

Late Pliocene–Early Pleistocene Ecologic Changes in the Arctic Ocean Borderland

By CHARLES A. REPENNING *and* ELISABETH M. BROUWERS

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LATE PLIOCENE–EARLY PLEISTOCENE ECOLOGIC CHANGES IN THE ARCTIC OCEAN BORDERLAND

By CHARLES A. REPENNING¹ and ELISABETH M. BROUWERS¹

ABSTRACT

Between 3.0 and 0.85 million years ago the Arctic Ocean Borderland experienced two cycles of climatic deterioration, both of which ended in major Northern Hemisphere continental glaciation. The first cooling cycle ended 2.1 million years ago; continental ice extended down the Mississippi River Valley as far south as Iowa. The cooling cycle was followed by a collapse of glaciation and an extremely warm climate that extended to the northernmost land, almost to the North Pole. This warmth at the beginning of the second cooling cycle was greater than that known at the beginning of the first cycle and cooling began again 300,000 years later, late during the Olduvai Normal-polarity Subchron and about 1.7 million years ago. The second cooling cycle culminated with the classic Ice Ages, beginning 850,000 years ago.

At the beginning of the first cooling cycle, forests, dominated by mixed conifers and including minor but moderately diverse deciduous broadleaf trees, bordered a warm-subfrigid Arctic Ocean. Treeline at lat 79°39' N. indicated an arctic climate remarkably warm by comparison with today's conditions but not so warm as the climate at the beginning of the second cooling cycle. Two successive terrestrial records in Siberian Beringia, four separate but sequential marginal-marine records in Alaskan Beringia and northern Canada, and core records from the Arctic Ocean indicate that between 3.0 and 2.4 million years ago the Arctic Ocean Borderland progressively changed to a conifer forest-tundra ecosystem, then to fully developed tundra. With these floral changes came the first record of permafrost and a frigid Arctic Ocean.

Two Alaskan and one Arctic Canadian shallow-marine records indicate that the Arctic Ocean lacked complete ice cover at least until 2.4 million years ago. Faunal

and depositional records from the central part of the Arctic Ocean indicate that this lack of ice cover may not have extended to the pole; from about 2.48 until about 2.1 million years ago perennial ice cover is thought to have formed intermittently in the central Arctic Ocean. There are no known terrestrial records in the Arctic Ocean Borderland for much of the time of this first intermittent perennial sea ice; however, both sediments and fauna from the beginning and from the end of this period indicate that ice cover did not extend to the shore.

Continental glaciation began in both Scandinavia and Laurentia almost 2.6 million years ago while the Arctic Ocean Borderland was still being deforested and the first permafrost formed in northeastern Siberia and Alaska, but before the first intermittent perennial ice developed in the central ocean area. Continental glaciation in the Northern Hemisphere was most extensive between 2.2 and 2.1 million years ago; in North America ice then flowed down the lowlands of the Mississippi River as far south as Iowa.

Two and a half million years ago, shortly after ice began to accumulate in Scandinavia and Laurentia, depositional records in the Arctic Ocean indicate an abrupt increase in ice-rafted debris, presumably from more extensive mountain glaciation, and a change to prominent density stratification, possibly caused by the influx of greater amounts of freshwater. Arctic marine depositional and faunal records between 2.4 and 2.2 million years ago suggest that the terrestrial climate of the Arctic Ocean region fluctuated. After glacial intensity peaked about 2.1 million years ago, a variety of evidence from the Arctic Ocean, North America, Scandinavia, and northwestern Europe indicates that Northern Hemisphere glacial conditions ended abruptly, thus ending the first cooling cycle.

With the disappearance of glacial climate, mixed coniferous forests returned to the shores of an unfrozen Arctic Ocean, and about 2.0 million years ago treeline

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extended as far north as the northernmost point of Greenland, lat 82°30' N. and about 500 mi (800 km) south of the North Pole. Faunal and floral records at lower latitudes in North America and Europe indicate warmer temperatures and increased precipitation at this time.

The beginning of the second cycle of cooling is not sharply marked. Near the end of the Olduvai Normal-polarity Subchron (about 1.7 m.y. ago), proxy evidence in Europe and North America indicates that the extreme warmth in the Arctic Ocean Borderland had ended. Faunal, floral, and depositional records from the Arctic Ocean, Europe, and North America indicate an oscillation of cool and warm periods between 1.7 and 1.2 million years ago. Both Arctic and Atlantic oceanic temperatures suggest that average conditions were similar to those of today, but the Arctic Ocean was not perennially ice covered.

The severity of the cold periods increased, and about 1.1 million years ago the Arctic Ocean again began to have periods of perennial ice cover. A cold and dry climate, more severe than today, is indicated in western Beringia at this time, and the abundance of arboreal plants was at a minimum. This cold climate was recorded as far south as the Netherlands; however, less than 100,000 years later, during the Jaramillo Normal-polarity Subchron, the pollen spectrum of eastern and western Beringia was similar to that of today, if not somewhat warmer. About 850,000 years ago, ice again flowed down the Mississippi River Valley, beginning the Ice Ages in the United States, and bottom sediments in the Arctic Ocean indicate that perennial ice cover was more frequent. Although the Ice Ages began 850,000 years ago, climatic deterioration apparently did not reach its extreme until about 400,000 years ago.

INTRODUCTION

This synthesis of ecologic changes in the Arctic Ocean Borderland during the late Pliocene and early Pleistocene draws upon published information for Europe westward across North America to eastern Siberia and for areas to the south where information has a bearing on the interpretation of high-latitude sites. Our discussion concentrates, however, on information from Beringia, a vast area covering the Arctic Ocean Borderland in northeastern Siberia and northwestern North America (fig. 1).

More or less by definition Beringia has never been covered by continental glaciation. The region thus extends eastward almost to the Mackenzie River and its delta in northern Northwest Territories, Canada, the western limit of continental glaciation; the western limits of Beringia are beyond the Kolyma Lowlands of Yakutia but have never been clearly defined.

A.V. Sher (1986) applied the name "Beringida" to this area, apparently restricting the name "Beringia" to the area immediately adjacent to the Bering Sea and the Bering

Strait, but his failure to provide a geographic description makes the synonymy uncertain. He also provided the "Beringidian" mammalian age names, the Kutuyakan, the Capedeceitan, the Oiyoran, the Akanan, and the "upper Paleolithic of Siberia"; these lack typification and characterization and also suffer from a lack of geographic constraints, and they are not used herein. The need for age names to aid discussion of these extremely provincial Arctic Ocean Borderland mammalian faunas is discussed by Repenning (in press), although he does not use Sher's terms either.

Throughout the Tertiary and Quaternary, both northwestern North American and Siberian Beringia were a single faunal and floral province; minor east-west differences resulted from ecologic restrictions created by either the Bering Strait or the Bering Land Bridge. About three million years ago the Bering Strait opened (Gladenkov, 1981) and separated eastern and western Beringia, possibly for the first time since the mid-Cretaceous (Marincovich and others, 1990), although recent marine evidence from Kamchatka (Gladenkov and others, 1991) indicates that there may have been an earlier opening about 4.2 million years ago. The Bering Strait has been sporadically present since 3.0 million years ago.

Evidence from the Arctic and North Atlantic Oceans indicates that the opening of Bering Strait 3 million years ago introduced North Pacific marine invertebrates to the Arctic and North Atlantic Oceans. These Pacific invertebrates provide a useful, but somewhat imprecise, maximum age wherever found in deposits of the Arctic Ocean. The use of Pacific invertebrates for other correlation is limited by problems in identifying and dating faunal events. The first appearance of Pacific taxa in Arctic and North Atlantic deposits marks the oldest marine events discussed in this synthesis.

This opening of the Bering Strait is not so clearly marked in terrestrial deposits; it occurred during climatic deterioration of the Northern Hemisphere that began about 3.5 million years ago according to oxygen isotope records. Nonmarine records discussed here are not that old; the earliest, from Meighen Island in Arctic Canada, are inferred to be between 3.0 and 2.9 million years old on the basis of molluscan fauna and paleomagnetic evidence (Fyles and others, 1991), evidence that does not convincingly exclude the possibility that the deposits may be between 4.3 and 3.9 million years old.

The age of about 3.0 million years is supported by Alaskan mammal faunas that cannot be this old but are associated with deposits apparently representing a near-continuous record of both marine and terrestrial climate change with some mollusk and ostracode species that are identical to those found in the Meighen Island locality; these Alaskan faunas shortly postdate the opening of the Bering Strait. The age interpretation of the Meighen Island deposits that is preferred by Fyles and others (1991) is also

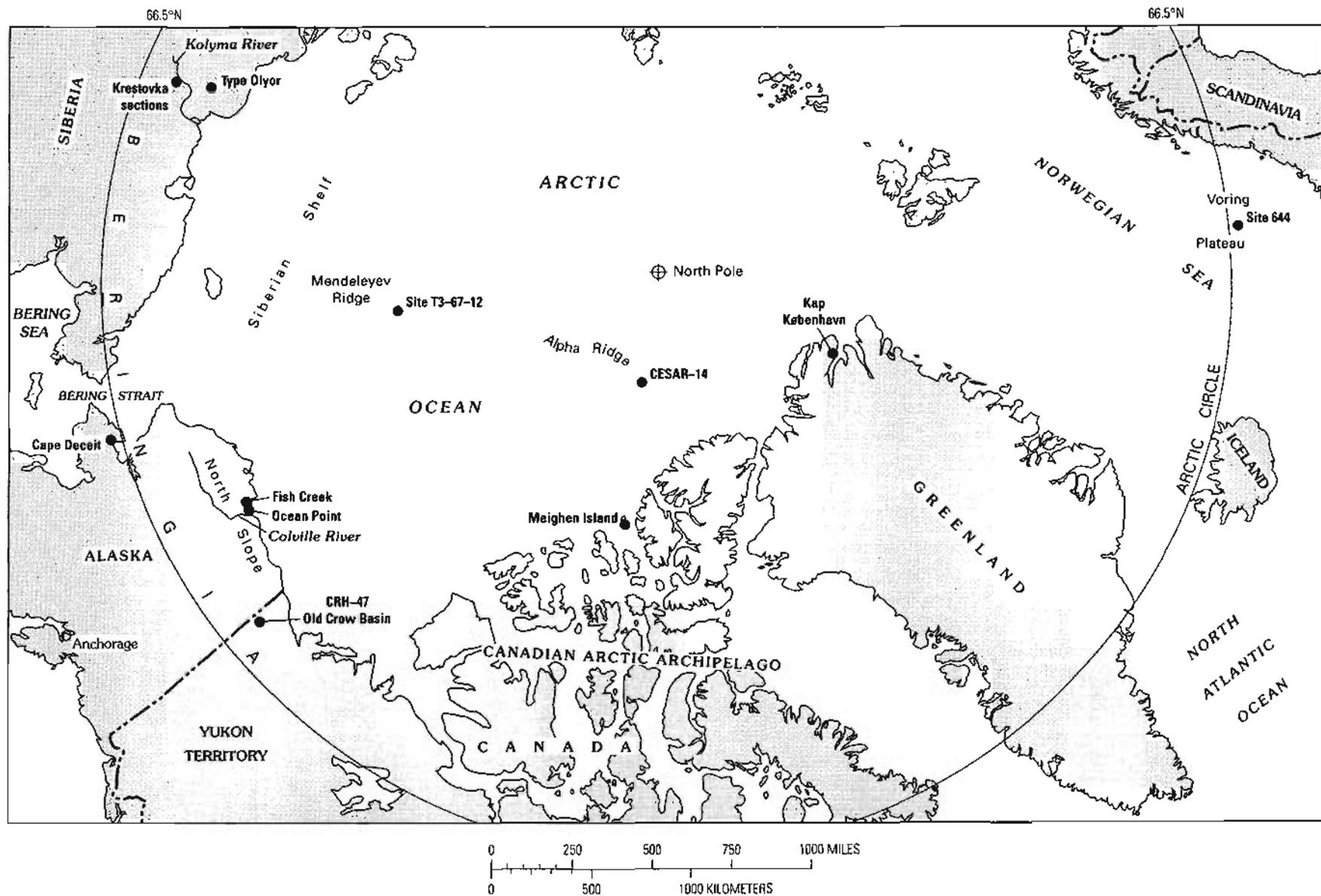


Figure 1. Arctic Basin showing localities discussed in text.

preferred by us. These deposits from Meighen Island of Arctic Canada are the oldest known post-Bering Strait record in the Arctic Ocean basin.

Our reconstruction of climatic changes in the Arctic Ocean Borderland ends with the initiation of the traditional Ice Ages, marked by the first North American ice advances down the Mississippi River Valley that have been called the "Nebraskan" glaciation. This is a term abandoned by Hallberg (1986); Hallberg referred all former named North American Quaternary glaciations before the Illinoian glaciation to the "pre-Illinoian." The first ice advance of the Ice Ages is generally correlated with oceanic isotope stage 22, the peak of which has been dated about 850,000 years ago. All terrestrial evidence regarding the age of this first ice advance of the Ice Ages is in agreement with this date. Some authors have selected a pre-Nebraskan (late Pliocene) glaciation for the beginning of the Ice Ages, but this definition is of recent usage and it is not followed here. Neither usage equates with the beginning of the Pleistocene, which was typified in Italy and, in recent literature (Aguirre and Pasini, 1985), is defined as having begun about 1.67 million years ago.

The temporal limits of our synthesis are dictated by the current limitations of field evidence. Work in progress in the Arctic Ocean Borderland will extend this history prior to the opening of the Bering Strait and will also identify and precisely date localities that are younger than events considered in the present discussion. At present, however, the ages of these localities are not known with sufficient certainty to incorporate into a history of climatic change; significant climatic changes may occur in as few as 20,000 years.

As herein reconstructed, the Arctic Ocean Borderland underwent two major cycles of climatic deterioration during the time discussed, both of which terminated with continental glaciation that developed in and flowed from the European Scandinavian and the North American Laurentian centers. The reason for these two preferred locations of major continental ice accumulation is oceanic; that is, the locations are the result of major latitudinal shifts in North Atlantic current patterns. Behind this immediate cause is a wide spectrum of contributing factors including adjustment to changes in the global oceanic density and solar energy input that results from orbital, atmospheric, and albedo differences. These factors must combine in critical relationships to trigger the onset of glaciation, and no single factor is the cause of glaciation on a continental scale.

If a single cause must be named, it is the gradual rise of Central America between 15 and 3 million years ago. Repenning and Ray (1977) briefly discussed some of the consequences of this shoaling and its age, but the literature on this subject is very large. The shoaling and ultimate emergence of this area completely disrupted former equatorial current patterns in the World Ocean and forced north

and south transequatorial flow to make density adjustments that had previously been accomplished by less complex circulatory cells within individual oceanic basins, with interbasinal differences being adjusted by equatorial currents between the Atlantic and the Pacific. The process of density adjustment has recently been illustrated by computer modeling (Maier-Reimer and others, 1990). Transequatorial flow also varies with other factors such as equatorial heat and geographic configurations; atmospheric climate varies, in turn, with oceanic conditions.

Although the loss of equatorial mixing initiated global climatic deterioration, the regions most sensitive to this change were at polar latitudes. A warming of the North Atlantic produces more atmospheric moisture over adjacent lands such as Laurentia and Scandinavia. If these lands are subject to freezing temperatures, ice will accumulate. Temperatures must be below freezing for the accumulation of ice but need not be much below freezing. Today's minimum temperatures vastly exceed the requirement, but today's precipitation is not adequate for ice accumulation. An increase in atmospheric moisture also produces greater cloud cover, which in turn reduces summer ice melt.

As ice accumulates, other factors come into play, such as decreased insolation and the regional development of high-pressure areas above the ice. These alter global climatic patterns in proportion to the size of the ice field. They are also self-defeating with respect to the growth of continental ice fields; increased solar insolation is needed to provide the energy needed to remove oceanic water for the production of an ice field and greater insolation is needed to allow its growth to a continental scale, but the existence of ice fields reduces insolation and produces atmospheric conditions that deflect moist air masses away from the ice field. Hypothetically an ice field of continental scale should be shaped like a doughnut because the ice field creates a climatic situation in which no precipitation, only evaporation, can reach its center and precipitation, and ice accumulation, is greatest at its margins.

A better understanding of the history of climatic change in the Arctic Ocean Borderland is, consequently, of more than provincial interest. Arctic climatic history is reflected in global climatic history. Furthermore, it is expressed in a region where, because it is cold enough to accumulate ice, this history has had most dramatic effects.

High-latitude climatic deterioration is a reflection of changes of oceanic current patterns, and therefore what happens in the Arctic Ocean region correlates with what happens in lower latitudes. This relationship is justification for the use of proxy data, data from extra-Arctic regions that are contemporaneous with Arctic events and can be used to interpret Arctic conditions. But, to be used in interpretation, the proxy data must correlate in time with events from the Arctic region.

For more than a decade it has been generally recognized that climatic changes approximately correlate with orbital factors of the Earth, which influence total solar insolation (Hays and others, 1976); how other factors modify this influence is still a subject of much consideration. Orbital parameters vary over relatively brief periods of geologic time, the shortest of which is about 20,000 years. To use proxy data in reconstructing the Arctic Ocean Borderland, some means of temporal correlation must be employed that can differentiate time periods as brief as 20,000 years.

Because the late Pliocene through early Pleistocene (3 million to 850,000 years ago) is beyond the useful limit of carbon-14 dating (about 40,000 years) and because opportunities for the use of other radiometric dating methods are few in the Arctic region, our ability to correlate rocks of this time interval in the Arctic region has not kept pace with our ability to correlate such rocks in the rest of the world. Other than paleomagnetic stratigraphy, little evidence of geologic age is available in the Arctic to support biochronologic age inferences. Further, biochronologies in the Arctic region are drastically influenced by the extreme latitudinal provinciality of Arctic faunas and floras. Attempts to project biochronologies developed in lower latitudes into the Arctic region have consistently met with the conflicting aspects of latitudinal provinciality, which causes the unique biotic associations of the Arctic.

