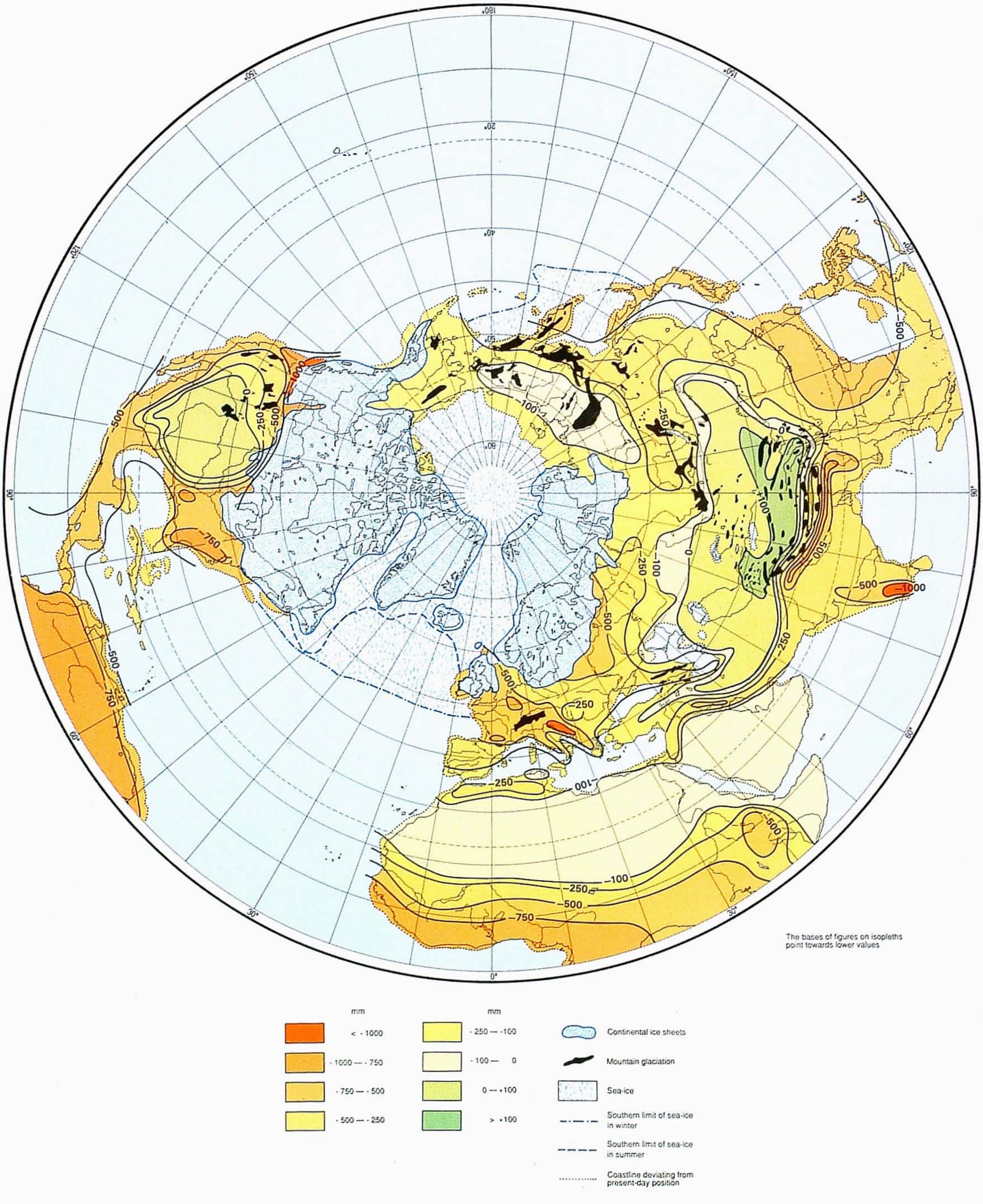
MAXIMUM COOLING OF THE LAST GLACIATION

(about 20,000 to 18,000 yr B.P.)

ANNUAL PRECIPITATION

Minimal deviations from present-day values

by B. Frenzel



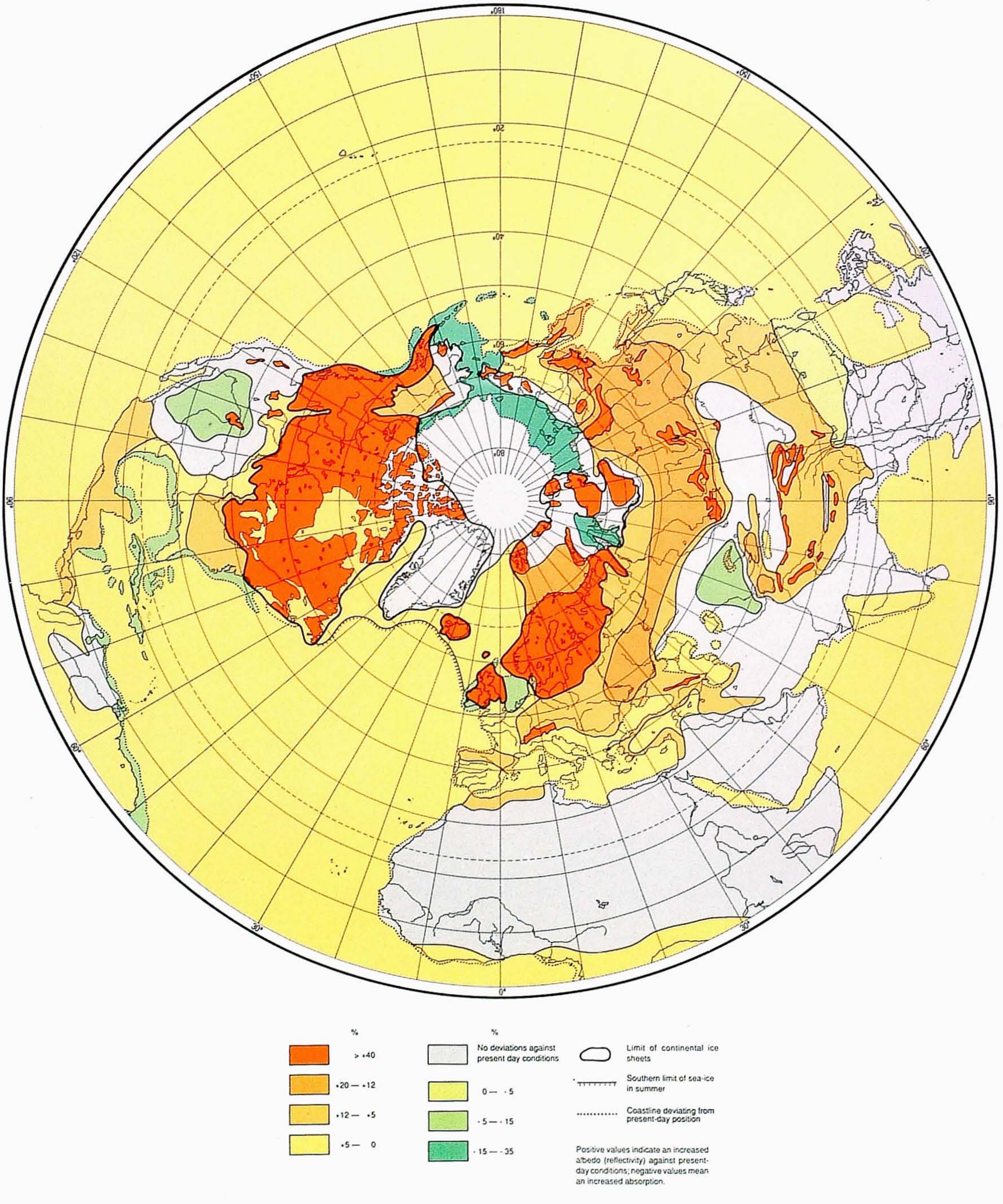
MAXIMUM COOLING OF THE
LAST GLACIATION

(about 20,000 to 18,000 yr B.P.)

SURFACE ALBEDO FOR THE SUMMER

Minimal deviations from present-day values

by B. Frenzel

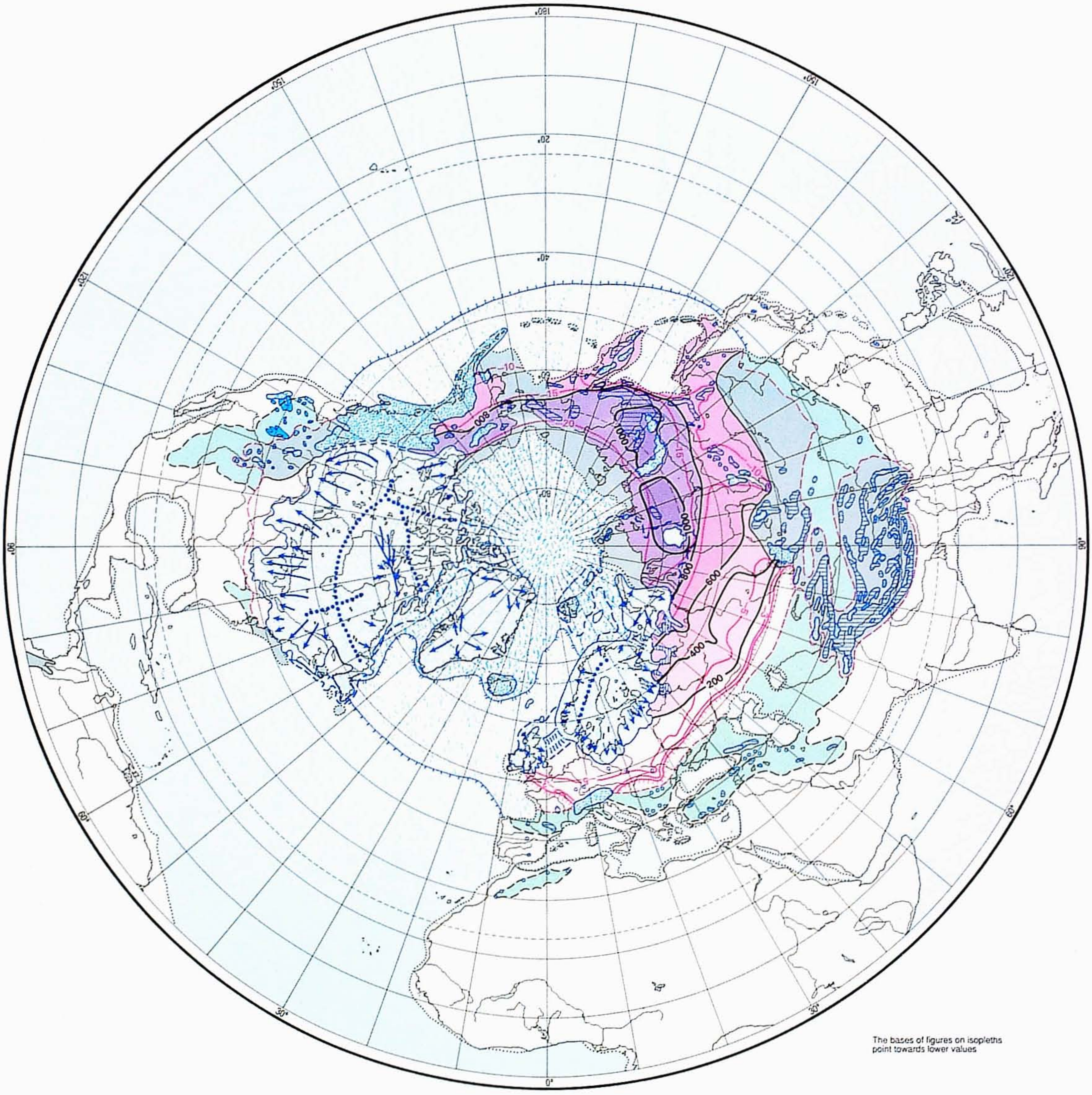


MAXIMUM COOLING OF THE LAST GLACIATION

(about 20,000 to 18,000 yr B.P.)

GLACIATION AND PERMAFROST

GLACIATION by O. Conchon, R. Fulton, M.A. Faustova L.L. Isayeva, J.E. Mojski, S.C. Porter, I.I. Spasskaya, S.N. Timireva, A.A. Velichko, W. Zagwijn
PERMAFROST by V.V. Baulin, N.S. Danilova, V.P. Nechayev, T.L. Péwé, A.A. Velichko



- Continental ice-sheets (another version see page 55)
- Mountain glacial complexes
- Areas of limited mountain glaciers
- Local mountain glaciers
- Ice sheet limit: defined
- Boundary between interacting ice sheets
- Assumed limits of ice extent („maximum"concept)
- Ice divides
- Ice lobes
- Ice flow direction
- Zones of interacting ice sheets (heterochronous advances of different ice sheets)
- Perennial sea-ice
- Seasonal sea-ice
- Perennial sea-ice limit
- Seasonal sea-ice limit
- Proglacial basins, probably short-lived lakes
- Pluvial lakes

Thickness of permafrost (m)

- > 1000
- 1000 — 800
- 800 — 600
- 600 — 400
- 400 — 200
- < 200
- Thickness unknown

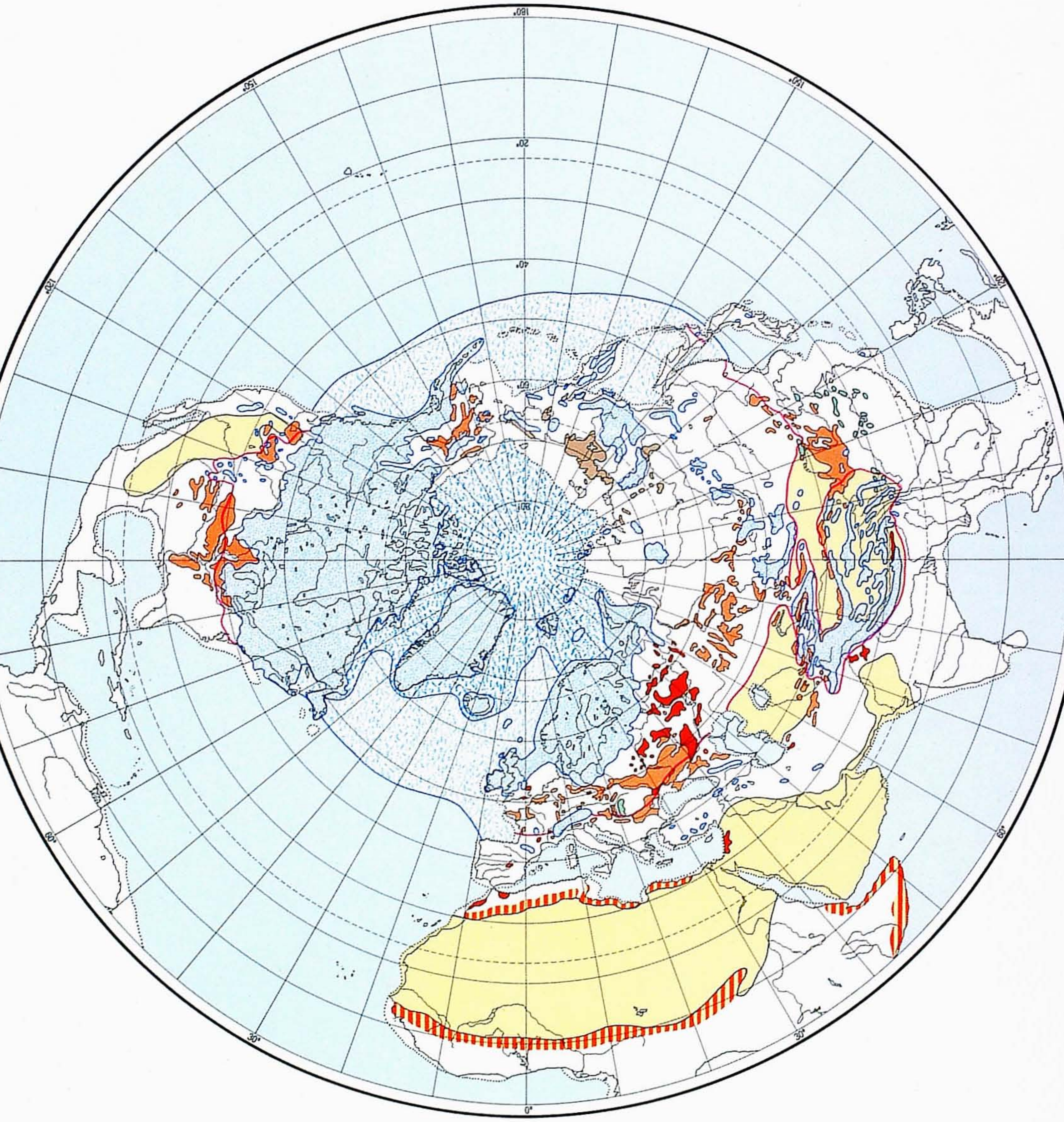
- Isolines of permafrost thickness
- Areas of discontinuous and sporadic permafrost
- Boundary of discontinuous and sporadic permafrost
- Areas of deep seasonal freezing
- Boundary of deep seasonal freezing
- -3 Geoisotherms of permafrost (°C)
- Boundary of continuous permafrost
- No traces of deep seasonal freezing found
- Coastline deviating from present-day position

MAXIMUM COOLING OF THE LAST GLACIATION

(about 20,000 to 18,000 yr B.P.)

LOESS

by M. Pécsi (USSR territory by A.A. Velichko and T.A. Khalcheva)



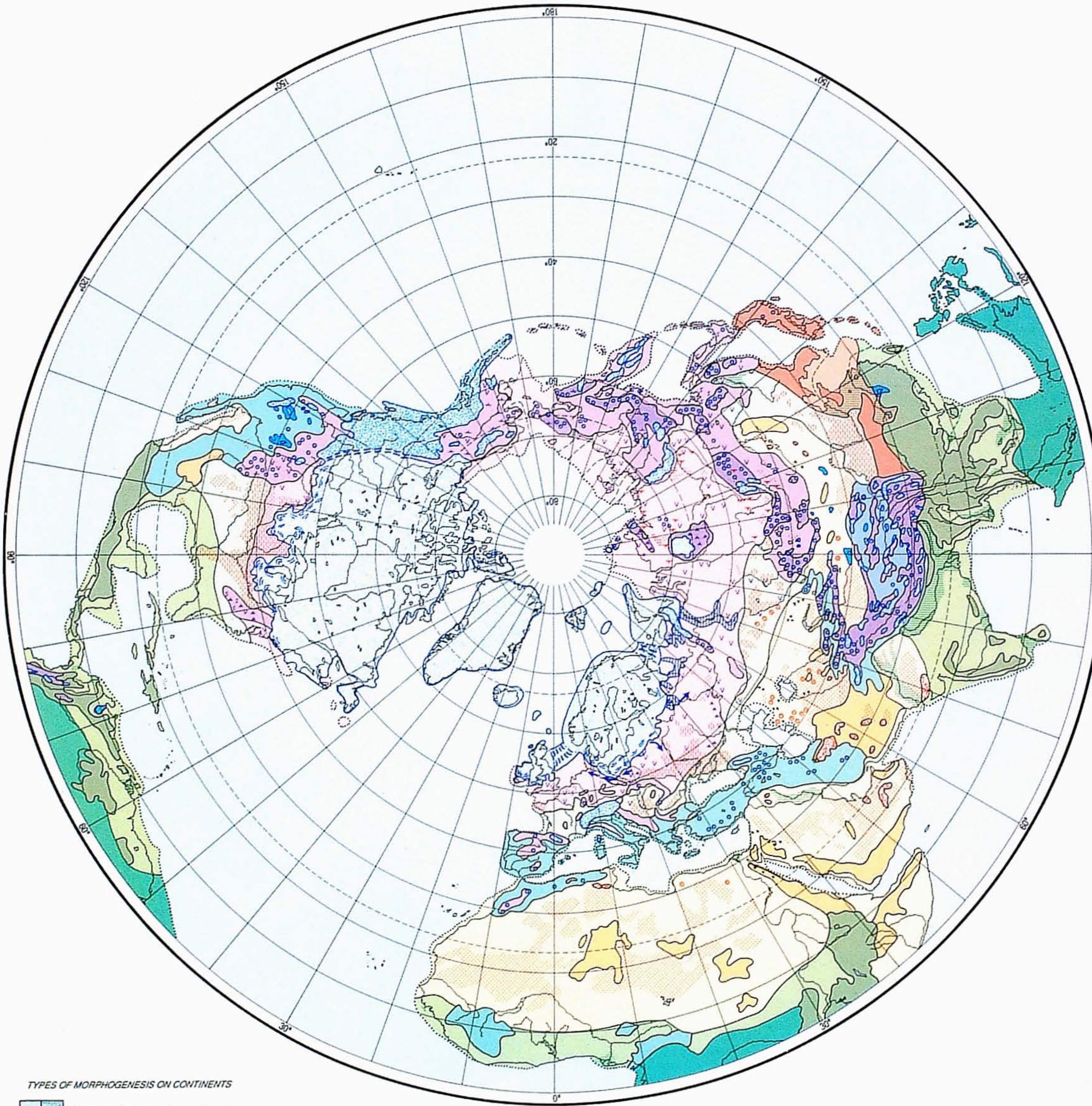
- Typical loess, sandy loess
- Loess derivates, continuous and sporadic distribution
- Ice-loess complex (yedoma)
- Flood-plain loess
- Loess-like formation, continuous and sporadic distribution
- Desert region (sandy, stony desert and semidesert)
- Glaciated area: ice sheets and mountain glaciation
- Perennial sea-ice
- Seasonal sea-ice
- Boundary of continuous permafrost
- Coastline deviating from present-day position

MAXIMUM COOLING OF THE LAST GLACIATION

(about 20,000 to 18,000 yr B.P.)