Except for pioneering work attempting to determine the age of the opening of the Bering Strait (Hopkins, 1967; Gladenkov, 1981; Gladenkov and others, 1991), success in determining geologic age in the Arctic has been minimal. The recognition that late Pliocene and early Pleistocene terrestrial faunal events are recorded in western Beringia has only existed for the last 20 years (Sher, 1971) and in central and eastern Beringia only for the last 8 years (Repenning, 1983, in press; Repenning and others, 1987). Terrestrial faunal events, which demonstrate the passage of time, have been recognized in temperate latitudes for more than 200 years, and their records are abundant. Only seven or eight localities of known age recording the terrestrial biota of the Arctic Ocean Borderland are discussed in this report.

AGE ASSIGNMENTS

The chronology used herein is based on the biochronology of the microtine rodents of the Northern Hemisphere (Repenning and others, 1990). This biochronology is new and, therefore, has not been successfully applied by others to faunas of the Arctic Ocean Borderland. Its use is limited to the past five and a half million years (the time during which microtine rodents have been a significant part of the world fauna).

Within this time span, the biochronology can distinguish time differences with a minimum accuracy of 250,000 years in most areas of the Northern Hemisphere; using adequate faunal samples of the better known lineages of microtine rodents, faunal differences can be identified that are as little as 5,000–10,000 years different in age (Repenning, in press). This is possible because microtine rodents evolved very rapidly, but it must also be admitted that adequate faunal samples of those microtine rodents having significance are not yet common in the Arctic region.

This time discrimination is sufficiently precise to identify paleomagnetic and climatic events of known age. Age assignments based on microtine rodent evolution and dispersal history can thus be refined by correlation with these events, and climatic events can be identified with sufficient certainty to be used in support of correlation. For example, the two normal-polarity excursions within the Reunion Normal-polarity Subchron of the Matuyama Reversed-polarity Chron are separated by 80,000 years, a span of time too brief to be detectable in much of the presently known microtine history of that period of time. Positive climatic evidence indicates, however, that continental glaciation was extensive in North America during the older of these two excursions and that climates were warm during and following the younger excursion. If the microtine biochronology can identify a polarity event in an outcrop as being one of the two Reunion excursions, it is considered reasonable to identify which excursion is represented on the basis of climatic evidence.

Mammals other than microtine rodents lived in or have dispersed through Beringia, but either their rates of evolutionary change were too slow or their history too imprecisely known for similarly accurate age interpretation. Correlations in the Arctic Ocean region that are based on marine invertebrates are basically the application of marine paleoclimatic stratigraphy with an often misleading assist from the history of the invertebrates in lower latitudes, just as floral evidence is an application of terrestrial paleoclimate stratigraphy. Neither biologic group has an evolutionary history that is well enough understood in the Arctic to be used in a precise biochronologic sense, although both are most helpful in reconstructing past environments.

Eastern Beringia has maintained many faunal features of the northeastern Asian ecosystem because of its isolation from the rest of North America during continental glaciation. Because of increasing latitudinal differences in environment during climatic deterioration, these features were retained during periods in which no continental ice was present. Lower latitude North America had very little influence on the Beringian fauna. Because of increasing latitudinal zonation and provinciality during the late Pliocene

and early Pleistocene, faunal influence from lower latitude Eurasia decreased. The Eurasian fauna filtered into Beringia and dispersed southward into North America, but very little of the North American fauna moved northward into Beringia.

For the most part, similarities between Beringian terrestrial faunas and those of temperate regions have resulted from the southward dispersal of fauna during cold periods of the late Pliocene and early Pleistocene. That northward dispersal of temperate faunas into Beringia took place during warm periods is obvious, but records from the Arctic Ocean Borderland and from regions to the south strongly suggest that these elements became locally extinct with the return of more severe climates. This ecological constraint has made Beringia a "filter" dispersal route (Simpson, 1940). The record of the passage of mammalian taxa through Beringia is mostly that of their appearance in temperate North America or Eurasia. Commonly there is no record of the dispersing organism in Beringian faunas because its passage through this filter dispersal route was brief and its discovery in the fossil record is, therefore, very fortuitous (Repenning, in press).

The biohistorical aspects of Arctic North America must be considered provincial; they resulted from different dispersal and evolutionary histories, as well as from different ecosystems, than those that produced the faunal history of the United States. If "North American" mammal ages are applied to Beringian faunas, they must be redefined. Logically, an immigrant mammal, whose first appearance from higher latitudes defines a particular age in the contiguous United States, will be characteristic of an older age in the higher latitudes from whence it came. One should expect an Arctic fossil mammal fauna that seems characteristic of a North American Provincial Mammalian Age to represent an older age in Beringia. And, in fact, Arctic faunas in both the New and Old Worlds have proven to be older than conventional faunal definitions indicate.

The North American Provincial Mammalian Ages of Wood and others (1941) were originally considered as representing the North American Tertiary mammalian fauna. They have come to be considered also as nonmarine biochronologic units, although Wood and others specifically included marine beds in which mammals are found. Possibly the last usage of such colonial thinking in the United States, the "provincial" part of their nomenclature was in distinction from "European ages," which were, in 1941, difficult to correlate to the United States. Although it would be unreasonable to think that they did not believe such existed, Wood and others (1941) expressed no awareness of latitudinal provincial distinctions, and they included Mexican and southern Canadian faunas and formations in their correlations. Their ages were assumed to be North American in scope but were defined primarily on the basis of the fossil faunas of the conterminous United States.

Consequently, although mentioned herein a few times in explanation, the North American provincial mammalian ages of Wood and others (1941) are not used in this discussion of the Arctic Ocean Borderland and the Beringian faunal province; faunal ages are given only in radiometric years.

ISOTOPE AND POLARITY RECORDS

The oceanic cores used for correlation in this report are (1) site 552A (Shackleton and others, 1984; re-evaluated by Raymo and others, 1989) on the western side of the Rockall Plateau south of Iceland; (2) site 644 (Jansen and others, 1988) on the Voring Plateau in the Norwegian Sea off of Norway (fig. 1); (3) site CESAR-14 (Scott and others, 1989) near the eastern end of the Alpha Ridge in the central Arctic Ocean (fig. 1); and (4) site T3-67-12 (Herman and others, 1989) on the Mendeleev Ridge off the Siberian Shelf in the western Arctic Ocean (fig. 1). Although used only occasionally, the numbers applied to oxygen isotope stages are those used by Ruddiman and others (1986) and extended by Raymo and others (1989); stage numbers mentioned in the text are shown in figure 2.

We accept most of the radiometric age data for the late Pliocene and Pleistocene polarity events from the original presentation by Mankinen and Dalrymple (1979) because the biochronology here used is (1) based on radiometric age determinations rather than presumed constant rates of sea-floor spreading, which have been used to modify radiometric ages, and (2) was constructed on the Mankinen and Dalrymple calibration of the paleomagnetic stratigraphy.

Several age assignments differ from those in Mankinen and Dalrymple. (1) The beginning of the Olduvai Normal-polarity Subchron of the Matuyama Reversed-polarity Chron is adjusted to 1.9 million years ago because the very securely dated KBS Tuff of Kenya is 1.88 ± 0.02 million years old, and this tuff lies at least 20 m stratigraphically above the base of the Olduvai Subzone (McDougall, 1985). (2) The age of the Cobb Mountain Normal-polarity Subchron of the Matuyama Chron is 1.10 million years (Mankinen and others, 1978). (3) The beginning of the Brunhes Normal-polarity Chron is close to 0.79 million years before present (Izett and Obradovich, 1991).

FLORAL TERMS

Use of the terms "mixed-conifer forest," "boreal forest" (taiga), and "tundra" is conventional. Pollen percentages cited herein that include some tundra plants as part of the flora could fall into the classification "forest-tundra" of some authors; however, such detail has little significance in the present discussion. The term "tundra" is intended to indicate extensive tundra well removed from any arboreal

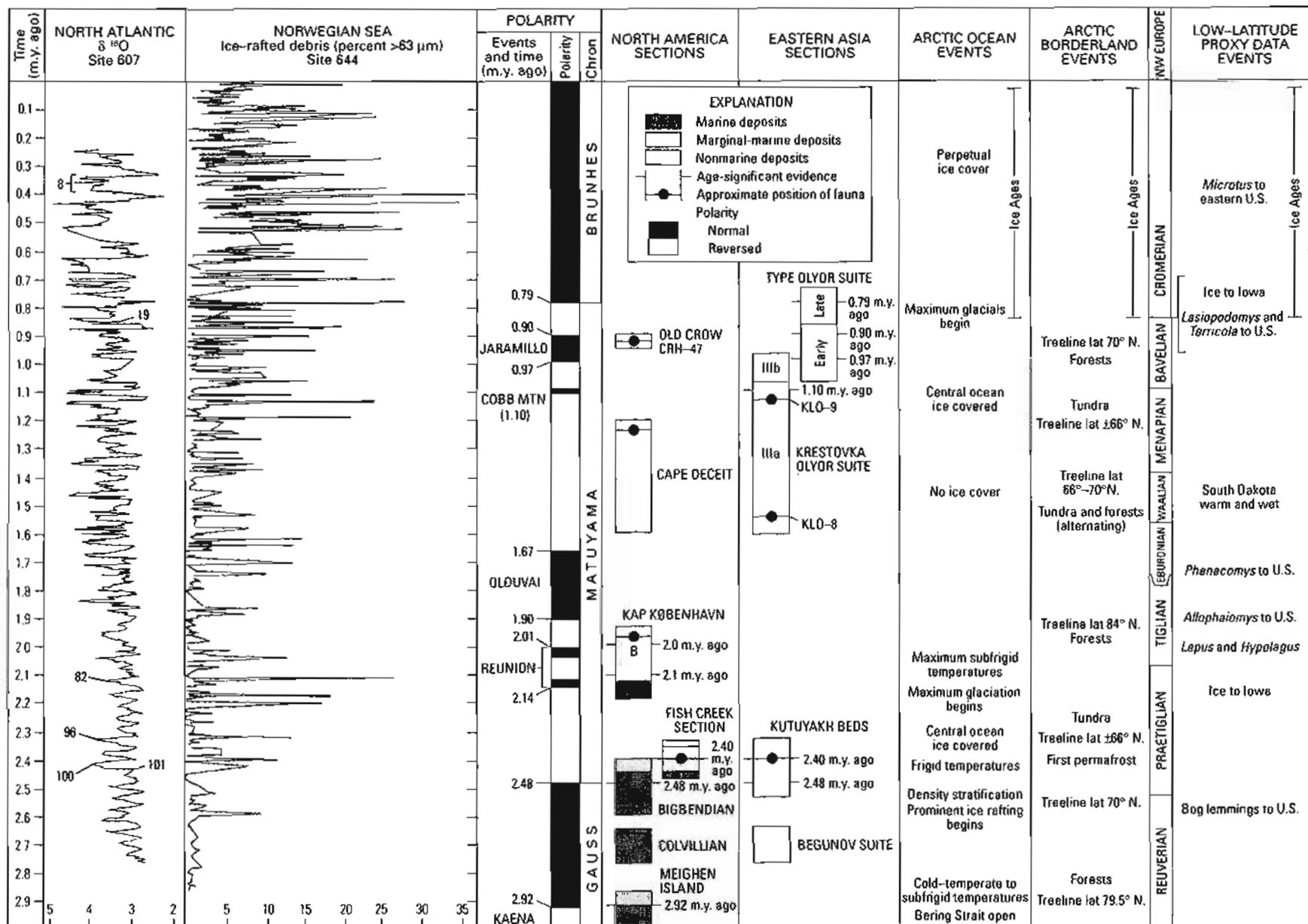


Figure 2. Correlation of localities and events discussed in text. Data for site 607 from Ruddiman and others (1989) and Raymo and others (1989); data for site 644 from Jansen and others (1988). Data for sites 607 and 644 have not been corrected for differences in polarity age assignments. Sources for other data are given in the text. Only those oxygen isotope stages that are mentioned in the text are numbered.

plants. In addition, many pollen samples cited have a high percentage of grass, such as in the younger part of the Kutuyakh Beds (Pliocene) of Sher and others (1979), and these floras might better be referred to as steppe or forest-steppe mosaic than as forest-tundra mosaic, but the distinction is not emphasized here.

Acknowledgments.—This summary of the late Pliocene and early Pleistocene history of the Arctic Ocean Borderland is a consolidation of information from many sources that has accumulated over the past 10 years. During this time we have benefited from the work of many people, and we would like to particularly thank Thomas Ager (U.S. Geological Survey, Denver), Ole Bennike (Geological Museum, University of Copenhagen), Julie Brigham-Grette (University of Massachusetts, Amherst), L. David Carter (U.S. Geological Survey, Anchorage), Oldrich Fejfar (Charles University, Czechoslovakia), Rolf Feyling-Hanssen (Aarhus University, Denmark), Svend Funder (Geological Museum, University of Copenhagen), Niels Jorgensen (Institute of Historical Geology and Paleontology, University of Copenhagen), Louie Marinovich, Jr. (U.S. Geological Survey, Menlo Park), John Matthews, Jr. (Geological Survey of Canada, Ottawa), David Penney (Aarhus University, Denmark), Andrei Sher (Academy of Sciences of the U.S.S.R., Moscow), Charles Schweger (University of Alberta, Edmonton), Jack Wolfe (U.S. Geological Survey, Denver), and Vladimir Zazhigin (Academy of Sciences of the U.S.S.R., Moscow). Thomas Ager, Tony Bryant, Oldrich Fejfar, Art Grantz, Patricia Holroyd, John Matthews, Hugh Rieck, Richard Tedford, and Jack Wolfe have reviewed the manuscript and offered suggestions for its improvement.

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3.0-2.0 MILLION YEARS AGO: THE FIRST COOLING CYCLE

As herein considered, the maximum age for the opening of Bering Strait is about 3.0 million years. This date is based on a dated lava flow that is interbedded with sediments containing the first occurrence of Pacific mollusks in Iceland (Gladenkov, 1981). Recently, evidence has been described (Gladenkov and others, 1991) that suggests an opening of the Bering Strait about 4 million years ago, but it contains some discrepancies between the biostratigraphic and radiometric dates. In addition, we are aware of no evidence for the dispersal of North Pacific invertebrates into the Arctic and North Atlantic Oceans during this time period. Known relationships between marine and terrestrial climatic trends and associated mammal faunas all favor the

interpretation of an opening of the Bering Strait about 3.0 million years ago, as has been discussed.

CANADA, MEIGHEN ISLAND: BEAUFORT FORMATION

The commingling of Pacific-derived mollusks and Atlantic-derived mollusks that were displaced from the Arctic Ocean in later faunas has been described from the Beaufort Formation (Pliocene) of Meighen Island, Canadian Arctic Archipelago (lat 79°30' N.) (Fyles and others, 1991) (fig. 1). The molluscan fauna is believed to represent a rather brief time immediately after introduction of Pacific forms into the Arctic Ocean and prior to extirpation of some endemic Arctic-Atlantic species. The deposits are reversely polarized and bounded above and below by normally polarized deposits.

The interpretation preferred by Fyles and others (1991) is that the reversed polarity associated with the molluscan fauna represents one of the two reversed-polarity events of the Gauss Normal-polarity Chron and indicates an age between 3.1 and 2.9 million years. This interpretation is in agreement with other evidence discussed herein and so is also preferred by us. The other possible interpretation is that some time within the Gilbert Chron is represented.

Available faunal data indicate a cold-temperate marine climate with temperatures comparable to the modern North Atlantic south of Newfoundland, water temperatures not present in the modern Arctic Ocean. Plant and insect fossils in the overlying, normally polarized organic deposits indicate a treeline flora that differs from modern treeline floras (now about 1,230 mi (1,980 km) to the south at lat 64° N.) only in having a richer variety of taxa, with as many as six types of conifers, some now extinct (Fyles and others, 1991).

Four species of shallow-marine ostracodes (*Cytheretta reshekpukensis*, *Acanthocythereis dunelmensis*, *Sarsicytheridea bradii*, and *Thaerocythere* aff. *T. crenulata*) are present. These forms of Atlantic origin permit little chronological interpretation. Their earliest known Arctic Ocean record is in the pre-Bering Strait deposits of the Nuwok Member (Oligocene to Pliocene, but see Brouwers, in press) of the Sagavanirktok Formation (Cretaceous to Pliocene) from the eastern Alaskan North Slope. Ecologically, they indicate a cold-temperate to warm-subfrigid marine climate, compatible with the environment indicated by the mollusks.

No ostracodes of Pacific origin are present. Present records from the Arctic Ocean suggest that ostracodes disperse more slowly than mollusks, presumably because of their benthic habits throughout their life cycle, as contrasted with the readily mobile and dispersed pelagic habit of molluscan veligers.

ALASKAN NORTH SLOPE: GUBIK FORMATION

In the Alaskan North Slope area (fig. 1), at about lat 70° N. and 9.5° farther south than the locality on Meighen Island, deposits younger than the opening of the Bering Strait are included in the Gubik Formation (Pliocene and Pleistocene). This formation was deposited across a stable coastal plain, the Alaskan North Slope, that is bounded by the Brooks Range to the south and the Arctic Ocean to the north. A number of eustatic sea-level rises have resulted in marine transgressions across this plain that deposited a series of interbedded shallow-marine and nonmarine deposits comprising the Gubik Formation. The transgressions have been identified by geomorphic characteristics, faunal suites, and stratigraphy; they have been informally named by Carter and others (1979). Originally thought to be of middle and late Pleistocene age, those now known to be of late Pliocene age (Repenning, 1983; Repenning and others, 1987; Carter and Hillhouse, 1992) have been called the Colvillian, Bigbendian, and Fishcreekian transgressions, listed in depositional order.