DOMINANT GEOMORPHIC PROCESSES

by I.I. Spasskaya



TYPES OF MORPHOGENESIS ON CONTINENTS

- a | b — Glacial (a- ice sheets; b- mountains)
- Cold arid (high mountain desert)
- Nivational (high mountains)
- a | b — Periglacial (a- plains; b- mountains)
- Cold temperate
- a | b — Mild temperate (a- plains; b- mountains)
- a | b — Arid (a- plains; b- mountains)
- a | b — Tropical and subtropical, semiarid and seasonally wet (a- plains; b- mountains)
- Mountain unspecified (semiarid, semihumid, humid)
- Equatorial (permanently wet)

M O R P H O G E N E T I C I N D I C A T O R S

of glacial processes:

- Small areas of cirque and valley glaciers (out of scale)
- Morainic ridges
- Proglacial basins, probably short-lived lakes
- Direction of meltwater drainage

of cryogenous (periglacial) processes:

- Ice wedges
- Solifluction features
- Altiplanation terraces
- Block fields, block streams

of arid processes:

- Area of wind blown sands
- Deflation hollows
- Deflation pavements
- Ventifacts
- Area of loess accumulation
- Solonchaks
- Features of physical weathering (on solid rocks)

OTHER SYMBOLS

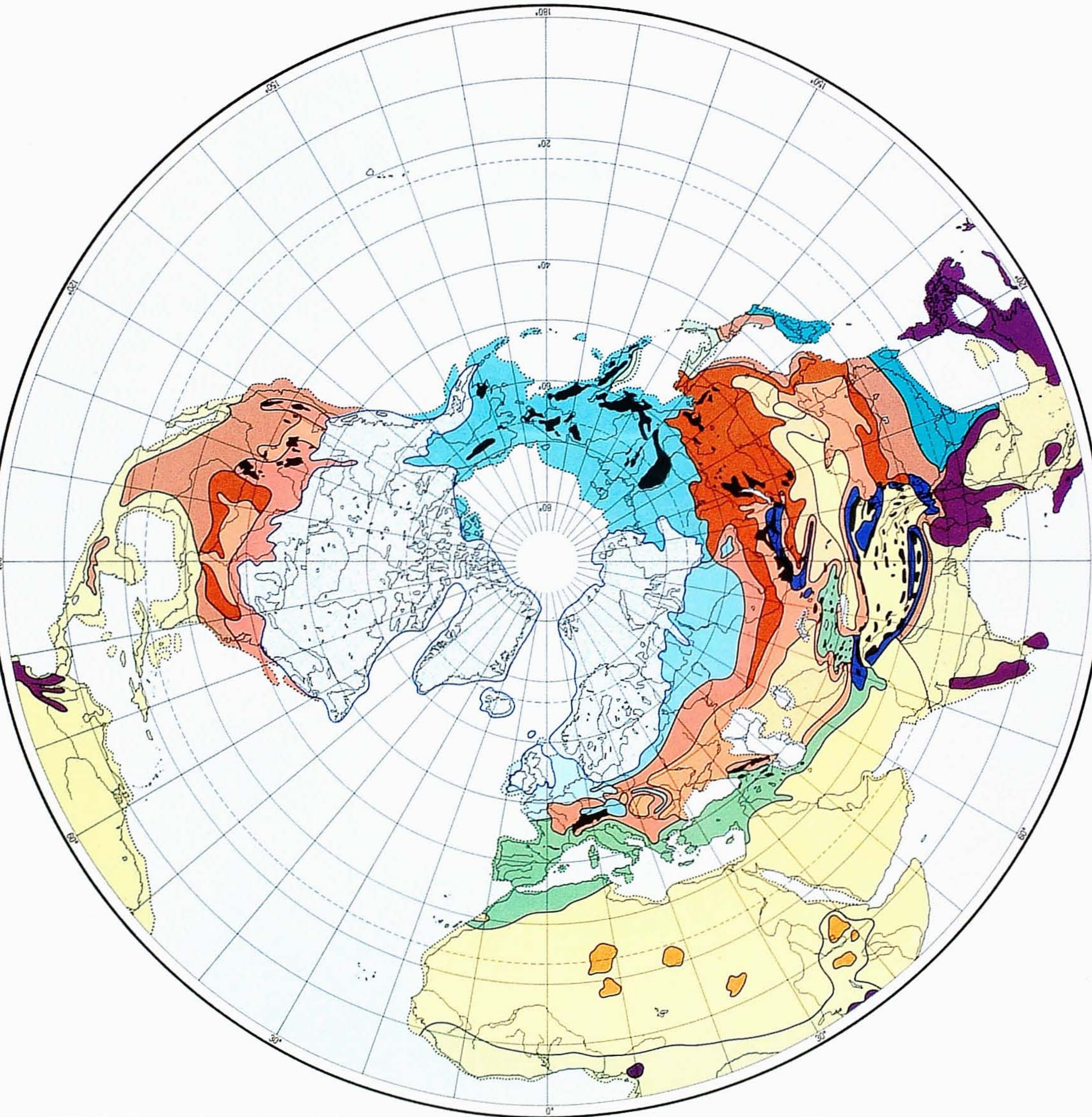
- Coastline deviating from present-day position
- Ice sheet limits: defined
- Ice sheet limits: presumed
- Zones of heterochronous advances of different ice sheets
- River valleys on the shelf (documented)
- Fluvial plains
- Pluvial lakes

MAXIMUM COOLING OF THE
LAST GLACIATION

(about 20,000 to 18,000 yr B.P.)

VEGETATION

by B. Frenzel



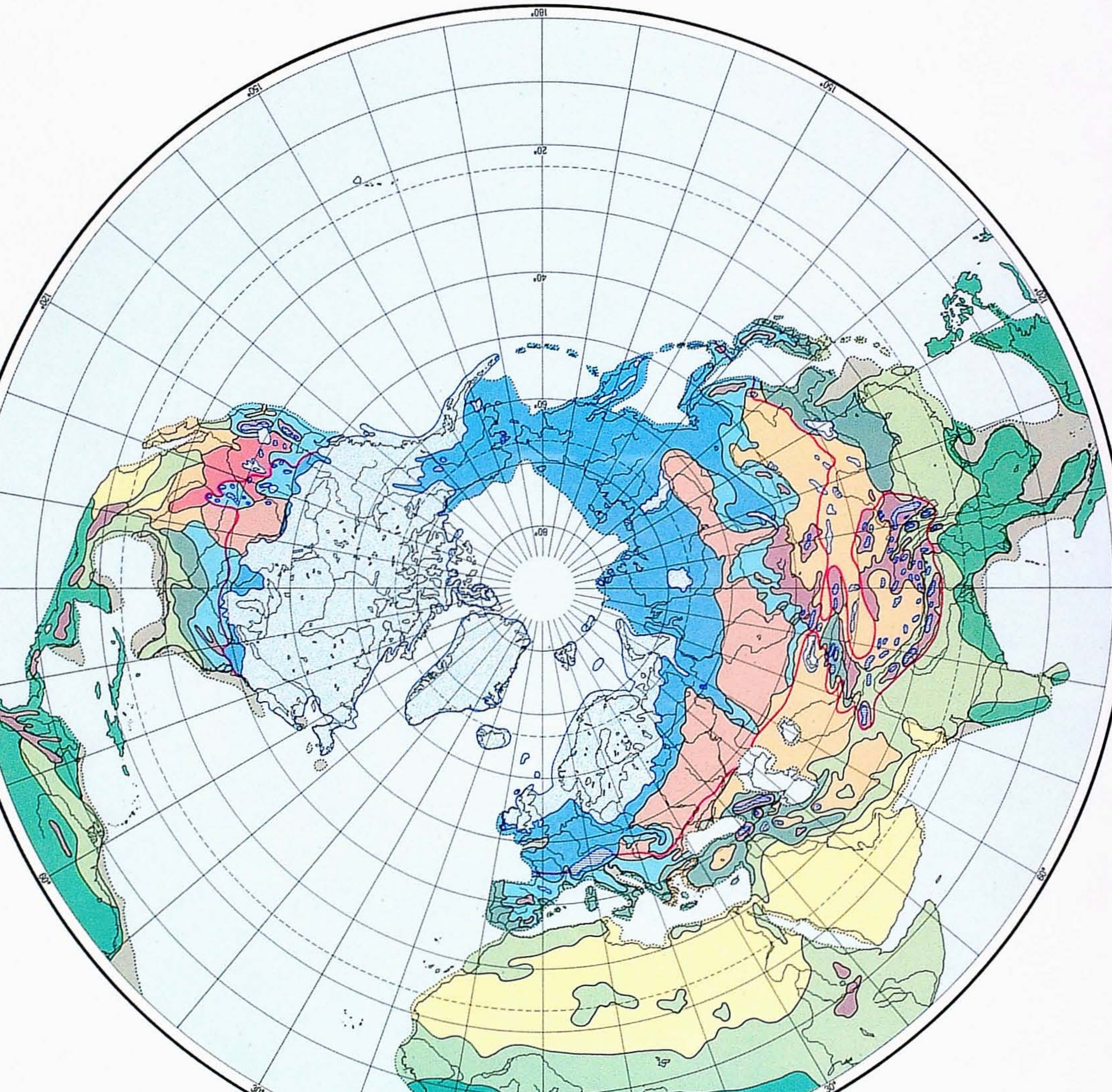
- Tundra-like plant communities of oceanic and suboceanic regions, to some extent with a lot of Rumex and dwarf-willow species, Gramineae, Cyperaceae and Selaginella. A very sparse plant-cover
- Mixture of steppe and tundra communities: tundra species in moist habitats; Artemisia, Chenopodiaceae, a wealth of herbs on dry places. A very sparse plant-cover
- Tundra-steppe, rich in herbs, Artemisia, Gramineae, Selaginella, and various lycopods. A very sparse plant-cover
- Steppe consisting most of all of Artemisia, herbs and Chenopodiaceae, together with Ephedra. Formation of loess and loess-loam up to approximately 400 to 600 m a.s.l.; a sparse vegetation
- Loess-steppe, being dominated by Chenopodiaceae, various herbs and Artemisia species. Facultative and obligate halophytes. An open vegetation
- Herb-steppe with various Chenopodiaceae species. Loess formation. An open vegetation
- Steppe-communities, in the northern parts of Siberia a mixture of steppe and tundra communities with some isolated stands of cold- and drought-resistant coniferous and deciduous tree-species. More or less continuous plant-cover
- Semideserts and deserts, rich in halophytes, chenopods, and sage-brush. In the surroundings of mountainous areas increasing amount of xerophilous herbs and grasses. Unprotected soil surface prevailed
- Xerotic vegetation of the mountainous areas of the Central Sahara, together with some stands of tree species in moister habitats. A mosaic of various, in general very sparse, plant communities
- Mediterranean to sub-mediterranean Artemisia-Ephedra-Juniperus-Compositae steppe of mountainous regions, most of all in their southern parts containing some small stands and woods of forest-communities rich in ecologically different tree-species
- Herb-steppes of the Central Asiatic high mountains, with some stands of Myricaria, Juniperus and Ephedra. A very sparse plant-cover
- Dry savanna
- Subalpine and subboreal open coniferous and in general very frost resistant deciduous forests, containing large steppe areas. Vegetation often influenced by eolian sedimentation. Dense plant-cover
- Birch-alder forest-tundra to forest-steppe
- Hygrophytic mixed coniferous and broad-leaved forest of Japan and South China. Dense plant-cover
- Subtropical to tropical mountain forests
- Continental ice-sheets (another version see page 49)
- Mountain glaciation
- Coastline deviating from present-day position