COLVILLIAN AND BIGBENDIAN TRANSGRESSIONS

Diverse ostracode faunas, comprising 54 species, are present in deposits of the Colvillian and Bigbendian transgressions, the first deposits of the Gubik Formation of the Alaskan North Slope (Brouwers, in press). The oldest recognized transgression of the Gubik Formation, the Colvillian transgression, is marked by a mixture of Pacific and Atlantic species and therefore is younger than the Bering Strait.

Geographically widely separated localities in the eastern Alaskan North Slope (eastern Beringia) are uniform in species composition and contain comparable ostracode species associations that include *Acanthocythereis dunelmensis*, *Elofsonella concinna*, *Heterocyprideis sorbyana*, *Palmenella limicola*, *Pterygocythereis vannieuwenhuisei*, *Rabilimis paramirabilis*, and *Robertsonites tuberculatus* (Brouwers, in press). These assemblages are dominated by Atlantic-derived genera (*Sarsicytheridea*, *Rabilimis*, *Pterygocythereis*, and *Heterocyprideis*) but contain some Pacific-derived genera (*Finmarchinella*, *Palmenella*, *Hemicythere*, *Howeina*, *Pectocythere*, and *Robertsonites*).

These oldest post-Bering Strait assemblages of the Alaskan North Slope indicate a stable paleoenvironment with little to no fluctuation of the physical-chemical environment. These conditions are typical of deeper inner shelf to middle shelf water depths. Based on the ostracodes, the shallow Arctic Ocean was much warmer than today, typical of North American northern cold-temperate to southern subfrigid marine climatic zones (northern Nova Scotia and southern Labrador) and the Transitional and Norwegian biogeographic provinces of Europe (Hazel, 1970).

The ostracode faunas of these two earliest transgressions (Colvillian and Bigbendian) are indistinguishable (Brouwers, in press). The mollusk assemblage from the lower, normally polarized part of the Bigbendian deposits indicates that the deposits are younger than the about 3.0-million-year opening of Bering Strait and that the Arctic Ocean was not frozen close to the then-existing shore during the winters (Repenning and others, 1987; L. Marinovich, Jr., unpublished data). The presence of an archaic sea otter also indicates that the ocean was not frozen to its shore at any time during the year and that the deposits are less than 3 million years old (Repenning, 1983).

Deposits of both transgressions were first recognized in the vicinity of Ocean Point, along the Colville River, where deposits of both the Colvillian transgression and the older part of the Bigbendian transgression are normally polarized. The upper part of the Bigbendian deposits have a reversed magnetic polarity that is believed to represent the beginning of the Matuyama Reversed-polarity Chron (Carter and Hillhouse, 1992). The evolutionary stage of the archaic sea otter found at Ocean Point in a Bigbendian mollusk bed having normal polarity and possibly 16 ft (5 m) below the polarity reversal (stratigraphic position indicated by Carter and Hillhouse, 1992; specimen collected by Carter in 1975) indicates an age *older than* the Olduvai Subchron of the Matuyama Chron and *younger than* 3 million years (Repenning, 1983).

Pollen from the Colvillian and Bigbendian deposits at the Ocean Point section indicate a terrestrial climate considerably warmer than today, possibly warmer than modern Anchorage, Alaska, which is north of the modern range of some of the trees recorded (Nelson and Carter, 1985, 1992). Based on the flora, permafrost could not have been present. Nelson and Carter (1992) suggested that the flora of the Bigbendian transgression represents a more "open spruce-birch woodland or even parkland" than the flora of the Colvillian transgression. They noted the absence of hemlock and a reduction of pine in the Bigbendian flora and suggested that the climate was slightly more severe than that during the earlier Colvillian transgression.

There is no record of extensive tundra or permafrost in these earliest post-Bering Strait deposits at about lat 70° N.; today, the vegetation of the Alaskan North Slope consists of full tundra and extensive permafrost.

Although the mixed conifer and deciduous forests present in Alaska during the Miocene (Wolfe and Tanai, 1980) had mostly withdrawn southward from the Arctic Ocean Borderland, spruce-birch forests with some pine, hemlock, larch, and poplar were still present less than 3 million years ago in the latitude of Beringia (Nelson and Carter, 1985, 1992). To judge from the pollen records, particularly those of the Bigbendian transgression, these remnants of the early Pliocene forests included open areas that supported tundra plants and, at times, significant percentages of grass; a forest-tundra or forest-steppe mosaic is indicated.

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Coldwater ostracode faunas, typical of the Fishcreekian deposits, are present in the upper part of this basal silty unit (locality 78ACr60a in Repenning and others, 1987). The upper part of the lower silt unit also contains the oldest pollen flora, which indicates shrub tundra (Repenning and others, 1987, fig. 3). Amino acid ratios of marine mollusks presumed to be from the oldest part of the Fish Creek section also indicate that these basal silt deposits are older than the overlying part of the section; the amino acid ratios are similar to those from deposits of the Bigbendian transgression at Ocean Point (Brigham-Grette and Carter, 1992).

Although Brigham-Grette and Carter (1992) assigned the entire lower silt unit of the Fish Creek section (~13 ft (4 m) thick) to the Bigbendian transgression, the warmwater ostracode fauna was found only in the lowermost meter of this silt unit, and coldwater ostracodes of the Fish Creek fauna, as well as the oldest tundra pollen floras, were found in the upper part of this lowest unit. The climatic change from open forests and warm water of the Bigbendian transgression to shrub tundra and cold water of the Fishcreekian transgression occurred entirely during deposition of the lower silt unit of the Fish Creek section.

The entire Fish Creek section is reversely polarized (Carter and Hillhouse, 1992). A single basal sample reported to be normally polarized (Repenning and others, 1987) was recollected and the initial polarity determination found to be in error (Carter and Hillhouse, 1992). Thus, warmwater ostracodes are present later in the sedimentary record than the end of the Gauss Chron and are found in deposits having a reversed polarity that represents the earliest Matuyama Chron. Probably the climatic change from warm to cold water and from forest to tundra was (1) rather abrupt, (2) during the oldest part of the Matuyama Chron, (3) entirely within the lower silt unit of the Fish Creek section, and (4) not marked by any depositional change, occurring instead at the peak of the transgression because the remainder of the Fish Creek section consists of regressive deposits.

The type section of the Fishcreekian "transgression" grades upward from flatbedded, silty marine deposits to strongly crossbedded sandy units considered to be either tidal channel deposits or nearshore wave-washed sands (Repenning and others, 1987). These sandy units contain shallow-water benthic foraminifers and mollusks, as well as intermixed terrestrial and marine mammals. Upsection, the ostracode fauna contains increasing numbers of nonmarine species. The marginal-marine deposits are overlain by nonmarine sands.

The type section of the Fishcreekian is a regressive sequence and apparently records the end of the Bigbendian transgression as well as the first cooling period following the warm, transgressive phase of the Bigbendian. It also apparently records the first cooling period during the Matuyama Chron. How this regressive sequence relates to

evidence of a Fishcreekian transgression elsewhere depends on field evidence that we do not have.

The polarity change from normal to reversed in the Bigbendian deposits represents the Gauss-Matuyama boundary, an interpretation of Carter and Hillhouse (1992) that is not disputed. The Ocean Point and Fish Creek localities record a single history of regression from the peak of the Bigbendian transgression. The transgression began with a warmwater fauna and open forests on adjacent lands. The regressive deposits record a northward (seaward) retreat of the Arctic Ocean from Ocean Point to a point north of Fish Creek and a coldwater fauna with tundra on adjacent lands. The Bigbendian transgression culminated during the earliest Matuyama Reversed-polarity Chron, and the Arctic Ocean withdrew north of Fish Creek as the oceanic water cooled abruptly and terrestrial climatic deteriorated.

During the withdrawal of the Arctic Ocean, its mollusk fauna remained about the same. Primitive sea otters are known from both the "warm" (Bigbendian) and "cold" (Fishcreekian) parts of the transgression, but the ostracode fauna changed to a modern coldwater fauna before the end of deposition the oldest unit of the Fish Creek section. The terrestrial vegetation changed dramatically, from the forest described by Nelson and Carter (1992) for the Colvillian, to an open forest lacking pine and hemlock during the Bigbendian transgression, to shrub tundra and evidence of distant forests during the early Fishcreekian regression, to a wet tundra and possible permafrost and less evidence of distant forests during the late Fishcreekian regression. The youngest deposits of the Fish Creek section are nonmarine.

Subsequent to the publication of Repenning and others (1987), much has been learned about the ages of the depositional units shown on their figure 13 (Repenning and Grady, 1988; Repenning, in press). However, aside from raising the base of the Fish Creek section so that normal polarity (then thought to have been present) is not shown at its base, the age assignment of 2.4 million years for the mammal fauna in the upper part of the section is still most likely. The oxygen isotope record for North Atlantic site 552 (Shackleton and others, 1984; modified by Raymo and others, 1989) indicates that the first and most prominent eustatic sea-level regression following the end of the Gauss Chron began about 2.44 million years ago and peaked about 2.40 million years ago (site 607, fig. 2).

If the regression recorded by the Fish Creek section represented any later, and *less extreme*, sea-level regression, one would expect to see some fluctuation in depositional environments and oceanic temperatures as reflected in the ostracode faunas of the Fish Creek section, and none was found. If other transgressive cycles had taken place between the Bigbendian transgression and the Fishcreekian regression, more time would be involved and one would expect the warmwater ostracodes at the base of the Fish Creek section to differ from the Bigbendian ostracode fauna; this was the time of the most marked change of the

ostracode fauna of the Arctic Ocean. One would also expect some dissimilarity in the amino acid ratios of the Bigbendian and basal Fish Creek deposits if the basal beds of the Fish Creek Section did not represent the transgressive phase of the Bigbendian. The terrestrial climate trends, as indicated by the pollen from Colvillian and Bigbendian deposits, show the same uninterrupted trend in environmental cooling, and no suggestion of fluctuation. That a climatic fluctuation would be detectable in the pollen profile is demonstrated by the record of the youngest part of the Kutuyakh Beds in Yakutia, to be discussed; this pollen record was deposited during the same time period but lasted a bit longer, in that it records the return of forests following the conclusion of the cold period about 2.4 million years ago.

In brief, there is only one uninterrupted record in the Fish Creek Section showing regression, marine cooling, and faunal and floral change between the Bigbendian transgression and the Fishcreekian regression; no evidence suggests that the Fishcreekian regression represents anything other than a single (the first) sea-level drop following the Bigbendian transgression that began before the end of the Gauss Chron.

Carter and Hillhouse (1992) suggested that considerable time elapsed between the reversely polarized part of the later Bigbendian deposits at Ocean Point and the reversely polarized Fishcreekian regressive deposits of the Fish Creek section, apparently because of a misunderstanding of the age significance of the fossil evidence.

The primitive bog lemming *Pliotomys mimomiformis* is present in the Fish Creek fauna near the top of the section (Repenning and others, 1987; Repenning, in press). A more advanced species of this genus (worldwide, the last record of the genus) became extinct in the United States before the two normal-polarity excursions of the Reunion Subchron of the Matuyama Chron (that is, more than 2.14 m.y. ago) (Repenning and Grady, 1988). The Fish Creek record of *P. mimomiformis* in east-central Beringia is the only record of this species outside of Eurasia; the youngest Eurasian record is in the Kutuyakh Beds of the Krestovka sections, Yakutia, in westernmost Beringia (in the same faunal province as Fish Creek), which were deposited during the first cooling period after the polarity change at the end of the Gauss Chron.

In the Krestovka sections of west Beringia, the climatic trend of terrestrial floras is essentially identical in time and nature to that of the Colvillian-Bigbendian-Fishcreekian sequence (Repenning, in press). Although there are no marine deposits in the Krestovka sections, the sequence records the change from open forests (Begunov Suite), to forest-tundra (older Kutuyakh Beds), and to the first permafrost and treeless tundra (younger Kutuyakh Beds, above the Gauss-Matuyama polarity boundary).

In western Beringia, the entire record is present at one locality, and there is less uncertainty than in reconstructing

its history from localities 16 mi (26 km) apart on the Alaskan North Slope. The principal difference between the Alaskan and Yakutian climatic records stems from the much greater (but unknown) distance of the Krestovka sections from the late Pliocene Arctic Ocean and the continentality of the Asian land mass; the compared floras of Yakutia are consistently drier in nature and have a forest-steppe or steppe composition; grass is much more abundant in the pollen record of Yakutia than in the Colvillian-Bigbendian-Fishcreekian climatic succession of Alaska.

As mentioned, the Krestovka record extends a bit later in time than does the Fish Creek section, and it shows the first of the following return of forests (youngest Kutuyakh Beds) about 2.38 million years ago based on the calibration of the oxygen isotope record of sea-level change. The top of the Kutuyakh Beds is marked by an erosional unconformity that is overlain by deposits about 1.5 million years old (Repenning, in press).

The following points summarize the age-significant data pertaining to the Fish Creek section. The transgressional-regressional history of the Bigbendian-Fishcreekian sequence, exposed in two localities 16 mi (26 km) apart, represents one sea-level cycle of the Arctic Ocean tied together by an unbroken continuum in (1) cooling trends of both marine and terrestrial climates, (2) depositional types, (3) amino acid ratios, (4) marine faunas other than ostracodes, and (5) paleomagnetic patterns and by the lack of irregularities in these patterns and trends that would suggest temporal gaps in the record which would eliminate evidence of more than one eustatic sea-level cycle.

This interpretation is supported by an essentially identical record at one locality in Yakutia that is correlated with the Alaskan record by (1) paleomagnetic pattern, (2) unique mammalian faunal elements, (3) very similar flora history, and (4) very similar climatic changes. This Yakutian record lasts briefly into the next climatic (eustatic) cycle, recording the return of warmer climate and of boreal forests; it thus indicates that, if more than one cycle had been involved in the Fish Creek section, the climatic-floral change would have been recognizable, as presumably would the change from regressive to transgressive deposits, as has been mentioned.

Thus, all available evidence suggests that the Fish Creek fauna in the upper part of the Fish Creek section of the Alaskan North Slope is within the regressive sea level indicated between isotope stages 100 and 101 (fig. 2) and is about 2.40 million years old. We see no reason to change our age interpretation of the Fish Creek section as being about 2.4 million years old (Repenning and others, 1987) and see unnecessary complication and additional uncertainty in a younger age interpretation, as suggested by Carter and Hillhouse (1992) and Brigham-Grette and Carter (1992).

Amino acid racemization (alle/le) suggests a significant difference in age between deposits of the Colvillian and Bigbendian transgressions (Brigham-Grette and Carter, 1992). The similar marine ostracode faunas indicate, however, stability and great geographic extent (as far north as Meighen Island) of marine temperatures during the time represented by these transgressions. Also indicated by comparison between Meighen Island and the deposits at Ocean Point is the fact that late Pliocene terrestrial climates, in contrast with shallow-marine climates, differed greatly with latitude, as they do today.

Although the first two marine transgressions of the Alaskan North Slope represent a marine climate similar to that farther north at Meighen Island, they are believed to be younger because they lack some Atlantic mollusks, as indicated by the Bigbendian fauna, that are present in the marine beds of Meighen Island. Subsequent to the Meighen Island fauna, these Atlantic mollusks were no longer present in the Arctic Ocean. The Colvillian and Bigbendian transgressions may span the time from shortly after 3 million years ago until shortly after the end of the Gauss Chron, 2.48 million years ago. As has been noted, the Meighen Island flora represents a treeline environment; the differences between the Colvillian and Bigbendian floras and those from Meighen Island can be explained by the difference in latitude.

At present there is no way of estimating the amount of time between the Colvillian and Bigbendian transgressions, but the similarity of the ostracode faunas and the terrestrial floras during the two events suggests little change in marine and terrestrial temperatures. Differences in the terrestrial flora, including the disappearance of hemlock and reduction in percentages of pine, suggest some lowering of air temperature and possibly a drier climate.

Oxygen isotope ratios from the World Ocean indicate many fluctuations in sea level during the time ascribed to the Colvillian and Bigbendian transgressions (fig. 2), whereas the deposits of these transgressions indicate fewer. Although this relationship may indicate a lack of adequate depositional records on the Alaskan North Slope or inadequate detail of their examination, the Colvillian and Bigbendian records and others in the Arctic Ocean Borderland suggest that shallow-marine deposits and, particularly, terrestrial deposits dampen the sensitivity of the record found in marine isotope ratios; minor eustatic sea-level changes may result in little lithologic change if the locality is not crossed by a shifting strand line, and some of the isotope fluctuations may reflect Antarctic events.

FISHCREEKIAN "TRANSGRESSION"

A third Pliocene marine sequence was informally named the Fishcreekian transgression (Carter and others, 1986). The type section is exposed along Fish Creek, 16 mi

(26 km) northwest of Ocean Point. Its faunal and floral character were discussed in greatest detail by Repenning and others (1987). The Fishcreekian type section is composed of reversely magnetized marginal-marine deposits that contain a fauna of coldwater ostracodes much different than those of the Colvillian and Bigbendian transgressions and essentially identical to the living ostracode fauna of the inner sublittoral Arctic Ocean.

The mollusks in the Fishcreekian type section, however, are distinctly warmer water forms than the modern Arctic Ocean fauna: they represent a cold-temperate to subfrigid marine climate similar to that of the southern Bering Sea and the northern Gulf of Alaska (Marincovich in Repenning and others, 1987). The presence of an archaic sea otter indicates a similar marine climate. The indications of the marine climate overlap in the subfrigid climate range.

The presence of some marine mollusks and an archaic sea otter in the Fish Creek fauna indicates that, as during the Bigbendian transgression, the ocean could not have been frozen to the shore during the winter. Small crystal clusters (about 5 mm in diameter) of glendonite (a calcite pseudomorph after ikaite) in the mollusk- and vertebrate-bearing deposits in the upper part of the section indicate that the water was not much above freezing temperatures (maximum 5°C) (Kennedy and Pickthorn, 1987).