MAXIMUM COOLING OF THE LAST GLACIATION

(about 20,000 to 18,000 yr B.P.)

MAIN TYPES OF VEGETATION (ECOSYSTEMS)

by V.P. Grichuk



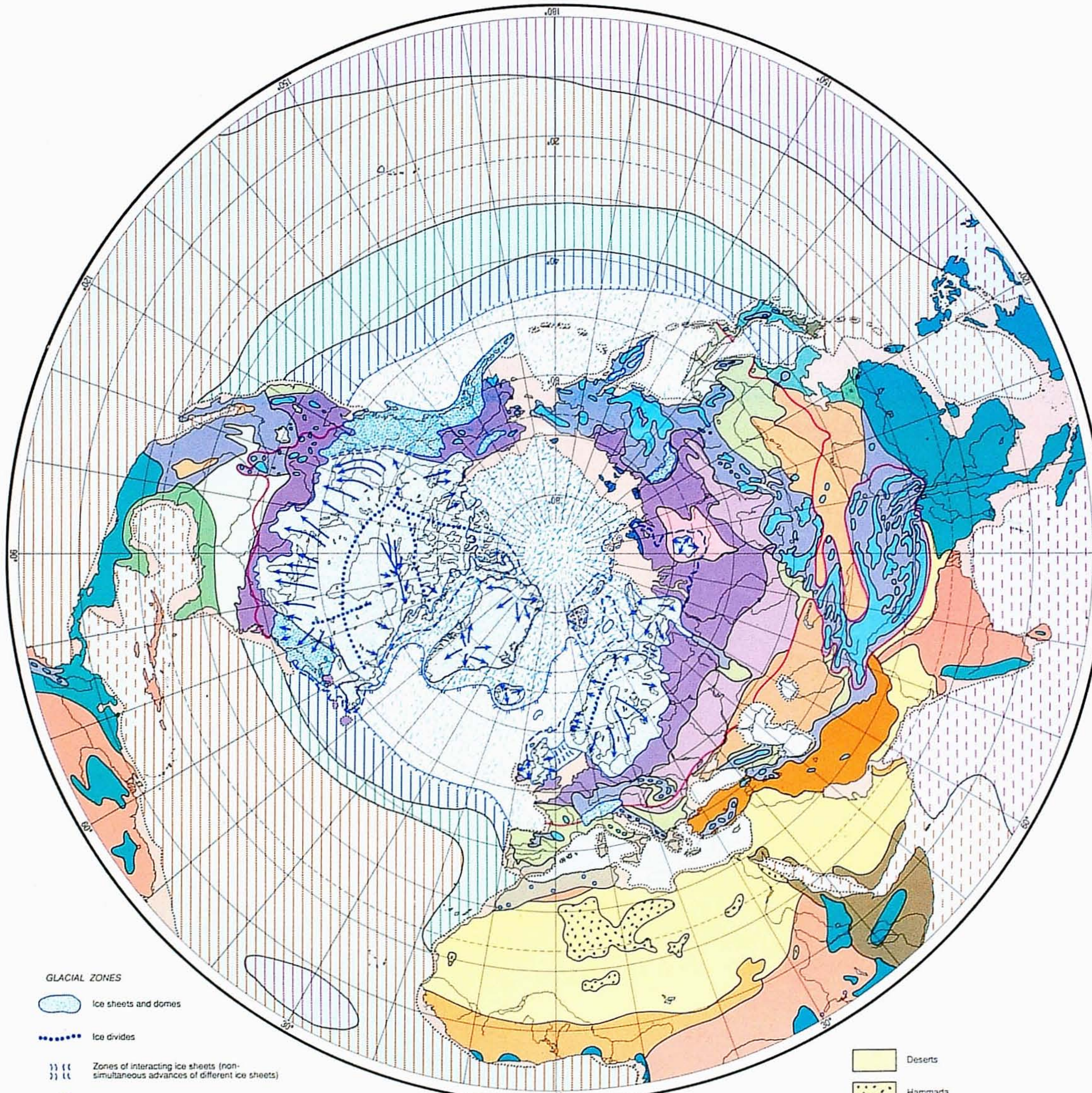
- Periglacial tundra
- Periglacial steppe
- Periglacial open woodlands
- Boreal type of vegetation
- Nemoral type of vegetation
- Steppe and desert
- Subtropical savanna vegetation
- Tropical forest
- Tropical steppe and desert
- High mountain vegetation
- Unknown type of vegetation on emerged shelves
- Continental ice-sheets (another version see page 55)
- Areas of mountain glaciation
- Southern limit of permafrost
- Lakes
- Coastline deviating from present-day position

MAXIMUM COOLING OF THE LAST GLACIATION

(about 20,000 to 18,000 yr B.P.)

LANDSCAPE TYPES

by A.A. Velichko, L.L. Isayeva



GLACIAL ZONES

- Ice sheets and domes
- Ice divides
- Zones of interacting ice sheets (non-simultaneous advances of different ice sheets)
- Main ice flow lines
- Individual small ice domes
- Assumed limits of ice extent („maximum" concept)
- Perennial sea-ice
- Limit of perennial sea-ice
- Seasonal sea-ice
- Limit of seasonal sea-ice
- Mountain glacial complexes
- Areas of limited mountain glaciers
- Local mountain glaciers

CRYOBOREAL AND BOREAL ZONES

- Arctic desert
- Tundra-steppes
- Periglacial steppes
- Open woodlands
- Steppes and semideserts
- Deserts, locally semideserts
- Stone deserts, tundras and alpine meadows in high mountains associated with steppes and forests in low mountains and foothills
- Open forests (including dark coniferous and broad-leaved refugia) on mountain slopes
- Mountain steppes, locally with trees

- Limit of continuous permafrost

WARM TEMPERATE, SUBTROPICAL, TROPICAL AND EQUATORIAL ZONES

- Forests, mostly broad-leaved
- Mixed forests (broad-leaved and coniferous species)
- Mountain forests and bushes
- Open mountain forests, with broad-leaved species
- Savannas
- Dry savannas
- Deserts
- Hammada
- Evergreen and deciduous forests more open than their modern equivalents

CLIMATIC ZONES IN OCEANS

- Polar and subpolar
- Temperate
- Subtropical (a= proved, b= suggested)
- Tropical and equatorial (a = proved, b = suggested)

SPECIAL SYMBOLS

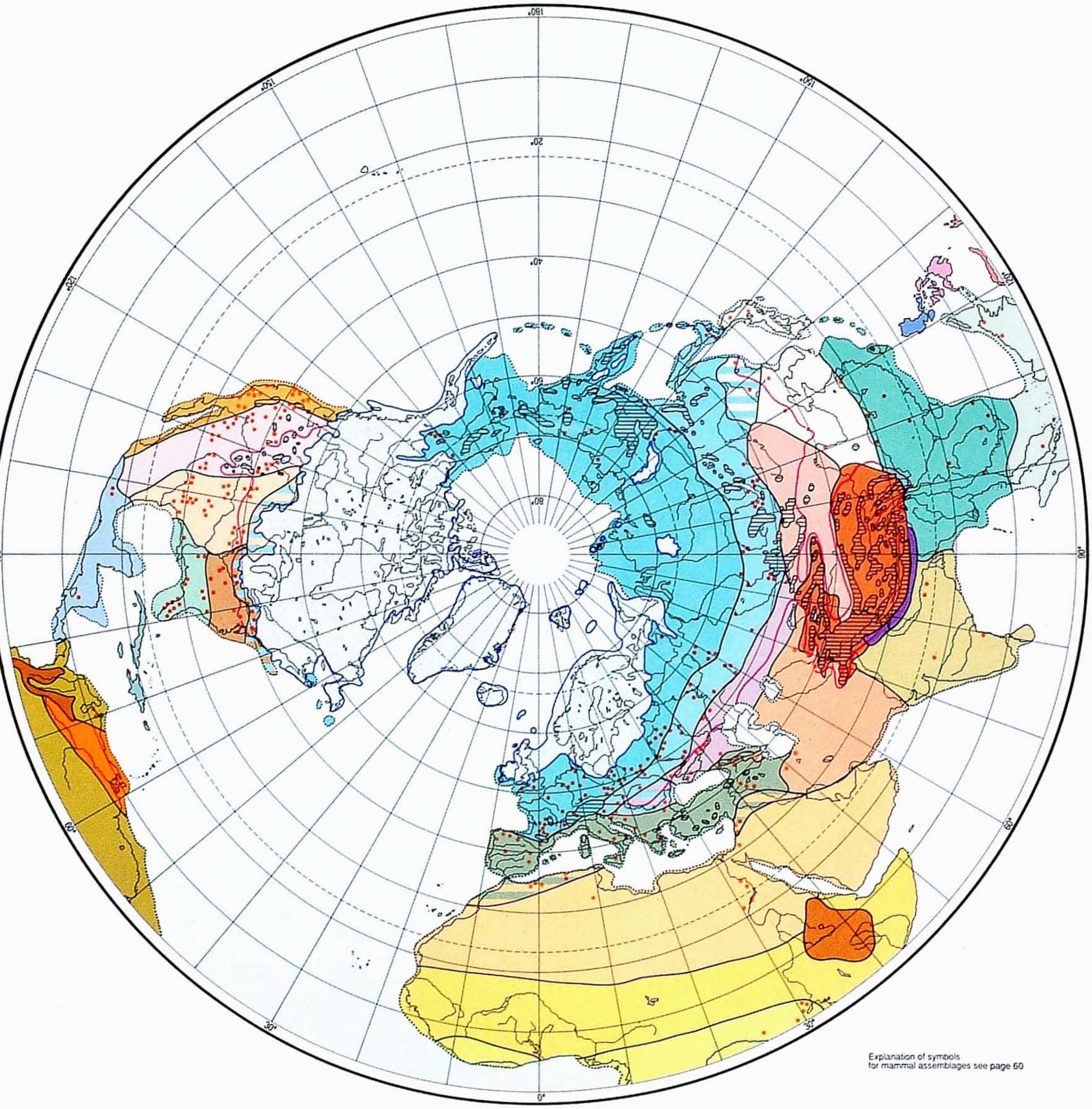
- Coastline deviating from present-day position
- Emerged shelf
- Inland sea and lakes