Pollen floras in the Fish Creek section show increasing climatic severity upsection, ranging from shrub tundra, possibly with some larch, in the lower part of the section to herb tundra near the top. Pollen of pine, spruce, and larch are sufficiently abundant in samples from the lower part of section to suggest that the limit of the boreal forest was farther north than today. Decreases in the percentages of these elements upsection suggest a southward withdrawal of distant forests or a change in prevailing wind direction.

As the percentage of arboreal pollen decreases upsection, the pollen spectrum changes from shrub tundra to wet (herb) tundra, suggesting the possible development of permafrost inland from the marginal-marine deposits of the Fish Creek section. This change in floral aspect is accompanied by the appearance of the pollen of yellow water lily, indicating that the area was not as cold as it is today (Ager in Repenning and others, 1987).

The lowest part of the basal silty unit in the Fish Creek section (about the lowest 4 ft (1.2 m)) contains a sparse ostracode fauna, including *Pterygocythereis vannieuwenhusei*. This fauna indicates a markedly warmer ocean than does the overlying Fish Creek fauna. The warmwater ostracode fauna (localities 83EB171 and 83EB172 in Repenning and others, 1987, fig. 3) is comparable to the faunas of the Colvillian and Bigbendian transgressions. The presence of *P. vannieuwenhusei*, an extinct species normally found in association with the extinct *Rabilimis paramirabilis*, indicates a post-Bering Strait age; the warmth it indicates implies a pre-Fishcreekian age.

fir, hemlock, hazel, or linden in 70 percent of the samples (Sher and others, 1979) suggest the existence of more distant, cool-temperate vegetation and enhance the similarity of the Begunov and Colvillian floras. The Colvillian flora also contains rare fir, hemlock, and hazel (Nelson and Carter, 1992).

The Begunov pollen content differs little from the pollen suite in the lower part of the overlying Kutuyakh Beds except for a decrease in the percentage of arboreal pollen, a trend that continues upsection within the Kutuyakh; this difference is comparable to the floral difference between the Colvillian and Bigbendian deposits.

A polarity change from normal to reversed, correlated with the Gauss-Matuyama polarity boundary on the basis of the younger Kutuyakh microtine fossils, is present in the lower part of the Kutuyakh Beds (Sher and others, 1979, p. 63: paleomagnetic studies by E.I. Virina; microtine rodent studies by V.S. Zazhigin). It is difficult to follow the reasoning of the authors (from their description and illustrations), but, if we have correctly interpreted their statements, the polarity boundary is about 1.5 m (5 ft) above the base of the Kutuyakh Beds. Sher and others (1979, p. 63) mentioned that permafrost action in the upper part of the Kutuyakh Beds has distorted paleomagnetic directions so that they are anomalous, and the polarity boundary may fall within this zone of anomaly.

Thus, the age-discriminating fauna and the earliest evidence of permafrost, as well as the transition from open forest to tundra (or steppe) in the Kutuyakh Beds, are younger than the polarity reversal and contemporaneous with the early part of the Matuyama Chron, as is indicated in the Bigbendian and Fishcreekian deposits.

The arboreal pollen content of the younger Kutuyakh Beds decreases upward to essentially none in that part of the section containing most of the fossil microtines. There the "pollen" spectrum contains abundant pond weed (several extinct species) and seeds of yellow water lily (Sher and others, 1979), comparable to the Fish Creek flora of Alaska. The youngest samples from the Kutuyakh Beds, however, show an increased percentage of arboreal pollen of between 7 and 12 percent (Sher and others, 1979). The return of arboreal pollen is unknown in the Fish Creek section (Repenning and others, 1987), probably because no Fish Creek deposits are as young as the youngest Kutuyakh Beds.

Paleomagnetic determinations for the Colvillian and Bigbendian transgressions and the Fishcreekian regression suggest that these deposits and the Begunov-Kutuyakh sequence of the Krestovka sections are of essentially identical age. Both the Alaskan and Yakutian sections were deposited during the Gauss and earliest Matuyama Chrons, have similar pollen records and trends in floral composition, and are at almost the same latitude. Both sections contain a distinctive bog lemming (*Pliotomys mimomiformis*) in the same part of the paleomagnetic and paleoclimatic

record, and both are in the Beringian faunal province. The Kutuyakh Beds record the return of forests and are believed to include slightly younger deposits than the Fish Creek section in Alaska, about 2.38 million years old (fig. 2).

The lesser percentage of arboreal pollen in the Begunov-Kutuyakh sequence indicates a somewhat more severe or more arid climate during the last of the Gauss Chron than that recorded for Bigbendian deposits, although both sections show a continuous decrease in arboreal flora upsection. Some hemlock is present in the Colvillian flora but apparently absent in the Bigbendian flora; it is very rare in the Begunov records and absent in the Kutuyakh flora. Considering the continentality of the Asian climate and the greater distance of the Krestovka sections from the ocean (Fish Creek section, 23 mi (37 km); Krestovka sections, 200 km (124 mi) today and possibly much farther 2.5 m.y. ago), the similarity is rather remarkable.

The Colvillian and Bigbendian floras, before the end of the Gauss Chron, have a greater percentage of arboreal pollen at the beginning of their record and do not show as great a decrease upsection as the Begunov and Kutuyakh sequence, but a decrease is indicated (Nelson and Carter, 1992), and the nearby and slightly younger Fish Creek section, which records full and extensive tundra during the time of the early Matuyama Chron, certainly indicates that this decrease in arboreal pollen continued. The two Alaskan localities have a less complete pollen sampling throughout the section, and the paleomagnetic record of the Krestovka sections is ambiguous at the time of the polarity change. These differences frustrate comparisons of the respective floras at the only identifiable point of contemporaneity, the Gauss-Matuyama boundary, but as far as comparisons can be made the timing and nature of floral changes are similar in both areas.

Within the perspective of rapidly decreasing forests in the Arctic Ocean Borderland, the similarities in pollen content suggest that the Begunov Suite is only slightly older than the Kutuyakh Beds. The Begunov Suite had been considered to be early Pliocene in age because of its floral similarity to late Miocene floras of Siberia (Sher and others, 1979). Based on the above relationships, the Begunov Suite more likely was deposited during the late part of the Gauss Chron, slightly before the Kutuyakh Beds and after the opening of the Bering Strait. No reversed magnetic polarity has been reported in either Alaska or the Yakutia, a polarity that would suggest that the Begunov or the Colvillian transgression was as old as the Kaena Subchron of the Gauss Chron, ending 2.92 million years ago.

Permafrost is not indicated in the sediments of the Colvillian or Bigbendian deposits or the Begunov Suite; the spruce-birch forest in both areas suggest a lack of permafrost. Evidence of permafrost was first reported by Sher and others (1979) in the youngest part of the Kutuyakh Beds, upsection from the Gauss-Matuyama polarity reversal. Permafrost may first have appeared in northern Alaska

fir, hemlock, hazel, or linden in 70 percent of the samples (Sher and others, 1979) suggest the existence of more distant, cool-temperate vegetation and enhance the similarity of the Begunov and Colvillian floras. The Colvillian flora also contains rare fir, hemlock, and hazel (Nelson and Carter, 1992).

The Begunov pollen content differs little from the pollen suite in the lower part of the overlying Kutuyakh Beds except for a decrease in the percentage of arboreal pollen, a trend that continues upsection within the Kutuyakh; this difference is comparable to the floral difference between the Colvillian and Bigbendian deposits.

A polarity change from normal to reversed, correlated with the Gauss-Matuyama polarity boundary on the basis of the younger Kutuyakh microtine fossils, is present in the lower part of the Kutuyakh Beds (Sher and others, 1979, p. 63: paleomagnetic studies by E.I. Virina; microtine rodent studies by V.S. Zazhigin). It is difficult to follow the reasoning of the authors (from their description and illustrations), but, if we have correctly interpreted their statements, the polarity boundary is about 1.5 m (5 ft) above the base of the Kutuyakh Beds. Sher and others (1979, p. 63) mentioned that permafrost action in the upper part of the Kutuyakh Beds has distorted paleomagnetic directions so that they are anomalous, and the polarity boundary may fall within this zone of anomaly.

Thus, the age-discriminating fauna and the earliest evidence of permafrost, as well as the transition from open forest to tundra (or steppe) in the Kutuyakh Beds, are younger than the polarity reversal and contemporaneous with the early part of the Matuyama Chron, as is indicated in the Bigbendian and Fishcreekian deposits.

The arboreal pollen content of the younger Kutuyakh Beds decreases upward to essentially none in that part of the section containing most of the fossil microtines. There the "pollen" spectrum contains abundant pond weed (several extinct species) and seeds of yellow water lily (Sher and others, 1979), comparable to the Fish Creek flora of Alaska. The youngest samples from the Kutuyakh Beds, however, show an increased percentage of arboreal pollen of between 7 and 12 percent (Sher and others, 1979). The return of arboreal pollen is unknown in the Fish Creek section (Repenning and others, 1987), probably because no Fish Creek deposits are as young as the youngest Kutuyakh Beds.

Paleomagnetic determinations for the Colvillian and Bigbendian transgressions and the Fishcreekian regression suggest that these deposits and the Begunov-Kutuyakh sequence of the Krestovka sections are of essentially identical age. Both the Alaskan and Yakutian sections were deposited during the Gauss and earliest Matuyama Chrons, have similar pollen records and trends in floral composition, and are at almost the same latitude. Both sections contain a distinctive bog lemming (*Pliotomys mimomiformis*) in the same part of the paleomagnetic and paleoclimatic

record, and both are in the Beringian faunal province. The Kutuyakh Beds record the return of forests and are believed to include slightly younger deposits than the Fish Creek section in Alaska, about 2.38 million years old (fig. 2).

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correlates exactly with the "pre-Nebraskan till" in Iowa, which includes the 2.14-million-year-old beginning of the Reunion Subchron (Easterbrook and Boellstorff, 1984), and with oxygen stage 82 in oceanic records (fig. 2).

At the eastern end of the Alpha Ridge, north of Ellesmere Island, Scott and others (1989) reported the intermittent presence of calcareous benthic foraminifers in CESAR core 14 between the Gauss-Matuyama boundary and just before the first Reunion normal-polarity excursion. These foraminifers were interpreted as indicating perennial ice cover. In that the Fish Creek fauna indicates that ice cover did not extend to the shore as recently as 2.4 million years ago, the presence of these calcareous foraminifers in CESAR core 14 suggests that perennial ice formed on the central Arctic Ocean long before it extended to the shores in the winter. The appearance of these foraminifers also suggests that the intermittent presence of perennial ice cover began shortly after initial activity of the Scandinavian ice sheet.

Following the first period of glacial activity, a quiet period in Scandinavian ice activity (correlative with a lack of ice cover in the central Arctic Ocean at CESAR site 14) is indicated at site 644 immediately following the Reunion Subchron between 2.0 and about 1.7 million years ago (ending during the late part of the Olduvai Subchron). During this period, Scandinavian ice-debris percentages are at near-background levels; two minor exceptions of increased debris are at the beginning of the Olduvai Subzone (Jansen and others, 1988, fig. 8) (fig. 2).

Before the Olduvai Subchron the site 644 record of no ice activity correlates closely with a time of warm climate from northern Greenland (Funder and others, 1985). During the first of the two minor peaks of rafted debris recording the brief break in warmth at the beginning of Olduvai Subchron, the microtine genus *Allophaiomys* dispersed southward from Beringia into the United States. *Allophaiomys* is also thought to have entered Beringia from lower latitude Asia at this time; it was a rodent that lived in temperate to subtropical climates and was not likely to have entered Beringia when arctic climate prevailed during maximum Northern Hemisphere glaciation (Repenning, in press). In North America, *Allophaiomys* is first found at, or just below, the base of the Olduvai Subzone in Colorado, immediately preceding in time the brief increases in ice-rafted debris in the Norwegian Sea. The genus appears in Holland (from the Black Sea area to the south) immediately after this minor peak of ice-rafted debris in the Norwegian Sea record, first occurring in Holland with late Tiglian warm floras in the older part of the Olduvai Subzone (van Kolfschoten, 1990a).

The end of the period of minor to no ice activity of the Scandinavian ice sheet correlates well with the beginning of the Eburonian cold floral age of Holland (beginning about 1.7 m.y. ago late in the Olduvai Subchron) (van Kolfschoten, 1990a). At this time, the microtine genus

Phenacomys (first known from the 2.4-million-year-old Fish Creek fauna of central Beringia) dispersed southward into the United States, where its oldest record is below the top of the Olduvai Subzone (about 1.7 m.y. ago) in Idaho (Repenning, in press). These ages are approximations based on the relative positions of the increase in ice-rafted debris of the Norwegian Sea, of the oldest Eburonian flora in Holland, and of *Phenacomys* in Idaho within the Olduvai Subzone.

The record of ice-rafted debris between the beginning of the Eburonian and 1.2 million years ago is relatively minor at site 644 (Jansen and others, 1988). Only a few modest peaks are concentrated in the older part of this core segment, between 1.7 and 1.5 (or perhaps 1.6) million years ago (fig. 2), and these are separated by longer periods of near-background percentages. These peaks would seem to correlate with the Eburonian cold flora of Holland, although the termination of this floral age is not dated. The remaining period, between 1.5 (or perhaps 1.6) and 1.2 million years ago, has lower peaks and longer periods of near-background percentages of debris; it probable correlates with the Waalian warm floral zone of Holland, although again this floral zone is not dated except as following the Eburonian and between the Olduvai and Jaramillo events.

Jansen and others (1988) noted that, during this quiet period, the marine climate was probably more latitudinally zoned: the water of the Norwegian Sea was more stable and less mixed with North Atlantic water than at present, and the thermal gradient from this sea to the North Atlantic was much steeper. The southward restriction of the North Atlantic surface current pattern and reduced flow to the Atlantic of Arctic water may have accentuated the warmth of the Waalian floral age in Holland and may have chilled the surface of the Arctic Ocean, permitting intermittent perennial ice to develop for the second time on the central Arctic Ocean.

At about 1.2 million years ago the ice-rafted debris percentage peaks became greater in core at site 644 and periods of near-background percentages became very brief. Although poorly dated by interpolation between the Olduvai and Jaramillo Subchrons, this activity probably correlates with the beginning of the Menapian cold flora age of the Netherlands. As reviewed by Jansen and others (1988), these peaks in ice-rafted debris persisted until 600,000 years ago, after which very few of the low percentage points are at near-background levels and none persisted for a recordable length of time; despite evidence from more closely spaced samples, there were only very brief times of low glacial activity.

In Holland, Zagwijn and de Jong (1984) introduced the term Bavelian for a "warm" floral age between the Menapian and the Cromerian (the first glacial advance of the Ice Ages, not well marked by extensive ice records in Europe). The Bavelian includes, from the base upward, an

"inter-glacial" (warmer flora), a "glacial" (colder flora), a second "inter-glacial," and a second "glacial."

The Bavelian began just before the Jaramillo Subchron and ended shortly after it. Although not appearing very "warm" in the ice-rafted debris record of the core from site 644 or in the immigrant microtines of Holland (which, according to van Kolfschoten (1990b), include the first *Microtus*, a cool climate indicator), the climatically alternating floral events are represented in the debris percentages of core at site 644 in proper relationship to the Jaramillo Subchron.

The difference in prominence of Jaramillo warm climates between the floras of Holland and the Scandinavian ice-rafted debris of the Norwegian Sea could have resulted from the exaggerated temperature gradient between the Norwegian Sea and the North Atlantic. This steep temperature gradient did not begin to diminish until after the Jaramillo Subchron (Jansen and others, 1988, fig. 9). In addition, the difference between the warm Bavelian flora of Holland and the prominence of ice-rafted debris in the Norwegian Sea may actually represent a correlation between warmer climate and ice accumulation (warm climate in Holland=warm North Atlantic=greater atmospheric moisture=greater precipitation on the Scandinavian and Laurentian ice sheets).

The end of the Bavelian was the beginning of the Ice Ages and is marked by the first of many prominent peaks in ice-rafted debris in the core of site 644 at about 850,000 years, although the age indicated by Jansen and others (1988) is slightly different because they used a different age for the beginning of the Brunhes Chron (730,000 years ago; we use 790,000 years).

At this time the Cromerian of northwestern Europe began, the microtine genera *Lasiopodomys* and *Terricola* dispersed from Beringia into the United States (Repenning, in press), and the first of several ice advances called "Nebraskan" in North America is recorded in the Mississippi River drainage basin. This time also correlates with oxygen stage 22 in records of the World Ocean but is not evident in the isotope record of site 644 because of the lack of full mixing with Atlantic waters and because of the effects of carbonate dissolution in the Norwegian Sea and Arctic Ocean.

The history of the Scandinavian ice sheet, as revealed in core 644 on the Voring Plateau of the Norwegian Sea, correlates, in varying degrees of precision, with the history of the two cooling cycles of the Arctic Ocean Borderland. Climatic events recorded in the Colvillian-Bigbendian-Fishcreekian sequence of Alaska, the Begunov-Kutuyakh sequence of Yakutia, and southward microtine dispersals from Beringia into the United States correlate exactly, to the extent that the ages of events shown at site 644 can be interpolated.

The missing history in the borderland records between about 2.38 and 2.0 million years ago is one of repeated ice

peaks that reached a maximum in frequency and degree at about 2.1 million years ago. At 2.14–2.12 million years ago, the first normal-polarity excursion of the Reunion Subchron was recorded in glacial till in Iowa and adjacent regions of the United States. Before this excursion, bottom sediments on the eastern part of the Alpha Ridge provide evidence indicating the first development of perennial ice cover on the central Arctic Ocean.

Just after the second Reunion normal-polarity excursion, about 2.0 million years ago, continental glaciation in Iowa was lacking and the climate in northernmost Greenland was extremely warm (Funder and others, 1985); the first cooling cycle of the Arctic Ocean Borderland was at an end.

The last break in the pattern of ice-rafted debris at site 644 noted by Jansen and others (1988) was at about 450,000 years ago, which is when the microtine *Microtus* dispersed southward into the United States. *Microtus* had dispersed as far south as southern Canada much earlier (Repenning, in press). Scott and others (1989) suggested that permanent perennial ice cover did not develop on the Arctic Ocean until about 300,000 years ago on the basis of their correlation of isotope stages. Due to very slow depositional rates (1 mm/1,000 yr) and the prevalence of samples too small for isotope measurement, their isotope stage identifications are questionable. This time period is later than the events discussed in this report.

CENTRAL ARCTIC OCEAN: EASTERN ALPHA RIDGE

Scott and others (1989), in their study of cores taken by the Canadian Expedition to Study the Alpha Ridge (CESAR), reported an abrupt replacement of agglutinated benthic foraminifers by calcareous benthic foraminifers at about the Gauss-Matuyama boundary. This replacement ended about 2.15 million years ago, before the older excursion of the Reunion Subchron as recorded in CESAR core 14 on the eastern end of the Alpha Ridge of the central Arctic Ocean.

This replacement is in the upper part of lithostratigraphic unit AB of Mudie and Blasco (1985) and consists of about three peaks of calcareous benthic foraminifers corresponding, in a general way, with the pattern of oxygen isotope fluctuations in the World Ocean between 2.48 and 2.15 million years ago. Scott and others (1989) inferred that this change represents the first appearance of perennial ice cover of the Arctic Ocean. If so, one might assume that some of the lower sea-level peaks of the oxygen isotope records in the World Ocean would correlate with ice cover of the central part of the Arctic Ocean following the Fish Creek-Kutuyakh deposition in Beringia. With the exception of isotope stage 96, about 2.4 million years ago, these

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Arctic Ocean was at a high eustatic level and covered much of the Alaskan North Slope. Treeline was at lat 79°30' N., farther north than most of the shoreline of the Arctic Ocean Borderland, and there is no indication of ice cover in western or central Arctic Ocean cores. Spruce-birch forests along the coast were more diverse than today and contained both extinct taxa and taxa that now live only far to the south.

In temperate latitudes this time equates with the Rebielice small-mammal age of Europe, the Reuverian floral age of northwestern Europe, the late MN 16 zone of the Mediterranean region, the late Kujalnikian north of the Black Sea (Tatiana Svetitskaya, written commun., 1992), the early part of the Kisichian in Western Siberia, the Itantsian in western Transbaikalia, Chinese zone III as correlated by Zheng and Li (1990), and the Blancan IV microtine age of the United States (Repenning and others, 1990). Each of these areas had its own local climate, but all of the areas were climatically related to the Arctic Ocean and, in general, indicate warmth; in some cases, decreasing warmth is indicated.

Toward the end of this time (toward the end of the Gauss Chron, 2.48 m.y. ago) the forests were thinning in Beringia to mosaics of forest-tundra or, in Asia, forest-steppe because grasses were more common in western Beringia than in the east. The Scandinavian ice sheet began to develop about 2.56 million years ago; at this time the Arctic Ocean became density stratified as eustatic sea level began to lower, although much of the Alaskan North Slope was still submerged. About 100,000 years after increased activity of the Scandinavian ice sheet was evident in the Norwegian Sea, evidence of ice-transported debris on the Arctic Ocean floor was more abundant and intermittent perennial ice cover began to form on the central Arctic Ocean; for at least 60,000 years, however, ice cover did not extend to the coast in the winter.

The southward movement of some microtine rodents (bog lemmings) from Beringia to the United States about the time that Scandinavian ice growth became apparent (the arrival in lower latitudes of other northern mammals is not demonstrated at this time) suggests that increasingly cold weather extended far south in North America. The bog lemmings of today still inhabit northern forests of Canada and the northeastern United States.

By perhaps 2.45 million years ago, shortly after the end of the Gauss Chron and the beginning of the Matuyama Chron 2.48 million years ago, treeline had shifted south of the average latitude of Beringia (70° N.), and full tundra was associated with permafrost in both eastern and western Beringia. About this time the ostracode fauna of the Arctic Ocean changed with dramatic abruptness to an assemblage characteristic of subfrigid or frigid temperatures, but the mollusk fauna changed little;

in combination, the faunal evidence indicates a subfrigid marine climate that was still warmer than today but colder than earlier records. The shoreline of the Arctic Ocean had receded northward of lat 70° N. in Alaskan Beringia with global eustatic lowering of sea level. By 2.4 million years ago, ice cover had not yet developed on the Arctic Ocean to the extent of reaching the Alaskan coast, and marine life that was dependent on year-round ice-free water still survived. Intermittent perennial ice cover is believed, however, to have been present in the central part of the Arctic Ocean, and mountain glaciation and the Scandinavian ice sheet were contributing icebergs to the Arctic Ocean as well as to the Norwegian Sea from where they drifted into and around the Arctic Ocean.

The remainder of the first cooling cycle of the Arctic Ocean region is not represented by deposits on land, possibly a result of lower sea levels that encouraged continental erosion rather than deposition. This last chapter of the first cooling period, during the earliest part of the Matuyama Chron, is recorded in the Norwegian Sea where activity of the Scandinavian ice sheet is indicated by ice-rafted debris. Intermittent increases in glacial activity climaxed at about 2.1 million years ago, when continental ice deposited till as far south as Iowa in the United States.

In temperate latitudes the later part of the first cooling cycle equates with the European Villanyian age, the Praetiglian floral age of northwestern Europe, zone MN 17 of the Mediterranean area, the late part of the Kisichian of Western Siberia, the earlier Dodogolan of western Transbaikalia, Chinese zone IV as correlated by Zheng and Li (1990), and the Blancan V' microtine age of the United States. During this time continental ice flowed south down the Mississippi River Valley to Iowa, and Blancan V microtine faunas both east and west of the Rocky Mountains are marked by the presence of bog lemmings (*Pliotomys* and *Mictomys*). In eastern Asia the ancestor (*Allophaiomys*) of the temperate to arctic microtine voles with rootless cheek teeth (*Microtus*, *Pitymys*, *Terricola*) evolved out of a subtropical to temperate late Pliocene form with rooted cheek teeth (*Miomys*). In North America the primitive, ubiquitous archaeolagine genus *Hypolaigus* was replaced by the modern, ubiquitous leporid *Lepus*.

At the close of the first cooling cycle of the Arctic Ocean Borderland, treeline was well south of its modern limits in both eastern and western Beringia, and continental ice had spread southward into the United States. In the brief time interval of about 100,000 years, by 2.0 million years ago, a treeline flora and an absence of Arctic Ocean ice cover is known from the northern end of Greenland at lat 82°30' N. (Funder and others, 1985). This climatic warmth was greater than that known at the beginning of the first cooling cycle, about 3 million years ago, and it marked the beginning the second cooling cycle.

2.0-0.85 MILLION YEARS AGO: THE SECOND COOLING CYCLE

Records from Scandinavia (Jansen and others, 1988), the central United States (Easterbrook and Boellstorff, 1984), and Greenland (Funder and others, 1985) indicate that, by 2.0 million years ago, the late Pliocene continental glaciation of the Northern Hemisphere had ended and the Arctic Ocean and its borderland were much warmer. The ostracode faunas from the Kap København Formation of northernmost Greenland at lat 82°30' N. indicate conspicuously warm waters in the Arctic Ocean and a treeline flora on land about 2.0 million years ago; it was much warmer than at lat 70°16' N. about 2.4 million years ago, when full tundra covered the Arctic Ocean Borderland, or than the modern Arctic Ocean (Brouwers and others, 1991).

To judge from the records on Meighen Island, the Kap København fossil flora record indicates an Arctic Ocean climate warmer than that which existed at the opening of the Bering Strait a million years earlier. No perennial ice cover is suggested at any location in the Arctic Ocean at this time, and dispersing microtine rodents entering North America from Asia indicate that at the lower latitudes of Beringia a temperate climate was present 2.0 million years ago.

In the conterminous United States winter freezes were rare across the northernmost Great Plains and in at least the southern half of inland California. To the east of the Rocky Mountains, on the northern Great Plains, summer rains were common, summer humidity was high, and maximum summer temperatures were lower than today (Repenning, 1990). Northwestern Europe was characterized by the warm Tiglian flora.

As the second cooling cycle progressed, Scandinavian glaciation again reached maximum activity between 1.2 and 1.1 million years ago, as indicated in the core at site 644, just before a normal-polarity excursion called the Cobb Mountain Normal-polarity Subchron (1.10 m.y. ago) (fig. 2). At this time full tundra returned to the Kolyma Lowlands of Yakutia, and, if paleomagnetic correlations of CESAR core 14 are as suggested by Repenning (in press), perennial ice cover was intermittent in the central Arctic Ocean.

Between 2.0 and 1.2 million years ago the nonmarine climate of the Arctic region vacillated between cool and warm periods, beginning late during the Olduvai Subchron (about 1.7 m.y. ago) at site 644 with the return of moderate peaks of ice-rafted debris from the Scandinavian ice sheet.

There are no records of the ecosystem of the Arctic Ocean Borderland during the Olduvai Subchron, and the history must be inferred from records of the Arctic Ocean and from lower latitude proxy records. The next record younger than the initial warm period 2.0 million years ago is from reversely polarized deposits before the Jaramillo

Subchron in Yakutia (Virina and others, 1984) and from the Cape Deceit Formation on the Seward Peninsula of Alaska that may be 1.5 million years old, deposited shortly following the Olduvai Subchron. The age of these deposits is inferred primarily from an incompletely known history of the microtine rodents (Repenning, in press). Based on these microtine records, the climatic record in the Arctic Ocean Borderland can be confidently, but approximately, correlated with more complete oceanic records and with continental records in lower latitudes.

During the Jaramillo Subchron (970,000-900,000 y.b.p.) of the Matuyama Chron, two relatively brief warm periods interrupted this second cooling trend, as indicated in Holland, in Yukon Territory of Canada, and in Yakutia. These two warm periods are not clearly detectable in Norwegian Sea or Arctic Ocean records, for uncertain reasons. Following the warm periods, the cooling continued, and the Ice Ages began 850,000 years ago with the advance of continental glaciation down the Mississippi River Valley in the United States. This first ice advance, along with several subsequent ones, has been referred to as the "Nebraskan" glaciation.

The only presumed record of the ecosystem in the Arctic Ocean Borderland at the close of the second period of climatic deterioration and the beginning of the Ice Ages is in the type section of the Olyor Suite in Yakutia, and it has not been studied in detail. The Yakutian record may not be complete, but it does contain reversely polarized deposits younger than the Jaramillo Subchron, a situation possibly unique in the Arctic Ocean Borderland.

Scott and others (1989) suggested that permanent perennial ice cover of the Arctic Ocean was not established until perhaps 300,000 years ago and that before this time perennial ice cover of the Arctic Ocean was intermittent. Intermittent perennial ice cover began just before an 8-cm-thick normal-polarity event in CESAR core 14, here believed to record the Cobb Mountain event.

GREENLAND: KAP KØBENHAVN FORMATION

Following the first Northern Hemisphere glacial maximum 2.1 million years ago, a spectacular return of forests to the shores of the Arctic Ocean is indicated in the record of the Kap København Formation of northernmost Greenland (Funder and others, 1985; Bennike, 1990), which is the northernmost terrestrial fossil record in the world. This warming is also recorded in (1) the abrupt reduction of activity of the Scandinavian ice sheet; (2) lower latitude North American faunas that indicate subtropical climates near the Canadian border (Repenning, 1990); (3) the St. Erth Beds of Cornwall, England, that indicate subtropical temperatures for the North Atlantic (Jenkins and Houghton, 1987); (4) the Tiglian warm floras of northwestern Europe;

and (5) immigration of the subtropical to warm-temperate microtine rodent to the United States.

Funder and others (1985) described a sequence of shallow-marine and marginal-marine deposits at Kap København in the northernmost part of Greenland, at lat 82°30' N. and about 525 mi (845 km) south of the North Pole (fig. 1). The older of two members of the Kap København Formation, member A (fig. 2), records shallow glaciomarine deposition, and the younger, member B, is composed of high-energy marginal-marine and coastal deposits. Member B, representing as much as three-quarters of the thickness of the formation, is subdivided into three units, referred to as B1, B2, and B3, in order of deposition.

The oldest quarter, member A, contains a modern Arctic Ocean mollusk fauna, with no extinct species, and dropstones thought to indicate nearby glaciation (Bennike, 1990). Deposition was in the absence of wave action presumably because of extensive ice cover. The possibility of ice cover suggested in CESAR core 14, about 550 mi (886 km) to the west-northwest, began about 2.48 million years ago and ended about 2.15 million years ago (Scott and others, 1989).

The mollusk fauna of member A is the oldest known Arctic Ocean mollusk fauna containing no extinct taxa; the Fish Creek fauna (2.4 m.y. old) contains one extinct form (Marincovich *in* Repenning and others, 1987). The fauna and the glaciomarine environment of member A indicate that it probably is younger than the Fish Creek fauna and is equivalent to unit II in the western Arctic Ocean (Herman and others, 1989) and to unit AB of CESAR core 14 (Scott and others, 1989).

Fossils and sedimentary structures of member B indicate that the Arctic Ocean could not have been perennially ice covered during deposition. Unit B3 contains evidence of a treeline biota with a mixture of boreal trees (trunks as much as 7.1 in (18 cm) in diameter and 15 ft (4.5 m) long) and tundra shrubs with a preponderance of forest beetles (with lesser numbers of tundra beetles) and temperate to low arctic coastal moss (Bennike, 1990; Bennike and Bocher, 1990). This biota indicates that the land adjacent to the ice-free ocean had a climate similar to that of today 27° farther south (1,900 mi, 3,000 km) (Bennike, 1990). An upsection warming trend is indicated by these fossil records in the Kap København Formation. Limited paleomagnetic study of member B indicates reversed polarity and that the section is not young enough to include the Olduvai Normal-polarity Subchron (Abrahamsen and Marcussen, 1986). The younger of the two events of the Reunion Subchron (ending 2.01 m.y. ago) may be present in member B, to judge from the paleomagnetic results of Abrahamsen and Marcussen (1986).

Unit B3 (the youngest) is mostly composed of marginal-marine deposits. An ostracode fauna in its basal part (sample 11514, Geological Museum of Copenhagen

University, locality 68) contains warmwater forms (*Cytheromorpha macchesneyi*, *Cytheromorpha* sp., *Leptocythere* spp., and *Bythocythere* sp.). This fauna is comparable to faunas from cold-temperate to warm-subfrigid marine provinces such as modern coastal Nova Scotia (Brouwers and others, 1991).

Underlying unit B3, unit B2 is a marine silt containing mollusks now living in warmer Atlantic waters (O. Bennike, written commun., 1987). Ostracodes in unit B2 indicate subfrigid water temperatures, which today are found in the Atlantic Ocean off of the coast of Norway and Newfoundland. The ostracode fauna indicates inner to middle shelf depths.

The oldest unit in member B, unit B1, is a coastal deposit that contains a flora similar to that of unit B3 but with no trees. Except for the lack of arboreal plants, the dwarf shrub heath present in unit B1 is similar to that of unit B3, and a subarctic climate near treeline is inferred (O. Bennike, written commun., 1988). Ostracodes were recovered from only one sample from unit B1 (sample 11494, Geological Museum of Copenhagen University, locality 74); the assemblage represents an inner shelf environment and is similar to the ostracode fauna from unit B2.

Unit B3 (the youngest) contains an unusual association of modern and primitive rabbits; the Pliocene archaeolagine *Hypolagus* and the Pleistocene (and living) leporine *Lepus* are present. *Lepus* was not mentioned by Funder and others (1985), but its provenance was confirmed by O. Bennike (oral commun., 1988). Only three other faunas in North America (Anita, Arizona; Borchers, Kansas; Big Springs, Nebraska) have this unusual association; two of these localities are 2.01 million years old, and the Anita fauna is not dated. *Hypolagus* is common in most North American localities older than about 2.0 million years, but *Lepus* is unknown, whereas *Lepus* is common in most fossil localities younger than 2.0 million years, but *Hypolagus* is unknown, these time relations suggest that unit B3 is close to 2.0 million years old.

The relative warmth recorded in member B suggests that it postdates the time of till deposition in Iowa, peak glacial growth in the Scandinavian ice sheet, and the earliest period of ice cover on the central Arctic Ocean (all ending 2.1 m.y. ago). Because the normal polarity of the Olduvai event was not recorded in the paleomagnetic study of the Kap København Formation, deposition must predate 1.90 million years ago. The unusual association of rabbits confirms the age range of 2.1-1.9 million years and further suggests that unit B3 is constrained to an age of about 2.0 million years.

The progressive warming upsection indicates that the younger part of the Kap København Formation records the transition from the first glaciation into a warm period; that is, the transition from the first cooling cycle into the second. The warming is also recorded in Holland, the Norwegian Sea, Great Britain, and the central United States as

described in the beginning of this section. Its cause may relate to northward displacement of marine climate in the North Atlantic Ocean (Repenning, 1990), but, if so, a similar situation existed in the North Pacific, as indicated by the dispersal of microtine rodents, to be discussed.

YAKUTIA: KRESTOVKA OLYOR SUITE

In the Kolyma Lowlands of Yakutia, deposits of the Krestovka sections that are between 2.0 and 0.85 million years old are placed in the Olyor Suite of Sher (1971), and this record does not begin until about 1.4 million years ago. These deposits are reversely polarized in their older parts and normally polarized in their younger parts. The contained fauna of large mammals is a typical Ice Ages fauna (Sher, 1971) and was initially considered to represent the beginning of the Ice Ages; the Krestovka Olyor Suite was believed to represent deposition during the latest part of the Matuyama Reversed-polarity Chron and the earliest part of the Brunhes Normal-polarity Chron, perhaps 0.85–0.7 million years ago (Sher and others, 1979; Sher, 1987) although there was no control of the younger time limit.

Subsequently, Azzaroli (1983) indicated that the Olyor large mammal fauna was representative of his Galerian (early Ice Age) large mammal fauna, which dispersed throughout Eurasia during what he called the End-Villafranchian dispersal event.