UPPER PLENIGLACIAL OF THE LAST GLACIATION

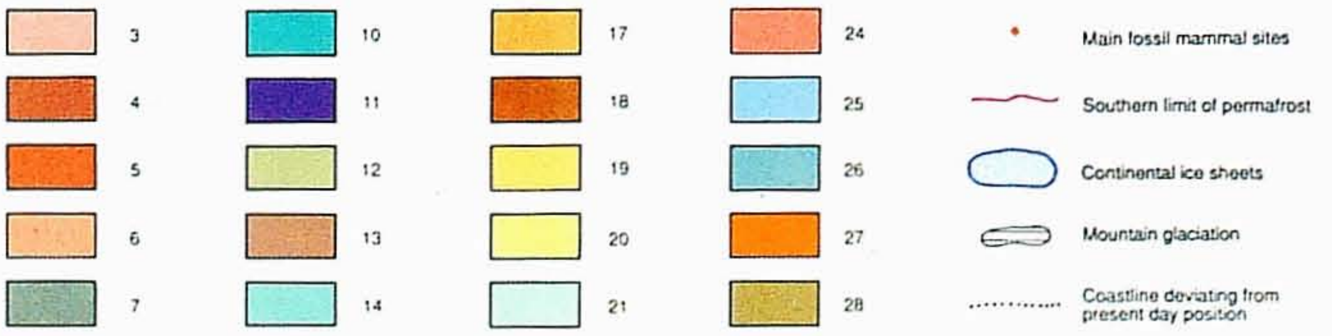
(about 24,000 to 12,000 yr B.P.)

MAIN MAMMAL ASSEMBLAGES

by G.F. Baryshnikov, A.K. Markova



Explanation of symbols
for mammal assemblages see page 60

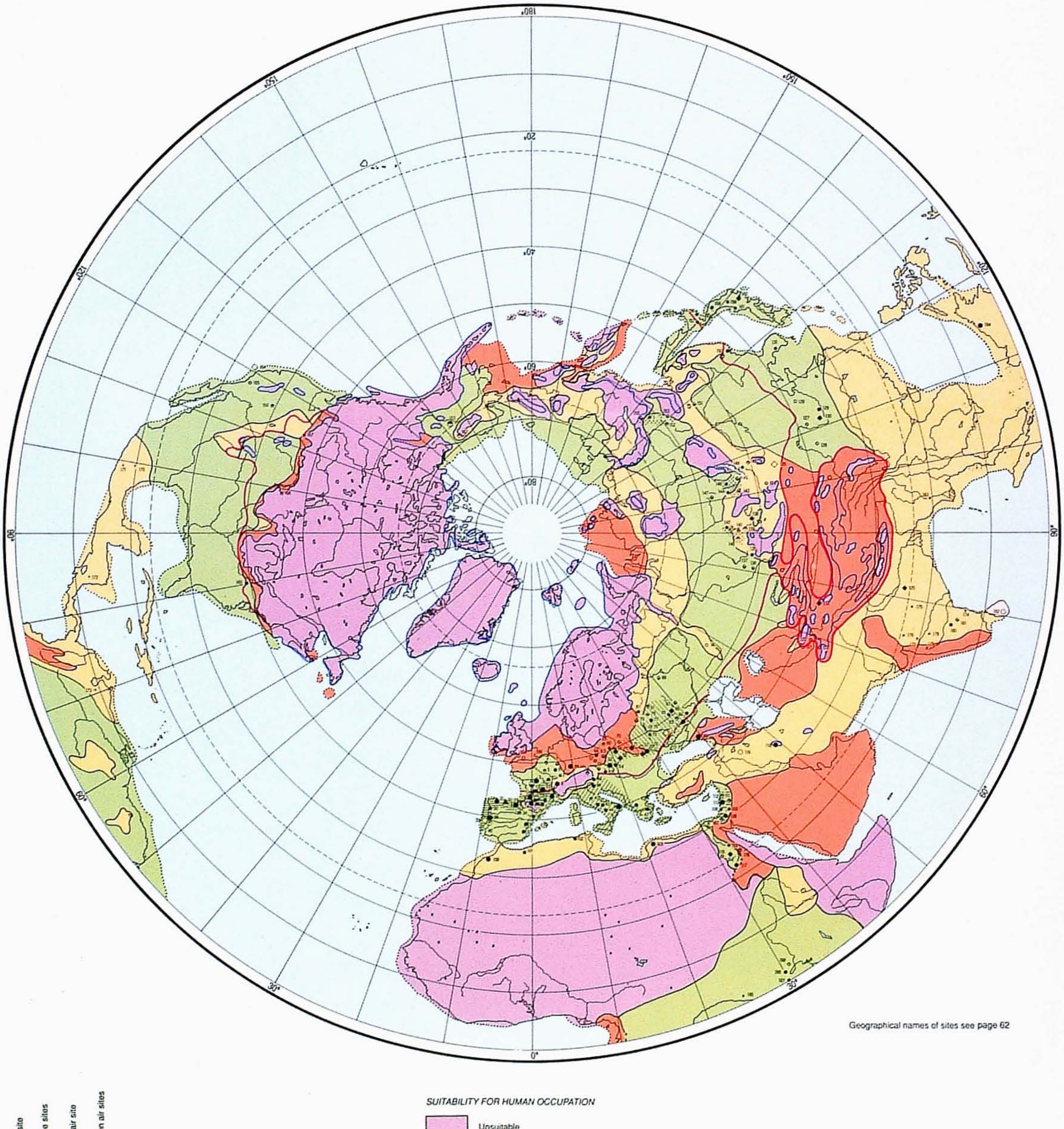


UPPER PLENIGLACIAL OF THE LAST GLACIATION

(about 24,000 to 15,000 yr B.P.)

HUMAN OCCUPATION

by T. Madeyska, O. Soffer, E. I. Kurenkova



Geographical names of sites see page 62

Single cave site	Group of cave sites	Single open air site	Group of open air sites	
•	•	•	•	Sites dated to 15 000 — 16 000 years BP.
•	●	•	◆	Sites dated to 22 000 — 16 000 years BP.
•	•	•	•	Sites dated to 24 000 — 22 000 years BP.
▫	○	∘	◇	Sites without precise dating

★ Upper paleolithic sites, tentatively attributed to the Late Pleistocene

+ Pre-Clovis sites in America

SUITABILITY FOR HUMAN OCCUPATION

- Unsuitable
- Inhospitable
- Limited suitability
- Suitable

MAIN UPPPER PALEOLITHIC CULTURAL UNITS CHARACTERISTIC OF THE LAST GLACIATION

- Protosolutrean, Solutrean, Solutrean-Levantanian Group
- Tardigravettian
- Early Magdalenian (Badegoulian and related cultures)
- Central and East European late backed points cultures („East Gravettian", Ságvárian, Molodovian, Mezinian, Eliseevichian, Kostienkian, etc.)
- Siberian cultures (Diuktai, Malta-Buret, Afontova, Kokorevo)
- Kebaran culture

- Boundary of ice sheets
- Boundary of permafrost
- Coastline deviating from present-day position