The only lapse of Azzaroli (1983) was not to consider the possible age of the source of the Galerian fauna, which he recognized as being the Arctic Ocean Borderland. The age problem came into focus in 1984, when Virina and others published their paleomagnetic study of the type Olyor Suite. They indicated that the Olyor (Galerian) fauna was known before the Jaramillo Subchron, whereas Azzaroli believed that his End-Villafranchian dispersal event, presumed to initiate the Galerian fauna throughout Eurasia, took place during or slightly after the Jaramillo Subchron. Recognition of the Jaramillo Subzone in the type Olyor Suite indicated that, in its source area, the Galerian fauna was older than the dispersal event.

The microtine fauna of the Olyor Suite of the Krestovka sections had not been described in 1984, when Repenning, misled by differences in the use of taxonomic nomenclature in published faunal lists, indicated that the microtine fauna was much older than had been inferred in publications; the only critical microtine fossils from the faunas of the Olyor Suite illustrated in publication are those in Repenning (in press); attention is called to the differences in nomenclature in this report. On the assumption that he understood the meaning of the taxonomic names used, Repenning (1984) indicated that the climatic history, magnetic polarity pattern, and microtine history reported for the unit strongly suggested that the Olyor Suite in the Krestovka sections was between 2.4 and 2.0 million years old, representing the earliest part of the Matuyama Chron,

during and before the Reunion Subchron. This opinion was restated in Repenning and others (1987) and has been cited in other publications.

Unfortunately, the climatic history and magnetic polarity pattern present in the Olyor Suite are similar to those during both the first and the second cooling cycles of the Arctic Ocean Borderland, and correct identification of the microtine rodent fossils was the only key to separate these two similar segments of the history of the Arctic Ocean Borderland. Both interpretations were based upon reversed polarity in the older part of the section and normal polarity in the upper part. Both interpretations recognized a cooling trend with major glaciation and the coldest climate at the beginning of the normal polarity. The biochronologic significance of the microtine fossil called *Allophaiomys* was obscured by different definitions.

Agreement between A.V. Sher (the principal investigator in the study of the Krestovka sections) and Repenning on the interpreted age of the Olyor Suite could not be achieved by correspondence. In 1990 Sher invited Repenning to visit Moscow and review the undescribed fossil faunas. Major differences in the application of systematic nomenclature were thus revealed and corrected. The revisions in the nomenclature and in the interpretation of the age of the Olyor Suite of the Krestovka sections are explained in Repenning (in press).

The significant microtine faunas from the Olyor Suite in the Krestovka sections are now recognized as having lived during the Matuyama Chron before the Jaramillo Subchron (Repenning, in press), rather than before the Olduvai Subchron, as published faunal lists had suggested to Repenning (1984). The approximate age of the oldest part of the formation is now known to be about 1.4 million years, based on the dated records in the United States of lineages derived from Asia; the oldest part of the formation cannot be as old as the 1.67-million-year-old termination of the Olduvai Normal-polarity Subchron because normal polarity of this event is not recognized in the sections. The youngest, temporally significant fauna lived immediately before the Cobb Mountain Normal-polarity Subchron (1.10 m.y. old) (Repenning, in press). These age limits apply to the older member of the Olyor in the Krestovka sections (called Olyor IIIa); the younger member (Olyor IIIb) is normally polarized (a factor in the original age interpretation) and may or may not include the Jaramillo Subchron, as well as part of the Brunhes Chron. Sample sizes of the microtine faunas from the younger member (IIIb) are small and do not permit close age approximations.

The revised age interpretation is in satisfactory agreement with the inferred age for the type section of the Olyor Suite, located some 93 mi (150 km) north of the Krestovka sections. The type Olyor was deposited before, during, and after an event of normal polarity assigned to the Jaramillo Subchron (Virina and others, 1984; Sher, 1987). In a

second section in the same area, the type Olyor extends later, into the older part of the Brunhes Chron.

Near the base of the Olyor Suite of the Krestovka sections, the proportion of shrub and arboreal pollen is about 40 percent of the pollen spectrum (Giterman and others, 1982), suggesting a flora similar to, but somewhat more diverse than, the boreal forest at the site today. Upsection in the upper part of the lower member (IIIa) of the Krestovka Olyor, just below and within deposits having normal polarity, the proportion of shrub and arboreal pollen is only 5 percent (Giterman and others, 1982); arboreal pollen is "almost absent" (Sher and others, 1979). Tundra, possibly near treeline, is indicated. This is the most severe environment recorded in the Krestovka Olyor and is more severe than that represented by the taiga at the locality today. Permafrost was present during deposition of Olyor IIIa (Sher and others, 1979).

An inconsistency is evident when this decrease in shrub and arboreal pollen in the Krestovka sections is compared with the pollen record of the type Olyor Suite (outcrop 21 of Sher, 1971). Unlike the Krestovka sections, the type section of the Olyor Suite is in tundra today.

Virina and others (1984) reported that the Jaramillo Subchron is in the middle of the type Olyor section. Yet, in the type section, Giterman and others (1982) reported that the percentage of shrub and arboreal pollen is no lower than 20 percent of the spectrum; in most of the section it is well above this value. The type section contains a pollen record of open forests and tundra, possibly near treeline, in an area now occupied by extensive tundra. It does not contain evidence of a cold period more severe than that indicated by the modern flora.

The absence of evidence for a severe cold period during the Jaramillo in the type section, similar to that recorded in and just below the oldest normally polarized deposits of the Krestovka sections, casts doubt on correlation of the Jaramillo event to the Krestovka sections. Faunal considerations support this doubt (Repenning, in press).

As noted by Sher and others (1979), Sher (1987), and Repenning (in press), the youngest significant microtine fauna in the Krestovka Olyor, locality KLO-9 (0.8 ft (0.25 m) below the normally polarized deposits), contains several specimens of the genus *Microtus* that are the earliest record of this genus in the Krestovka sections and, as far as is known, in Beringia. They are of a primitive form and are characteristic of the earliest morphotypes of the genus found around the world, although these records differ greatly in age (Repenning, in press). This earliest record, and its primitive evolutionary stage, is inconsistent with the age for the beginning of the Jaramillo Subchron or younger normal-polarity events.

In the Old Crow Basin of Yukon Territory, Canada, within the Beringian faunal province but in its eastern part, a large microtine fauna with a good number of *Microtus* morphotypes, some quite advanced, is known (R.E.

Morlan, written commun., 1991) from deposits judged to represent the Jaramillo Subchron on faunal, climatic, and magnetic polarity criteria (Repenning, in press). Thus, in Beringia the genus was well established by the time of the Jaramillo event, which lasted 70,000 years. The deposits containing the advanced morphotypes of *Microtus* in the Old Crow Basin also contain a pollen flora indicating temperatures as warm as, or warmer than, today (J.V. Matthews, written commun., 1987), similar to the flora in the type Olyor and contrasting equally with the flora below and within the oldest normally polarized zone of the Krestovka Olyor.

In the well-documented evolution of *Microtus paroperarius* in the United States it took 140,000 years to evolve "quite advanced" forms out of early morphotypes like those from locality KLO-9 of the Krestovka sections (Repenning, in press). If it took as long to evolve advanced morphotypes in Beringia, locality KLO-9 in the Olyor Suite of the Krestovka sections, western Beringia, should be a minimum of 70,000 years older and a maximum of 140,000 years older than the Jaramillo event; however, locality KLO-9 is only 0.8 ft (0.25 m) beneath the base of the zone with normal polarity in the Krestovka sections. In addition, climatic evidence from this part of the section, discussed above, suggests that the oldest normal polarity in the Krestovka Olyor cannot represent the Jaramillo Subchron.

Sher and others (1979) reported a brief change to reversed polarity about 1.6 ft (0.5 m) above the base of the normally polarized deposits in the Krestovka sections. This change is indicated by a single polarity determination in uncleaned samples but is indicated by three samples after cleaning (Sher and others, 1979, fig. 6). These three polarity determinations of cleaned samples suggest that the lowest normal-polarity deposits in the Krestovka sections may be less than 1.6 ft (0.5 m) thick. The presence of these higher reversed samples has not been considered significant, but it is possible that the lowest normal polarity of the Krestovka sections represents a brief event (its zone possibly less than a foot (0.3 m) thick) before the Jaramillo Subzone.

This lowest normal polarity recorded in the Krestovka sections may represent the Cobb Mountain Subchron, 1.10 million years old. If the first record of *Microtus* in the Krestovka sections was slightly below the Cobb Mountain Subchron, rather than slightly below the Jaramillo Subchron, a plausible length of time (comparable to that documented for the evolution of *Microtus paroperarius* in the United States) would be indicated for the evolution and diversification of *Microtus* found during the Jaramillo Subchron in eastern Beringia, as represented in the Old Crow Basin locality.

The assignment of this earliest normal polarity to the Cobb Mountain Subchron would indicate a correlation of the most severe climate recorded in the Krestovka sections with

(1) the second beginning of intermittent perennial ice cover on the Arctic Ocean, as reported from CESAR core 14 on Alpha Ridge (Scott and others, 1989) (the first beginning of intermittent perennial ice cover was between 2.4 and 2.1 million years ago, which was Repenning's 1984 interpreted correlation) and (2) the beginning of intense activity of the Scandinavian ice sheet recorded in the Norwegian Sea at site 644 (Jansen and others, 1988, fig. 2). On the other hand, assignment of this short normal-polarity event to the Jaramillo Subchron would indicate a correlation of the most severe climate recorded in the Krestovka sections with (1) the Bavelian warm floral zone of Holland and (2) the Old Crow locality CRH-47 and the type Olyor, both of which contain evidence of a climate as warm or warmer than today. Assignment of this event to the base of the Brunhes would indicate a correlation with warm oxygen stage 19 and a correlation of the primitive and undiversified morphotypes of *Microtus* of locality KLO-9 with the advanced and diverse morphotypes of Beringia at the beginning of the Brunhes Chron, as illustrated by the fauna from the type Olyor (Sher, 1987) and from locality CRH-47 in Canada that lived during the Jaramillo Subchron.

Accordingly, the magnetically reversed and oldest part of the Olyor Suite in the Krestovka sections of Yakutia, as well as less than 1.6 ft (0.5 m) of overlying normally polarized deposits, is considered to be about 1.4-1.1 million years old.

The presence of an essentially treeless pollen flora just before and during the 1.1-million-year-old Cobb Mountain Subchron in the Krestovka sections is correlated with other indications of cold climate from the central Arctic Ocean and Scandinavia and with the Menapian cold flora of the Netherlands. This cold period, ending about 130,000 years before the Jaramillo Subchron, is not recorded in the type section of the Olyor Suite and may be missing.

The remaining normally polarized younger part of the Krestovka Olyor, perhaps 8.2 ft (2.5 m) thick, may represent the Jaramillo Subchron that is recorded in the type section, or it may represent part of the Brunhes Chron, or both. The microtine rodents are too poorly represented in this youngest part of the Krestovka Olyor to evaluate their age (A.V. Sher, oral commun., 1990); they do include more advanced types of *Microtus* (Sher, 1987), however, and cannot be older than the Jaramillo event insofar as knowledge of Beringian microtine history is known at present. In the younger, normally polarized part of the Krestovka Olyor the percentage of arboreal pollen is greater than, and the climatic implications are similar to, that of pollen from the Jaramillo and younger parts of the type section of the Olyor Suite and to the fauna of locality CRH-47 in eastern Beringia.

Member IIIa of the Olyor Suite is interpreted (Repenning, in press) (1) as having been deposited between 1.4 and, perhaps, 0.90 million years ago (including deposits above the Cobb Mountain Subchron); (2) to record the

Cobb Mountain Normal-polarity Subchron of the Matuyama Chron; and (3) to record cooling climate before and during the Cobb Mountain Subchron, with subsequent warming leading into the Jaramillo Subchron.

In the Krestovka exposures, there is no record of renewed cooling prior to the beginning of the Ice Ages and there are no younger reversely polarized deposits; unconformities may be present.

YAKUTIA: TYPE OLYOR SUITE

Sher (1971), Giterman and others (1982), Virina and others (1984), and Sher (1986, 1987) have provided information about the type sections of the Olyor Suite, on the Bol'shaya Chukoch'ya River about 93 mi (150 km) southwest of its (present) mouth at the Arctic Ocean and about an equal distance north of the Krestovka sections (fig. 1). It is somewhat difficult to correlate these several reports because different aspects of the section are emphasized in each.

Two sections were used to define the type Olyor Suite. The older parts of the formation are present in the "primary" type section (Sher, 1971) and include the Jaramillo Subchron (Virina and others, 1984). In the "referred" type section the polarity pattern is continued upward from the youngest reversed polarity of the Matuyama Chron into the lower part of the Brunhes Chron (Virina and others, 1984). The microtine rodent faunas reported from these sections (Virina and others, 1984; Sher, 1987) are from all parts of the section, as subdivided by the polarity determinations, but are all represented by faunal samples too small for population analysis (A.V. Sher, oral commun., 1990). The morphotypes present, as named in Virina and others (1984) and Sher (1986, 1987), are, however, in agreement with the assignment of paleomagnetic events.

The last occurrence of *Lasiopodomys* (listed as *Allophaiomys* in Virina and others, 1984, and Sher, 1986, 1987) is just below the base of the Brunhes Zone, in agreement with its time of extinction in the United States, and the species of *Microtus* listed from this horizon (Virina and others, 1984; Sher, 1986, 1987) are advanced and are not known to be older than the Jaramillo event. They are in agreement with the paleomagnetic assignments.

Giterman and others (1982) presented the pollen profile of the older (primary) type section; it does not match the floral trends of the Krestovka sections record and may contain gaps. Giterman and others (1982) conducted no paleomagnetic studies, and their thickness measurements do not agree with those in other publications. It is not possible, from published information, to identify the position of the Jaramillo Subchron in their pollen sequence except in approximate terms.

Giterman and others (1982) indicated a general similarity of the flora throughout the primary type section and

at the time of the change of tundra types in the Fish Creek section underlying the mammal fauna but also above the Gauss-Matuyama polarity reversal. The absence of physical evidence of permafrost in the Fish Creek section could easily be due to the marine depositional environment.

Pliotomys mimomiformis makes its only North American appearance in the Fish Creek fauna, but it is an abundant part of the reversely polarized Kutuyakh faunas of Yakutia. *Phenacomys gryci* is the other microtine from the Fish Creek fauna; it is the most primitive known species of the genus and is inferred to have been derived from the Asian genus *Cromeromys* (Repenning and others, 1987). A species of *Phenacomys* that is inseparable from *P. gryci*, associated with *Cromeromys*, is known from a locality 4 km (2.5 mi) upstream from the Krestovka sections. This upstream locality is believed to be somewhat younger than the preserved Kutuyakh Beds (Sher and others, 1979). The specimen of *Phenacomys* is the only Old World record of the genus, which is known only from Beringia at this early date. Thus the Fish Creek fauna of Alaska and the late Kutuyakh faunas (including the isolated locality) share two unique records for their respective continents.

The similarities between the Colvillian-Bigbendian-Fishcreekian sections of central Beringia and the Begunov-Kutuyakh units of western Beringia are remarkable and suggest close contemporaneity of deposition. The differences suggest greater aridity and a steppe-forest mosaic flora in the more continental environment of Yakutia.

MARINE AND PROXY RECORDS OF THE FIRST COOLING CYCLE

We are not aware of any terrestrial deposits in the Arctic Ocean Borderland younger than the Kutuyakh Beds of Yakutia (about 2.38 m.y. ago) and older than 2.0 million years—the time of the culmination of the late Pliocene cooling cycle of the Arctic Ocean Borderland.

The absence of terrestrial records is so widespread throughout the borderland that the question of a cause comes to mind. The depositional lacuna existed during a period of Northern Hemisphere continental ice accumulation, and, possibly, the resulting lower sea level made erosion, rather than deposition, the principal process in the Arctic Ocean Borderland. A similar lack of deposits during glacial periods is evident in the Atlantic coastal area of the United States (T.A. Ager, written commun., 1992). Marine records of this time period in the deposits of the Arctic Ocean and adjacent seas bear directly on the history of this coldest and final part of the first cooling cycle.

Marine records show (1) the activity of the Scandinavian ice sheet as indicated by evidence from the Norwegian Sea; (2) deposition within the central Arctic Ocean north of Canada at the eastern end of the Alpha Ridge; and (3) the

same record in the western Arctic Ocean from the Mendeleev Ridge, north of Yakutia and the Siberian Shelf. The dating of these records is supported by lower latitude climatic records in Europe and North America.

NORWEGIAN SEA: SCANDINAVIAN ICE SHEET

Three Ocean Drilling Program cores were collected on the Voring Plateau in the Norwegian Sea, at sites 642, 643, and 644 (Jansen and others, 1988). Site 644 (figs. 1, 2) has the best paleomagnetic control and the least core disturbance, and it is the closest to the coast of Norway; it thus provides the best record of ice-rafted debris resulting from activity of the Scandinavian ice sheet (fig. 2). In the following summary we discuss temporal correlations to other events of the late Pliocene and early Pleistocene cooling cycles of the Arctic Ocean Borderland and elsewhere in the Northern Hemisphere.

Site 644 records the last 2.85 million years of history in the Norwegian Sea. Activity of the Scandinavian ice sheet is indicated in the cores by the percentages of ice-rafted debris derived from this ice sheet when it flowed to the sea and dispersed as icebergs (fig. 2). Jansen and others (1988) reported that major expansion of the Scandinavian ice sheet began about 2.56 million years ago, before the end of the Gauss Chron and before the earliest permafrost in Yakutia and, by inference, Alaska. This was about 120,000 years before the major lowering of sea level indicated in the oxygen isotope records of the World Ocean (stage 100, fig. 2). Stage 100 must reflect Antarctic, as well as Arctic, changes because it does not appear to represent the greatest accumulation of continental ice in the Northern Hemisphere during the first cooling cycle.