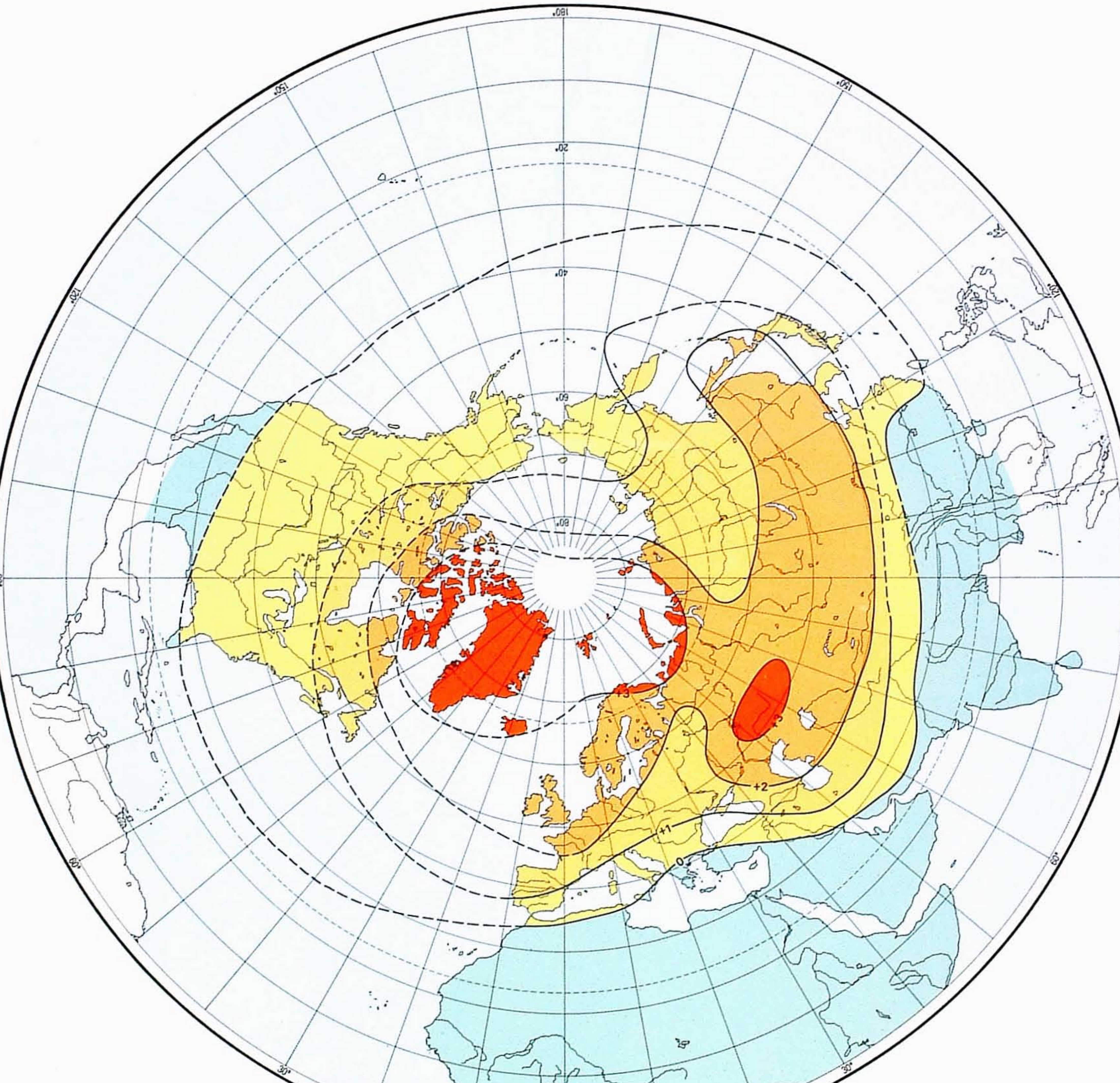
HOLOCENE

(about 6,000 to 5,500 yr B.P.)

JANUARY MEAN TEMPERATURE

Deviations from present-day values

by V.A. Klimanov, A.A. Velichko



°C

	> +3
	+3 — +2
	+2 — +1
	+1 — 0
	< 0

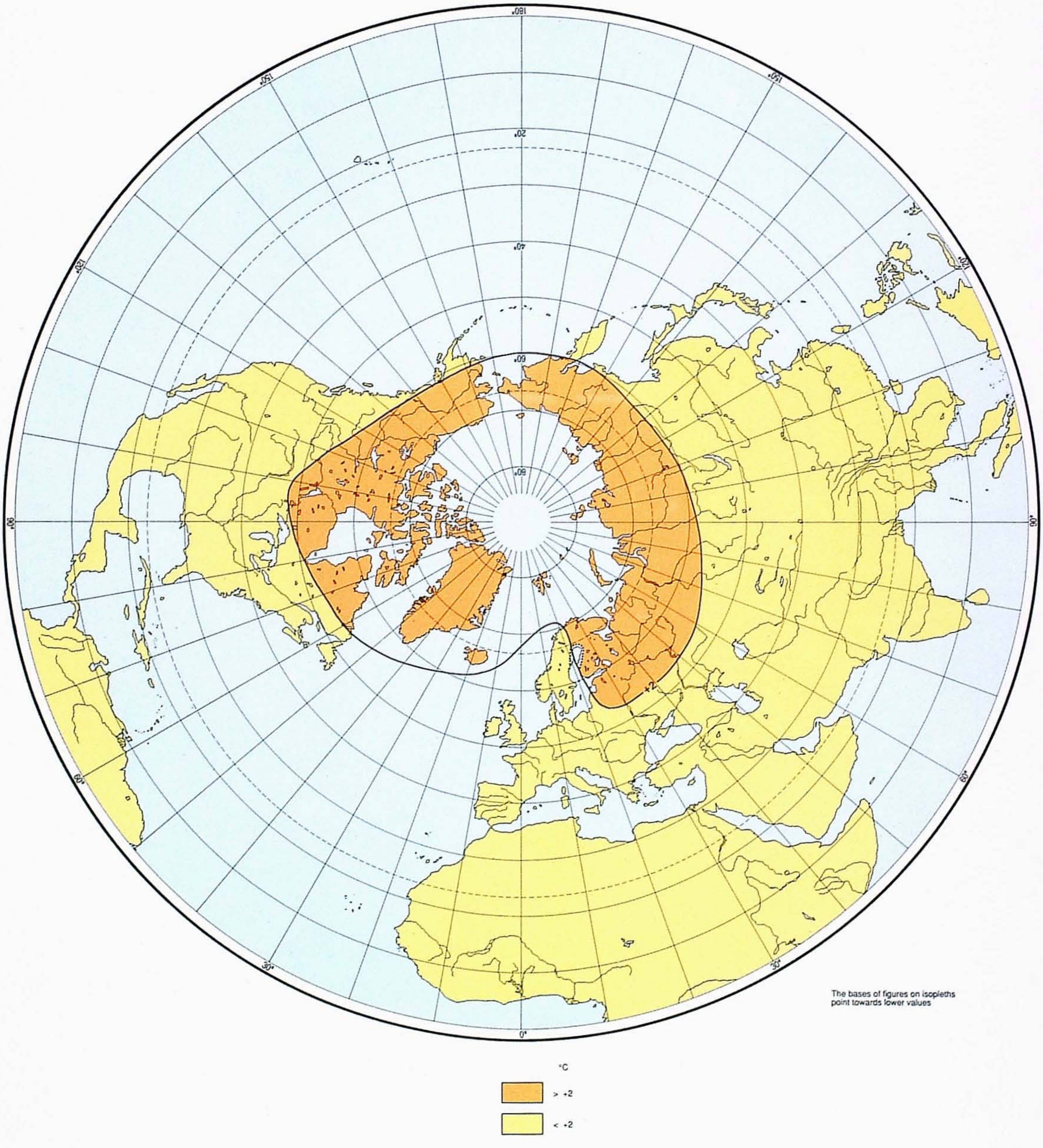
HOLOCENE

(about 7,000 to 6,500 yr B.P.)

FEBRUARY MEAN TEMPERATURE

Minimal deviations from present-day values

by B. Frenzel



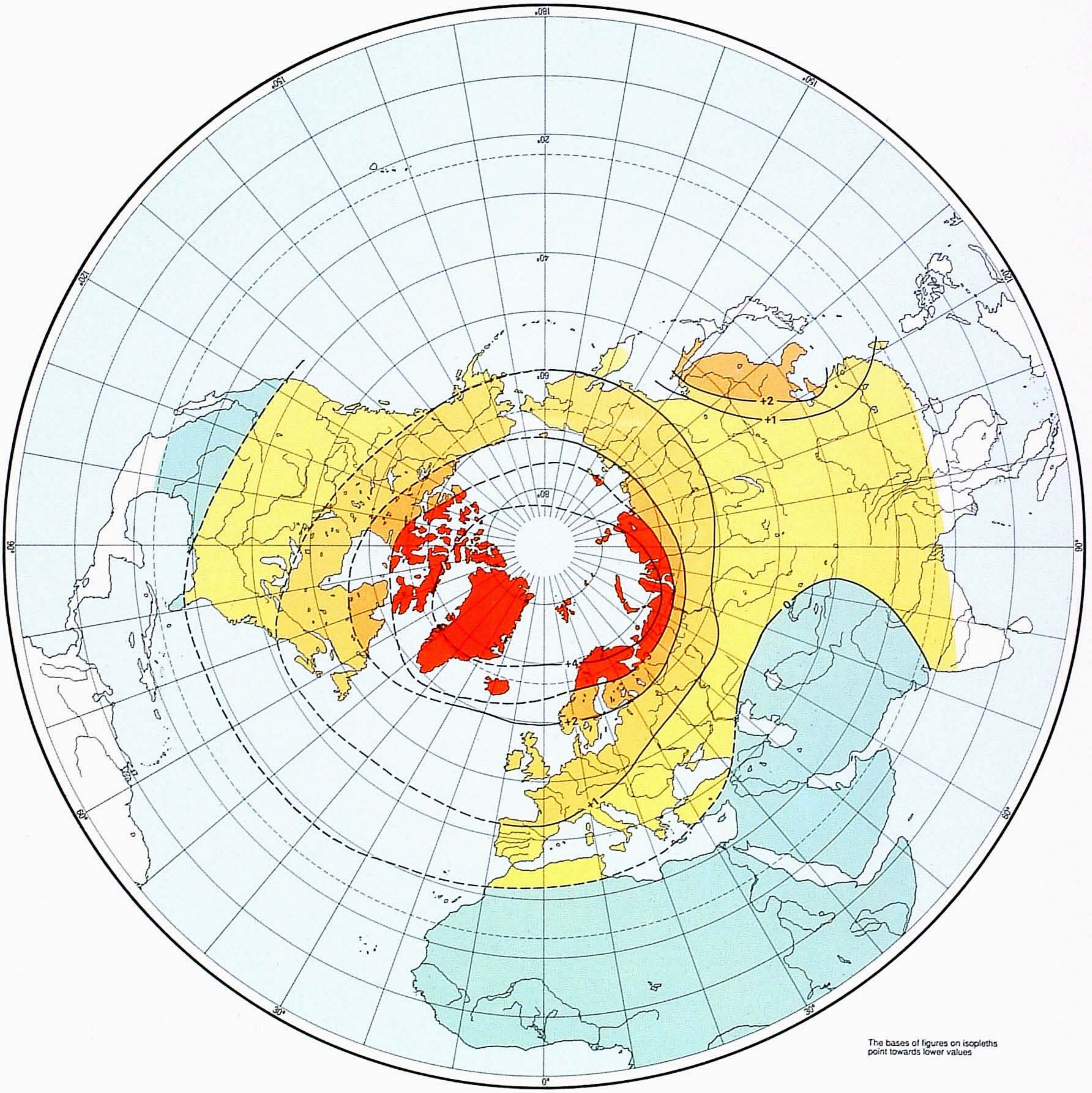
HOLOCENE

(about 6,000 to 5,500 yr B.P.)