About 2.56 million years ago the microtine bog lemming genera *Pliotomys* and *Mictomys* (a genus deriving from *Pliotomys*) dispersed southward from Beringia into the United States (Repenning and Grady, 1988). In the United States the date of this dispersal has been approximated (by interpolation between the Kaena Subchron and the end of the Gauss Chron) at about 2.6 million years ago on the basis of the earliest records of these immigrant genera between 10 and 15 ft (3.05–4.57 m) downsection from the Gauss-Matuyama magnetostratigraphic boundary in Texas, Arizona, and Idaho (Repenning, 1987). Because the dispersal probably was climatically stimulated, the 2.56-million-year date of initial dramatic Scandinavian ice activity is considered a closer approximation.

In the core from site 644, peaks of major glaciation (marked by high percentages of ice-rafted debris, comparable in size to those recorded during the late Pleistocene but not as frequent) persisted until about 2.0 million years ago, with relatively strong peaks between 2.2 and 2.0 million years ago (by interpolation between the Olduvai Subchron and the end of the Gauss Chron). The strongest peak

oldest microtine fauna of the Krestovka Olyor, but no *Microtus* is present at this locality. The species is different and possibly more primitive than *L. deceitensis* from the Cape Deceit fauna (Repenning, in press). Locality KLO-8 is believed to be between 1.6 and 1.3 million years old (Repenning, in press) but cannot be as old as 1.67 million years because the normal polarity of the Olduvai Subchron is not present. Locality KLO-8 cannot be as young as the overlying locality KLO-9. If the difference noted in the *Lasiopodomys* from KLO-8 is significant, then perhaps locality KLO-8 is older than the Cape Deceit fauna.

Predicrostonyx hopkinsi from the Cape Deceit fauna is the most primitive collared lemming known. The ancestry of the collared lemmings is unknown; the genus is thought to have originated in submerged parts of the Arctic Ocean Borderland now forming the Siberian Shelf.

In the Krestovka Olyor Suite, *Predicrostonyx hopkinsi* is known only as reworked specimens derived from unknown pre-Olyor deposits; a more advanced species, *P. compitalis*, in the oldest microtine fauna of the Olyor at locality KLO-8 (Sher and others, 1989) suggests that Krestovka locality KLO-8 is younger than the Cape Deceit fauna, and *Predicrostonyx hopkinsi* from the Cape Deceit fauna suggests that the Cape Deceit fauna may be as old as the Olduvai Subchron, which ended 1.67 million years ago. Unfortunately, there are no paleomagnetic polarity determinations for the Cape Deceit fauna.

In the youngest significant fauna of the Krestovka Olyor, from locality KLO-9, *Predicrostonyx compitalis* is replaced by its descendant, *Dicrostonyx renidens*, which is associated with *Microtus* and *Lasiopodomys deceitensis* at that locality. *Dicrostonyx renidens* is replaced by *D. simplicior* at locality CRH-47 in the Old Crow Basin of Canada (R.E. Morlan, written commun., 1991), which was deposited during the Jaramillo Subchron. This is additional reason to think that the first normal-polarity zone in the Krestovka section is older than the Jaramillo.

The suspicion (Repenning, in press) that the *Lasiopodomys* from locality KLO-8 is more primitive than *L. deceitensis* from the Cape Deceit fauna conflicts with the older age suggested by *Predicrostonyx hopkinsi* from Cape Deceit. Both suggestions are weak: the supposed primitiveness of *Lasiopodomys* from locality KLO-8 is not supported by morphologic trends in any other fauna in the world and the identification of species of *Predicrostonyx* (and *Dicrostonyx*) is based on average population morphology, and the KLO-8 sample of the lemming is not very large, raising the doubt that it may not be representative of the population.

As a result, the Cape Deceit fauna is also considered to have an age range from possibly 1.6 to 1.2 million years (Repenning, in press), similar to the possible age range

assigned to locality KLO-8, and it is not certain which locality, if either, is older.

If the Cape Deceit fauna is as old as the Olduvai Subchron, it would be the oldest record of *Lasiopodomys* in the world except for the record of *L. praebrandti* from the Lake Baikal area (Erbaeva, 1976). This record, from the Kudun fauna, is not dated by external evidence but on other faunal evidence is believed to be about the age of the Olduvai Subchron (Repenning and others, 1990). It is also believed to be a fauna of the Chinese faunal province of Asia, which may have a biochronology with a somewhat different time scale and history than those under discussion.

In his analysis of the environments of the Cape Deceit Formation, Matthews (1974, table 6) summarized six successive environments for the formation and for the periods both immediately before and immediately after its deposition. The environment before deposition of the formation was cold and the environment following deposition was much warmer than today.

Within the formation, Matthews recognized four climatic units: two tundra environments, colder than today, and two forest or forest-tundra intervals, warmer than today. The oldest warm period was the warmer of the two recognized within the formation; at this time spruce grew at or near Cape Deceit. The youngest cold period was the colder of the two recorded in the sediments; at this time herbaceous tundra grew on Cape Deceit, and the deposits show evidence of permafrost. This was the period during which the majority of the small mammals of the Cape Deceit fauna lived.

In reality, the lack of forest today on the Seward Peninsula (and at Cape Deceit) is a coastal feature; forests extend farther north farther inland. The evidence of former forests in the Cape Deceit Formation possibly represents low sea levels, with the coast more removed from the locality, and thus a "cold" period could be inferred from the presence of forests in the Cape Deceit Formation.

The top of the Cape Deceit Formation, immediately above the unit containing most of the mammalian fauna, is truncated by an erosional unconformity believed to represent a much warmer time than the present. This warm period, which resulted in thawing, slumping, and erosion of the previously frozen sediments, is the youngest climatic period that Matthews (1974) associated with the Cape Deceit Formation.

This history of alternating tundra and forest in the Cape Deceit Formation, and a climaxing peak of cold climate (followed by temperatures much warmer than today), is in agreement with the climatic history of the Arctic Ocean Borderland for several periods of time between about 1.8 and 1.1 million years ago. It conforms broadly with the nature of the relatively quiet period of glacial

activity recorded at site 644 in the Norwegian Sea but does not obviously correspond to any particular part of it. No basis exists for refining the age of the formation by comparison with climatic events, but this climatic history does tend to support the general fluctuations of climate during the earlier part of the second cooling cycle that were inferred from the record of ice-rafted debris at site 644 in the Norwegian Sea.

CANADA: YUKON TERRITORY— LOCALITY CRH-47

Locality CRH-47, along the Old Crow River in the northern part of Yukon Territory, Canada (fig. 1), is near the Canada-Alaska border and about 100 mi (161 km) south of the Arctic Ocean in easternmost Beringia. The locality was discovered by C.R. Harington and contains a very large microtine rodent fauna that is being studied by R.E. Morlan. Sediments at the locality have normal magnetic polarity and a pollen flora indicating boreal forest (Schweger, 1989); the climate during deposition was as warm as, or possibly somewhat warmer than, today.

According to Morlan (written commun., 1991), microtine rodents in the fauna include *Lasiopodomys deicei*, a variety of *Microtus* morphotypes, *Phenacomys*, *Dicrostonyx renidens*-*Dicrostonyx simplicior* (intergrading from bottom to top of the section), *Mimomys (Cromeromys)* sp. cf. *M. (C.) virginianus*, and other genera.

As noted above, *Dicrostonyx renidens* replaced *Predicrostonyx compitalis* in the fauna from locality KLO-9 of the Krestovka sections in Yakutia, western Beringia, and only primitive morphotypes of *Microtus* were present at locality KLO-9. About 140,000 years were required to evolve the endemic *Microtus paroperarius* of the United States from a primitive morphotype comparable to the specimens at locality KLO-9 to a population morphologically varied, with advanced forms, comparable to those at locality CRH-47 (Repenning, in press). The presence of *Microtus* at this stage of evolution and of *Dicrostonyx* as more advanced species than those from KLO-9 indicates that the CRH-47 fauna is younger than the KLO-9 fauna.

The Cobb Mountain Subchron, immediately overlying locality KLO-9, was perhaps only 10,000 years long (Mankinen and Gromme, 1982), and the next normal polarity event was the Jaramillo, which thus must be represented by the normal polarity during deposition of locality CRH-47; the time involved between the ages of localities KLO-9 and CRH-47 (130,000-200,000 years) thus represents the time span of the species *Dicrostonyx renidens* and the evolution of varied and advanced morphotypes of *Microtus* in Beringia. This time span is comparable to that observed in the similar evolution of *Microtus paroperarius* in the United States (noted above) (Repenning, in press).

That locality CRH-47 in Yukon Territory cannot be as young as the next normal polarity (Brunhes Chron) is indicated by the presence of *Lasiopodomys deicei* and *Mimomys (Cromeromys)* sp. cf. *M. (C.) virginianus* in that fauna (R.E. Morlan, written commun., 1991). In the type locality of the Olyor Suite, western Beringia (Yakutia), *Lasiopodomys* (as *Allophaiomys*; Sher, 1987) became extinct before the Brunhes Chron, and the same date of extinction is recognized in the conterminous United States (Repenning, in press); lacking evidence to the contrary, it is reasonable to assume the same extinction date in eastern Beringia of North America. *M. (C.) virginianus* from the Cheetah Room fauna, West Virginia (Repenning and Grady, 1988), is older than the Brunhes Chron and is similar to *M. (C.) dakotaensis* from the much older (~1.5 million year old) Java fauna of South Dakota (Martin, 1989). No similar morphotype of *Mimomys (Cromeromys)* is known anywhere in the world during the Brunhes Chron, although the dissimilar but related *M. (C.) savini* (= *M. (C.) intermedius*) is. Finally, *Dicrostonyx simplicior* is known in Europe before the beginning of the Brunhes Chron, and it is not reasonable that its evolution out of *D. renidens*, recorded in locality CRH-47, could be an event of the Brunhes Chron. Thus the normal paleomagnetic polarity of locality CRH-47 in the Old Crow Basin was assigned to the Jaramillo Subchron by Repenning (in press).

C.E. Schweger (1989) examined the pollen of locality CRH-47 and concluded that it indicates an environment as warm as today or possibly somewhat warmer. This assessment is in agreement with the floral record during the Jaramillo Subchron as described from the type section of the Olyor Suite (a low percentage of arboreal pollen in an area now occupied by tundra) and with the warm Bavelian floral of Holland as described by Zagwijn and de Jong (1984), also of Jaramillo age. This warm period is not evident in the cores from site 644 in the Norwegian Sea or CESAR site 14 on the eastern Alpha Ridge but neither is such a warm period evident during the Brunhes Chron. The CRH-47 fauna is older than the 840,000-year-old Cheetah Room fauna of West Virginia based on the history of *Microtus* and on the normal magnetic polarity of locality CRH-47 (Repenning and Grady, 1988; Repenning, in press); however, locality CRH-47 may not be much older.

MARINE AND PROXY RECORDS OF THE SECOND COOLING CYCLE

Similar to the last part of the first cooling cycle, there are few records in the Arctic Ocean Borderland that give evidence of climatic change during the second cooling cycle. The Kap København Formation in northern Greenland provides evidence of treeline at lat 82°30' N. 2 million years ago at the beginning of the second cycle; this

position is supported by the marine record of little glacial activity in the Norwegian Sea and by other records that have been discussed. This was the Tiglian warm period as recognized in Holland.

The remainder of the second cooling cycle was reconstructed by the intercorrelation of the few records from the Arctic Ocean Borderland that have been discussed and lower latitude proxy evidence.

NORWEGIAN SEA: SCANDINAVIAN ICE SHEET

Between 2.0 and about 1.5 million years ago there is little evidence of climatic change in the Arctic except that from site 644 in the Norwegian Sea. The part of this record that corresponds to the warm Tiglian flora of Holland indicates minor peaks in glacial activity at the beginning (1.90 m.y. ago) and near the end (~1.7 m.y. ago) of the Olduvai Subchron (fig. 2). The first of these peaks is coincident with lower latitude proxy evidence of the dispersal of the distinctly temperate to subtropical microtine rodent *Allophaiomys* (1) from Beringia south to the United States at the beginning of the Olduvai Subchron (possibly somewhat more than 1.9 m.y. ago; Repenning, in press) and (2) from the Black Sea area northward into Holland during the earliest part of the Olduvai (van Kolfschoten, 1990a; somewhat less than 1.9 m.y. ago). The latter dispersal would be during the last half of the Tiglian warm floral age (Zagwijn and de Jong, 1984) and immediately following the minor peak in glacial activity indicated at the beginning of the Olduvai Subchron in the core from site 644 (fig. 2).

During the Olduvai the average ice activity was low (mostly background) in the record of the Norwegian Sea; this period coincides, on the basis of paleomagnetic correlation, with the last half of the warm Tiglian flora age of Holland (Zagwijn and de Jong, 1984). Again on the basis of paleomagnetic correlation, the period of minor, but increased, glacial activity late during the Olduvai Subchron (beginning ~1.7 m.y. ago) correlates with the beginning of the Eburonian flora of Holland and with the dispersal of *Phenacomys* southward into the United States. Both the appearance of Eburonian colder floras and the increase in ice rafting of debris occurred late in the period of time represented by the Olduvai Normal-polarity Subchron.

The end of the Eburonian cold flora and the beginning and ending of the following Waalian warm and Menapian cold floras of Holland (Zagwijn and de Jong, 1984) all are within the 700,000-year-long reversed-polarity part of the mid-Matuyama Chron, which is a period of reversed polarity almost as long as the normal polarity representing the Brunhes Chron. Except for the very brief Cobb Mountain

Subchron, the lack of polarity reversals during this 700,000-year period provides no basis for approximate age interpretation between the Olduvai and Jaramillo Subchrons. The only possible solution for an approximate age interpretation of these floral ages in northwestern Europe is in the correlation with oceanic records, where deposition rates were presumably more uniform and interpolation between age control points (polarity reversals) may be used with greater confidence than in continental deposits.

This correlation of these floral ages with the record of Scandinavian ice sheet activity suggests that (1) the Eburonian cold floral age terminated about 1.58 million years ago; (2) the Waalian warm floral age ended about 1.40 million years ago; and (3) the Menapian cold floral age ended about 1.10 million years ago and the Bavelian warm floral age began shortly before the Jaramillo Subchron (at the Cobb Mountain Subchron, which marks the last cold period before the Jaramillo Subchron) (fig. 2).

To judge from the record of ice-rafted debris at site 644, Scandinavian ice sheet activity less than 1.2 million years ago was more continuous and is characterized by a lack of prolonged minima in debris accumulation. The greatest amounts of ice-rafted debris were about 400,000 years ago, possibly correlating with the establishment of permanent perennial ice cover on the Arctic Ocean as inferred in CESAR core 14 (Scott and others, 1989), although Scott and others suggested a slightly later date.

MICROTINE IMMIGRANTS: WESTERN UNITED STATES

The distribution of microtine rodents is strongly constrained by climate; for the most part they are temperate to arctic animals. For 15 years it has been widely recognized that new appearances of microtines in various areas of the Northern Hemisphere coincide with colder climates, which allow more northerly forms to spread southward (Fejfar, 1976). Until quite recently, however, it has not been generally recognized that some microtine taxa more adapted to temperate climates spread northward during warmer climates. Because generally cooling climatic patterns are discussed herein, the examples of northward dispersal during warmer periods are few (Repenning, 1990, in press) in the present discussion. These few, however, are informative and provide support for climatic age assignments in the Arctic Ocean Borderland. The following is a review of this evidence, of both cooling and warming periods, more fully discussed in Repenning (1987, 1990, in press), Repenning and others (1987), Repenning and Grady (1988), and Repenning and others (1990).

FIRST COOLING CYCLE

The primitive bog lemming, *Pliotomys*, is first known about 4 million years ago in southern Russia and later in Hungary, Poland, and Mongolia. It is abundant in the upper Kutuyakh Beds of the Krestovka sections, where it is slightly younger than the Gauss-Matuyama polarity boundary; it is also known at this time from northern Alaska. *Pliotomys* dispersed southward from Beringia to the United States (1) just before the end of the Gauss Chron; (2) at the time that activity of the Scandinavian ice sheet began to increase; (3) with the density stratification of the Arctic Ocean; (4) when boreal forests were thinning; (5) as treeline was moving south from the shores of the Arctic Ocean; and (6) before full tundra and permafrost were established in Beringia. The genus persisted in the United States (as the more advanced species *Pliotomys rinkerii*) until sometime before 2.1 million years ago and then became extinct (Repenning and Grady, 1988). It was replaced, in the North American fauna, by the most primitive morphotype of the derived genus *Mictomys*. *Mictomys*, the living northern bog lemming, lives in boreal forests but avoids tundra.

The two genera *Pliotomys* and *Mictomys* are first known in the United States from deposits about 15 ft (~4.5 m) below the Gauss-Matuyama polarity boundary in Texas (*Pliotomys*) and in Arizona and Idaho (*Mictomys*). The significance of the dispersal of the bog lemmings is not so much that it corroborates the timing of these events in the Arctic, because these events are rather well dated, but that it provides an idea of the effect of climatic deterioration on climates in lower latitudes of North America.

Pliotomys became extinct before 2.1 million years ago (there are no records this young), following the dispersal of *Mictomys* around the southern end of the Rocky Mountains to the Great Plains. The earliest record of *Mictomys* in the Great Plains is in Kansas 2.0 million years ago and 100,000 years after the deposition of glacial till in Iowa; there are no known associations of early *Mictomys* with *Pliotomys*. The range of *Mictomys* has subsequently withdrawn northward with the Canadian forests, but, both east and west of the Rocky Mountains, much of this retreat has been after the end of the Pleistocene; during the Ice Ages it returned southward with several, perhaps all, ice advances, as presumably did the forests.

A lineage called *Synaptomys* diverged from *Mictomys* in the Appalachian Mountains about 840,000 years ago (Repenning and Grady, 1988) and now lives in the north-eastern United States and adjacent southeastern Canada; relict populations, surviving from a late Pleistocene distribution, live as far southwest as Kansas.