JULY MEAN TEMPERATURE

Deviations from present-day values

by V.A. Klimanov, I.I. Borzenkova, A.A. Velichko



°C

- > +4
- +4 — +3
- +3 — +2
- +2 — +1
- +1 — 0
- < 0

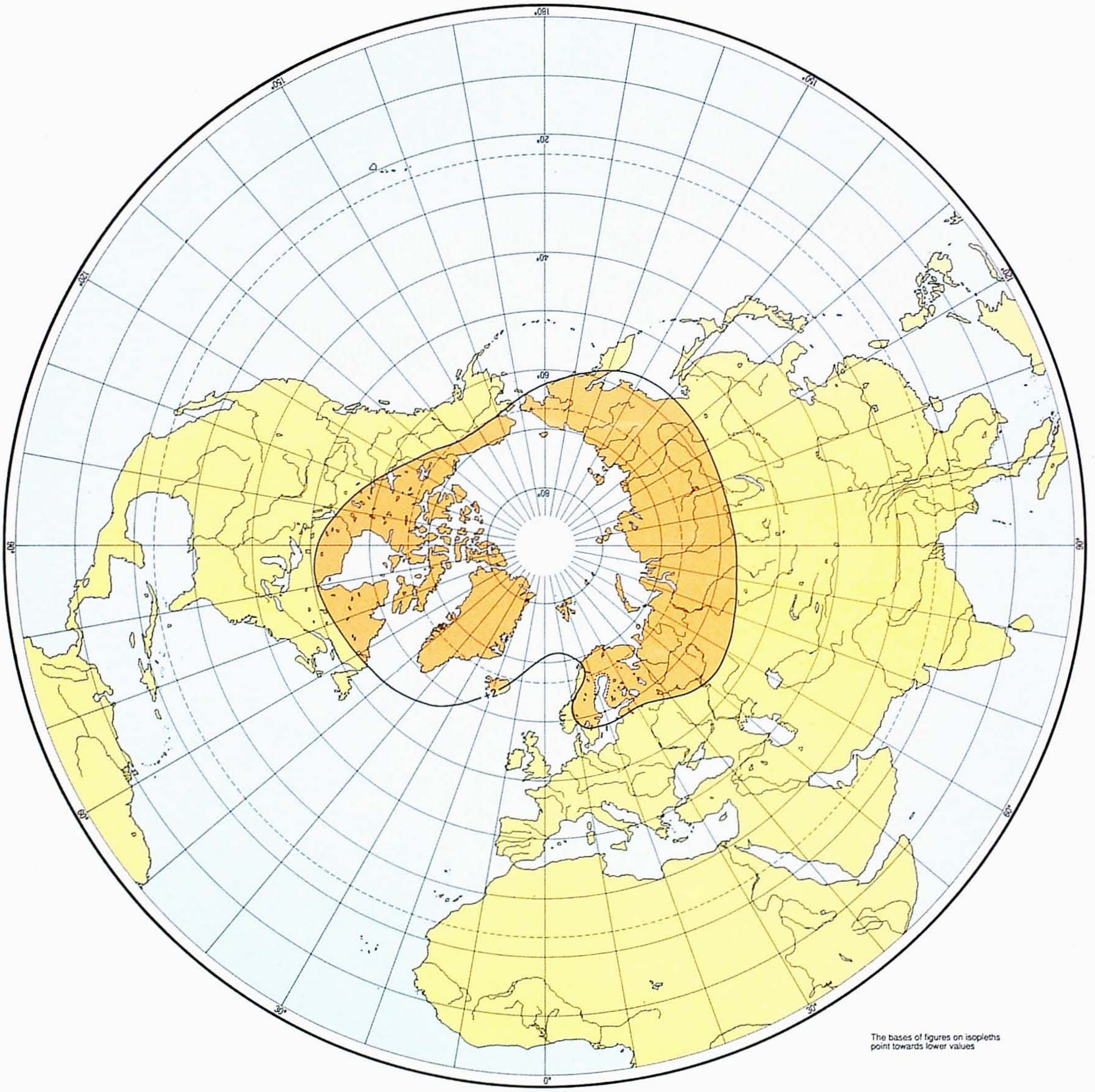
HOLOCENE

(about 7,000 to 6,500 yr B.P.)

AUGUST MEAN TEMPERATURE

Minimal deviations from present-day values

by B. Frenzel



°C

> +2

< +2

.... Coastline deviating from present-day position

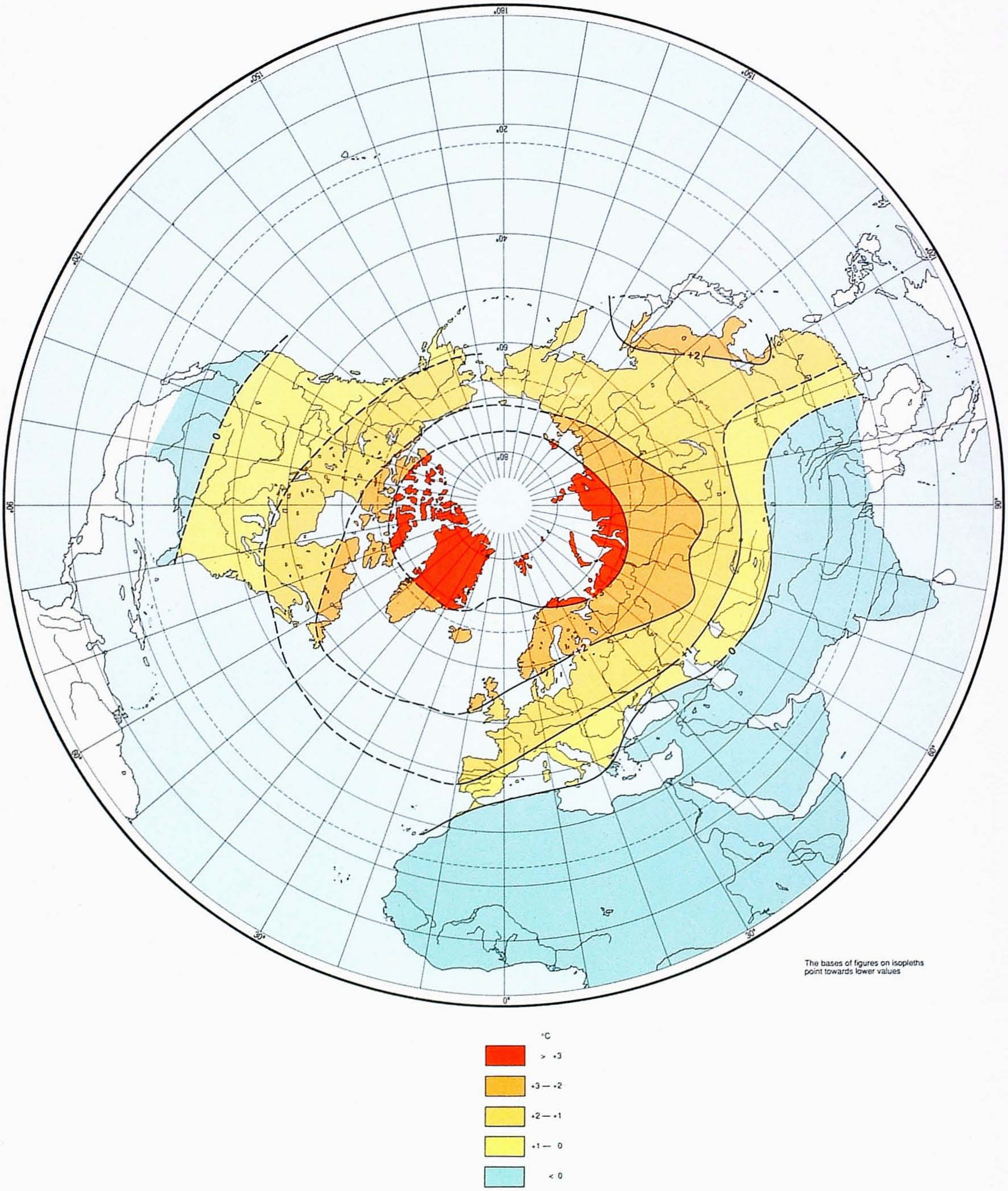
HOLOCENE

(about 6,000 to 5,500 yr B.P.)

ANNUAL MEAN TEMPERATURE

Deviations from present-day values

by V.A. Klimanov



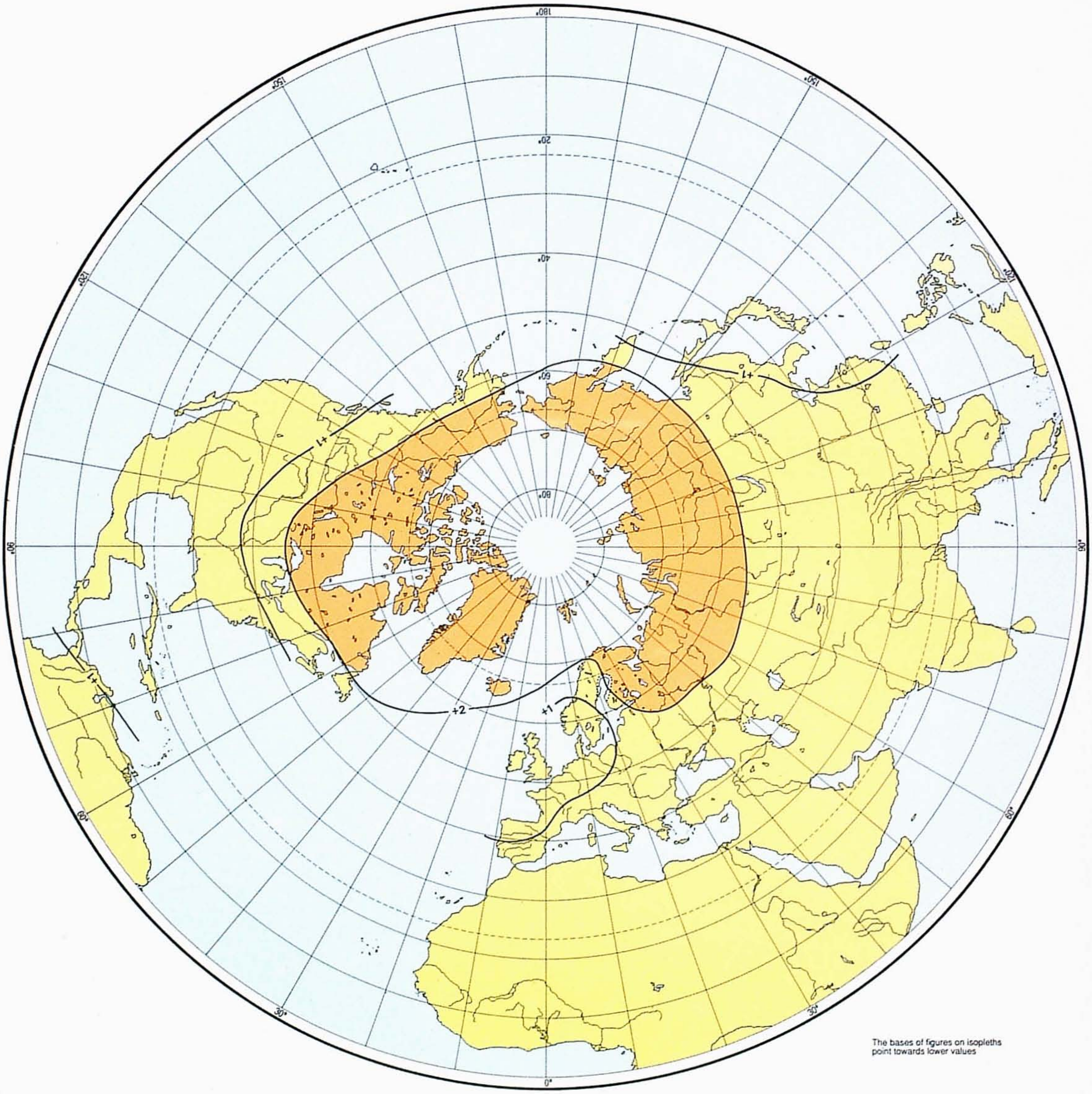
HOLOCENE

(about 7,000 to 6,500 yr B.P.)

ANNUAL MEAN TEMPERATURE

Minimal deviations from present-day values

by B. Frenzel



°C

\> +2

< +2

Coastline deviating from present-day position

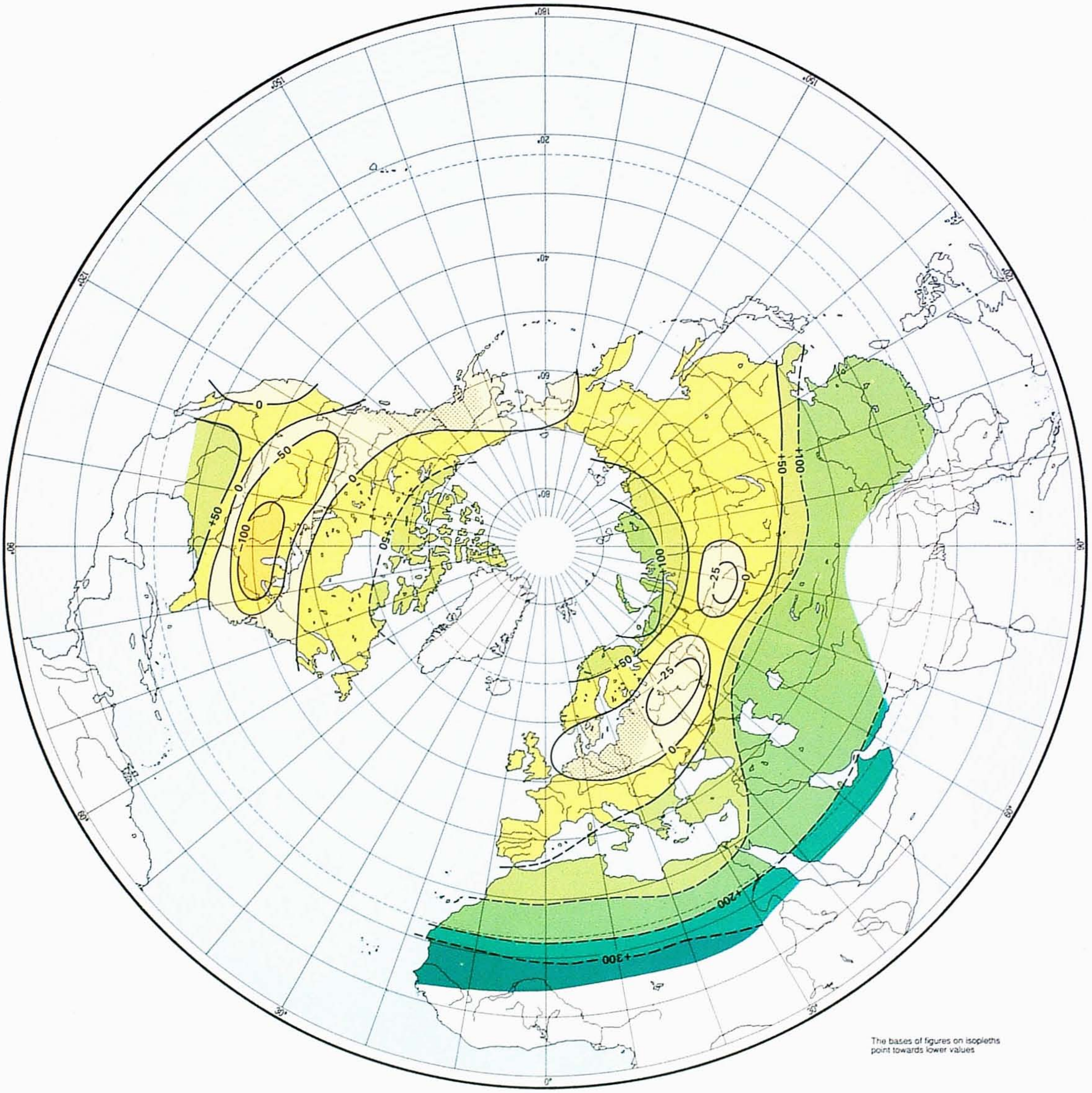
HOLOCENE

(about 6,000 to 5,500 yr B.P.)

ANNUAL PRECIPITATION

Deviations from present-day values

by V.A. Klimanov, I.I. Borzenkova, A.A. Velichko



The bases of figures on isopleths point towards lower values

mm		mm	
	< -100		+50 — +100
	-100 — -50		+100 — +200
	-50 — 0		+200 — +300
	0 — +50		> +300

No deviation

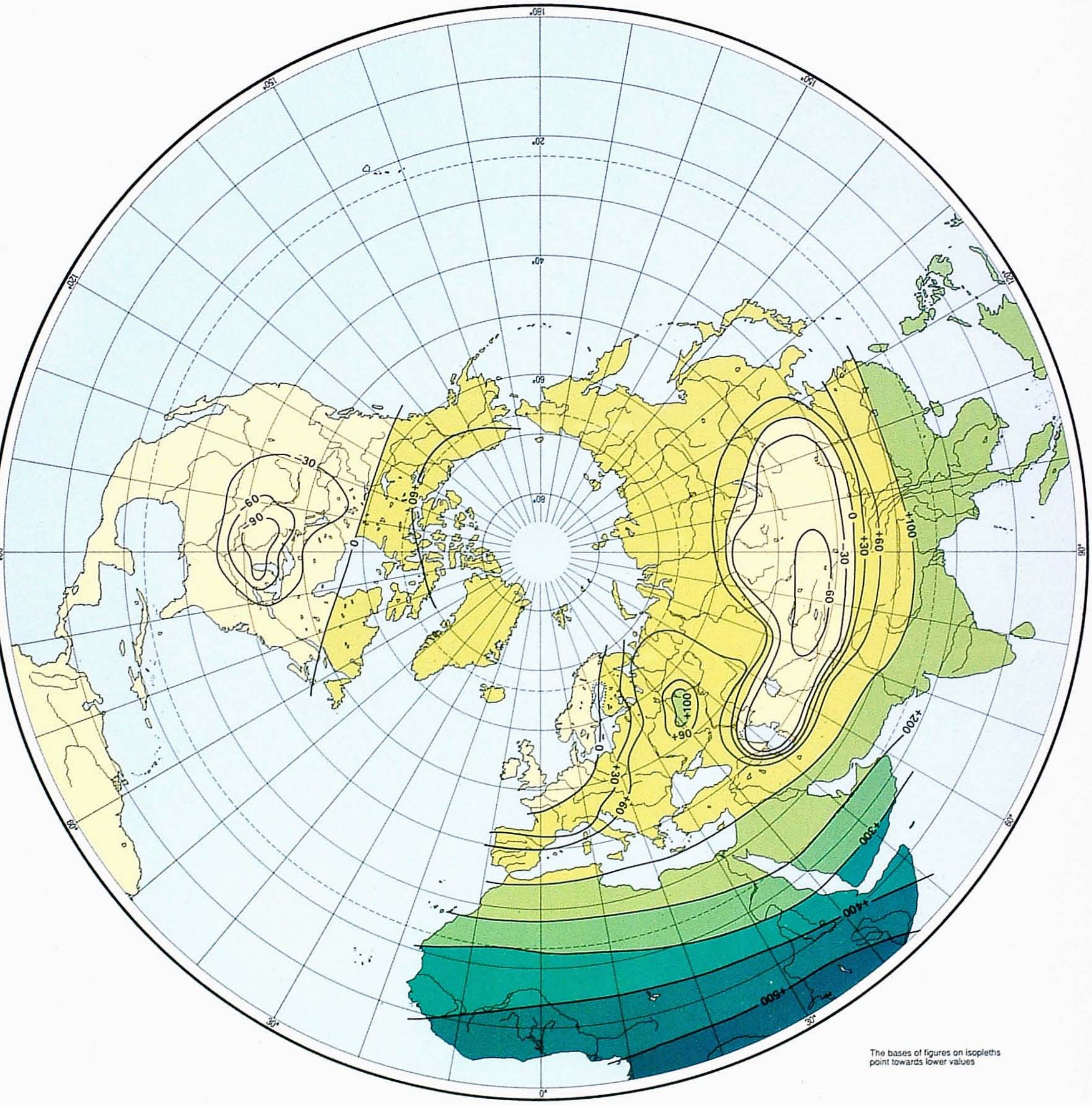
HOLOCENE

(about 7,000 to 6,500 yr B.P.)

ANNUAL PRECIPITATION

Minimal deviations from present-day values

by B. Frenzel



The bases of figures on isopleths point towards lower values

mm		mm
100 — 0		+300 — +400
0 — +100		+400 — +500
+100 — +200		> +500
+200 — +300		Coastline deviating from present-day position