Other than *Pliotomys* and *Mictomys*, no microtines are known to have dispersed southward in North America during the first cooling cycle.

BEGINNING OF THE SECOND COOLING CYCLE

The extinct microtine genus *Allophaiomys* had a distinct preference for temperate climates. From perhaps 2.2 million years ago (Beijing, lat 40° N.) to 0.83 million years ago (Maryland, lat 40° N.) it was abundantly present throughout the lower latitudes of the Northern Hemisphere (Repenning, in press). Its known latitudinal range was 50°–30° N. in North America and 52°–31° N. in Europe and Asia Minor. *Allophaiomys* may have originated in China (perhaps 2.2 m.y. ago) and dispersed to both Europe and North America, where its first record is about 1.9 million years old in both Holland (van Kolfschoten, 1990a) and southern Colorado (Repenning, in press). There are no records of *Allophaiomys* in the Arctic, although the genus must have arrived in North America by way of the Bering Land Bridge (about lat 63°–69° N.). One must assume that *Allophaiomys* crossed the Bering Land Bridge during a period of temperate climate but was forced southward in North America by the return of cold climate.

As discussed above, 2.1 million years ago a continental glacier flowed down the Mississippi River Valley as far south as Iowa, ice cover was present on the central Arctic Ocean, and ice-rafted debris from the Scandinavian ice sheet was as great as it was later during most of the Ice Ages (fig. 2); this was hardly a time of temperate climate on the Bering Land Bridge.

In North America *Allophaiomys* is first known in Colorado, where it was recovered from rocks recording the 1.9-million-year-old beginning of the Olduvai Subchron or from just beneath them (Repenning, in press). One must assume that between 2.1 (cold in Beringia) and 1.9 million years ago *Allophaiomys* crossed the Bering Land Bridge and was forced southward in North America during the brief cold period at the beginning of the Olduvai event that is indicated by the ice-rafted debris record of the Norwegian Sea (site 644) and by the isotope records of the World Ocean (site 607, fig. 2). If both assumptions are reasonable, one must then assume that Beringia had a temperate climate between 2.1 and 1.9 million years ago.

The coincidence of this constraint on the age of a temperate climate on the Bering Land Bridge with the age, inferred earlier, of the treeline forest in member B3 of the Kap København Formation of northernmost Greenland is mutually supportive of these interpretations. Thus the 1.9-million-year date of the oldest *Allophaiomys* from North America substantiates the age assignment that has been inferred for member B3 of the Kap København Formation, even though *Allophaiomys* is not known from Greenland or Beringia. It also indicates that the warm climate recorded in the Kap København Formation was not, as might be supposed, a local climatic extreme caused by conditions in the

North Atlantic Ocean but was also evident in the Bering Strait region of the North Pacific Ocean and thus may have been present throughout the Arctic Ocean basin.

This extreme warm period between 2.1 and 1.9 million years ago that is recorded in north Greenland and that probably existed in Beringia marked the end of the first cooling cycle and the beginning of the second. The rather widespread ability to correlate this time in the contiguous United States is due to the fortuitous eruption of the Huckleberry Ridge Tuff in the Yellowstone National Park area of Wyoming and Idaho (Christiansen and Blank, 1972) 2.01 million years ago, to the spread of its ash in still-visible deposits across much of the United States from the Pacific Ocean to the Mississippi River (Izett and Wilcox, 1982), and to the coincidence of this eruption with the younger normal-polarity event of the Reunion Subchron.

Perhaps 500,000 years later than its oldest record in the United States, there is no record of *Allophaiomys* in Beringia as indicated by the oldest fauna of the Krestovka Olyor Suite (locality KLO-8; possibly 1.4 m.y. old; Repenning, in press) nor is the genus known in younger Olyor faunas, in the Cape Deceit fauna, or the Old Crow Basin during the Jaramillo Subchron (0.97-0.90 m.y. ago).

In the United States *Allophaiomys* is not known west of the Rocky Mountains; for some unknown reason the dispersal route west of the Rocky Mountains was not open at this time. Yet before the end of the Olduvai Subchron the genus *Phenacomys* immigrated, west of the Rocky Mountains, from Beringia to Idaho; failure of *Allophaiomys* to accompany *Phenacomys* southward at this time suggests that *Allophaiomys* may have become extinct in the Arctic Ocean Borderland before the end of the Olduvai Subchron (Repenning, in press). This hypothesis is consistent with the inference that *Allophaiomys* was present in Beringia only during a very brief and exceptionally warm period, apparently about 2.0 million years ago.

During most of the Matuyama Chron *Allophaiomys* occupied the territory in North America from southern Saskatchewan to southern Texas and from the Rocky Mountains to the Atlantic coast. For more than a million years the genus thrived in the United States but it probably was able to endure Beringian climates for less than 100,000 years. One must assume that the temperate climate of Beringia between 2.0 and 1.9 million years ago was most unusual and, by inference, that the warmth 500 mi (805 km) from the North Pole at Kap København was equally unique.

Allophaiomys, possibly older than the Olduvai Subchron, is also known along the northern shore of the Black Sea and in southern Europe (Alexy Tesakov, Moscow, oral commun., 1990). However, it was not until after the beginning of the Olduvai Subchron that the genus first appeared in Holland (van Kolfschoten, 1990a). In the same

record (Zuurland Boreing; van Kolfschoten, 1990a) this warm climate genus possibly became extirpated from Holland with the beginning of the Eburonian, late during the Olduvai Subzone. The history of the distribution of *Allophaiomys* not only supports the age assignment of the forest-tundra in the Kap København Formation of North Greenland, and provides some insight into the probable climate in Beringia during an extremely warm 200,000 years that started the second cooling cycle of the Arctic Ocean Borderland, but it also suggests that this warming may have been influential as far away as northern Europe.

It is universally agreed that the meadow mouse *Microtus* is derived from *Allophaiomys*. During this evolution, *Microtus* developed arctic, as well as temperate tolerances (but lost subtropical tolerances), and it now lives throughout the Arctic Ocean Borderland and southward (in highlands) into Mexico and to the Mediterranean coastal area of North Africa. *Microtus* does, however, have climatic limits (Repenning, 1990). Although it was present in Beringia 1.1 million years ago (in the Krestovka Olyor locality KLO-9), was morphologically varied during the Jaramillo Subchron (0.97-0.90 m.y. ago), and was present in southern Canada after the Jaramillo Subchron (Repenning, in press), *Microtus* did not enter the United States (east of the Rocky Mountains) until 450,000 years ago (possibly as late as the development of permanent perennial ice on the Arctic Ocean according to Scott and others, 1989). The absence of *Microtus* in the United States during the second cooling cycle suggests a subtropical climate and (or) excessive rainfall (Repenning, 1990), an indication of the climate preferred by *Allophaiomys*, which was present in the United States during much of this time.

The replacement of temperate to subtropical *Allophaiomys* by temperate to arctic *Microtus* happened many times in various parts of the Northern Hemisphere. Because the derivation of *Microtus* from *Allophaiomys* is widely accepted, claims have been made of the origin of *Microtus* in France, Ukraine, Western Siberia, Buryat, Yakutia, China, and the United States. Most claims result from the ecologic replacement of *Allophaiomys* by *Microtus* during southward climatic shifts. The place of origin (or origins) of *Microtus* has not yet been identified, with the exception of the species *Microtus paroperarius*, which evolved endemically in the United States and never left that country (Repenning, in press).

EBURONIAN COLD

The warm spell at the beginning of the second cooling cycle began about 2.0 million years ago and lasted, with minor interruption, until peaks of ice-rafted debris from the Scandinavian ice sheet again appeared in abundance at site 644 in the Norwegian Sea, shortly before the end of the Olduvai Normal-polarity Subchron. This same

time constraint marks the beginning of the cold-climate Eburonian flora of Holland. Other than these two records, there is no further indication of the beginning of this cold event, although the southward dispersal of the heather vole *Phenacomys* to the United States is contemporaneous and suggestive. There are no deposits of this age in the Arctic Ocean Borderland, and there are no obvious signals suggesting the beginning of the Eburonian in the cores of the Arctic Ocean. The Eburonian return of colder climate (recorded in Holland) may have been a North Atlantic feature caused by a temporary collapse of the strong thermal gradient between the Norwegian Sea and the North Atlantic. Jansen and others (1988) felt that this gradient existed until 1.2 million years ago. Microtine dispersal provides one further clue.

The heather vole, *Phenacomys*, is first known about 2.4 million years ago in Beringia; the genus is of Beringian origin. Only one tooth representing it has ever been reported from the Old World, and this record is from eastern Beringia in the Kolyma Lowlands of Yakutia (Sher and others, 1979, p. 38).

In the contiguous United States *Phenacomys* first appears in the Froman Ferry fauna of western Idaho (unpublished data, Repenning). The associated fauna is otherwise completely characteristic of the older Grand View fauna of that area (Blancan V age); the presence of immigrant *Phenacomys* is considered as indicative of the next younger microtine age, Irvingtonian I. The species of *Phenacomys* from Froman Ferry is *P. gryci*, named from the Fish Creek fauna of the Alaskan North Slope. Preliminary paleomagnetic study of the Froman Ferry faunal sequence (J.T. Sankey, written commun. 1990) indicates that *Phenacomys* first appears in the section shortly below the top of the Olduvai Subzone. The age indicated would approximate that of the beginning of the Eburonian cold period and the increase in Scandinavian ice activity.

The immigration of *Phenacomys* to Idaho would have been along the coastal route from Alaska, west of the Rocky Mountains. East of the Rocky Mountains the oldest record of *Phenacomys* is in the Java fauna of South Dakota (Martin, 1989; listed as "*Hibbardomys*"). This fauna also has a high percentage of Blancan holdovers and is considered a very old Irvingtonian fauna (Martin, 1989).

The fossil reptiles and amphibians of the Java fauna indicate greater rainfall than today and considerably warmer winter air temperatures (Holman, 1977). The presence of the cotton rat, *Sigmodon*, in the Java fauna indicates that winter freezes were uncommon (or lacking), and a subtropical climate similar to modern St. Louis, Missouri (or farther south), rather than modern South Dakota, is indicated. Subtropical *Allophaiomys* is also present in the Java fauna, but temperate to arctic *Microtus* is not (Martin, 1989).

This climatic pattern in South Dakota does not suggest a correlation with the Eburonian cold flora of Holland, or with the Froman Ferry fauna of Idaho, but instead a correlation with the preceding Tiglian warm flora or the following Waalian warm flora. The paleomagnetic polarity of the Java locality is unknown, but the fauna represents a wet and warm environment; a climate brought into this area by a northward shift of the North Atlantic Bermuda High (Repenning, 1990) that warmed the regions of Holland and South Dakota is indicated.

If *Phenacomys* entered the United States on the east side of the Rocky Mountains at the same time that it entered on the west, it would be with a cooler winter, hotter summer, and drier climate more like today; the Bermuda High would have been in a southern position. Therefore, it is unlikely that the Java fauna could represent the time of the end of the Tiglian flora and the Froman Ferry fauna. The Java fauna may postdate the arrival of *Phenacomys* east of the Rocky Mountains and correlate with the Waalian warm flora of Holland (possibly 1.58–1.40 m.y. ago); early Irvingtonian I faunas of the United States as young as 1.40 million years contain a high percentage of Blancan V taxa. Present evidence does not exclude, however, the possibility that, on the east side of the Rocky Mountains, *Phenacomys* entered the United States with *Allophaiomys* 1.9 million years ago, before the genus followed the route down the Pacific coast; in which case the warm Java fauna would correlate with the Tiglian warm period.

BEGINNING OF THE ICE AGES

At the beginning of the Ice Ages, the end of the second cooling cycle as here used, the microtine genera *Lasiopodomys* and *Terricola* first entered the United States from Beringia; this was about 850,000 years ago (Repenning, in press). *Lasiopodomys* is known for perhaps 650,000 years before this time in Beringia, but *Terricola* was (and is) a southern form in Eurasia and, similar to *Allophaiomys*, has no record in Beringia. It also presumably passed across the Bering Land Bridge quickly and then retreated southward in both Asia and North America with the next cold period. The genus still is a temperate form in the living fauna of Eurasia, but in North America it may survive only in Mexico as *Terricola quasiater* (Repenning, in press). *Terricola meadensis*, the immigrant species, survived in the United States at least until the Porcupine Cave fauna in south-central Colorado (Barnosky and Rasmussen, 1988), which, because it contains both the endemic *Microtus paroperarius* and the immigrant *Microtus* sp. cf. *M. montanus*, is believed to be the age of the last immigration from Beringia and about 400,000 years old.

Endemic *Microtus paroperarius* apparently became extinct with the immigration of Beringian *Microtus*, and the climatic implications of the immigration of Beringian *Microtus* suggest that *Terricola* shortly moved southward to Mexico; thus the association of these taxa in the Porcupine Cave fauna suggests that the fauna preserves a relatively brief time of north-south faunal association. Unlike the record of climate assumed to have existed during the passage of *Allophalomys*, there is no proxy evidence in the Arctic Ocean Borderland suggesting a warm period that might have permitted the *Terricola* passage across the Bering Land Bridge.

It is not difficult to correlate the immigration event of *Lasiopodomys* and *Terricola* around the Northern Hemisphere because it was during a marked climatic change that permitted the fauna of the Arctic Ocean Borderland to move southward in both Old and New Worlds; Azzaroli (1983) called it the "End-Villafranchian Dispersal Event" in Eurasia and summarized the faunal change there. Repenning (in press) summarized the age control of microtine dispersal during this event in North America; the dispersal was presumed to correlate with oxygen isotope stage 22 (about 850,000 years ago) and with the first of the "Nebraskan" ice sheets. Dated microtine faunas in the United States constrain the time of dispersal to well after 900,000 years ago and well before 820,000 years ago. If large mammals in North America were associated with this dispersal event, as they were in Eurasia, their time of dispersal has not yet been identified with sufficient certainty to suggest an age.

During the Ice Ages there was one additional southward dispersal known in North America; Europe apparently experienced several (Repenning and others, 1990). The North American immigration was believed to have occurred about 450,000 years ago on the basis of the age of isotope stage 12 (Repenning, 1984), but evidence from the CESAR-14 site suggests that the immigration may have been during isotope stage 8. Both interpretations are based on the assumption that the dispersal occurred during an extreme cold period and not on any direct evidence of the faunal age. This is the dispersal that is recorded in Porcupine Cave, Colorado, which is not otherwise dated.

SUMMARY: LATE PLIOCENE-EARLY PLEISTOCENE ECOLOGIC CHANGE

About three and a half million years ago the average ratio of oxygen isotopes indicates that the volume of global ice stored on land (inverse-isostatic sea level) began to increase. Ice storage was greatest in the polar regions, but the records that indicate the development of ice storage in the Arctic region during the two cooling periods do not

closely match the global average of isotope changes. Thus, differences in proportionate ice storage in the Arctic, in mountain glaciers at lower latitudes, and in the Antarctic are indicated.

From 3.0 to 2.4 million years ago, evidence from both northeastern Siberia and northern Alaska and Canada indicates that the Arctic Ocean Borderland experienced a southward shift of treeline, changing from a coastline occupied by mixed coniferous forest, to treeline forest-tundra, and then to extensive tundra and the first record of permafrost. Treeline moved southward from almost lat 80° N. about 3 million years ago to less than lat 70° N. by 2.4 million years ago.

The records of marine ostracodes across this wide latitude zone indicate that about 3.0 million years ago the adjacent sea was cold-temperate to subfrigid, similar to sea temperatures adjacent to modern Labrador and Nova Scotia. By 2.40 million years ago the marine ostracode fauna had changed to one indistinguishable from the modern fauna of the Arctic Ocean with a temperature tolerance of subfrigid to frigid; tundra developed on shore. Climatic indications derived from associated mollusks overlap the temperature indicated by ostracodes in the subfrigid range at lat 70° N.

The Siberian record, in western Beringia, indicates that ice-wedge polygons, permafrost, and steppe-tundra first developed about 2.4 million years ago, contemporaneous with a similar change in northern Alaska that is inferred from the change in tundra type. Both invertebrate and vertebrate evidence in Alaska indicate that the Arctic Ocean was not yet frozen to the shore, even during the winter.

Somewhat earlier (about 2.56 m.y. ago), Beringian bog lemmings dispersed southward to the conterminous United States, to as far south as central Arizona. At the same time the Arctic Ocean became density stratified and the number of icebergs markedly increased; the Scandinavian ice sheet began to enlarge, sending greater amounts of ice to the shores of and across the Norwegian Sea.

From 2.5 to 2.1 million years ago, the climate of the Northern Hemisphere was dominated by the first significant continental glaciation of the late Tertiary. Records around the Arctic Ocean Borderland and in the central United States indicate that this glaciation developed to a maximum between 2.2 and 2.1 million years ago and then terminated. Direct evidence of increasing glaciation comes from (1) bottom sediments of the Norwegian Sea (Jansen and others, 1988), which indicate activity of the Scandinavian ice sheet; (2) North Atlantic site 552A (Shackleton and others 1984), which also indicates increases in ice-rafted debris; (3) the western Arctic Ocean site T3-67-12 (Herman and Hopkins 1980; Herman and others, 1989; their Unit II), which indicates increased iceberg activity and salinity stratification; (4) the central Arctic Ocean site

CESAR-14 (Scott and others, 1989), which indicates that the central Arctic Ocean had intermittent perennial ice cover at this time; (5) the central United States, which indicates continental glaciation flowing southward to Iowa 1.1 million years ago; and (6) the Kap København Formation in northernmost Greenland (Funder and others, 1985), which records the end of this first period of ice cover. Mollusk faunas from the oldest part of the Kap København Formation indicate that marine temperatures were as cold as today, and these fauna are the oldest Arctic Ocean mollusk fauna with no extinct elements.

For most of this first glaciation there are no known terrestrial deposits in the Arctic Ocean Borderland, possibly because of a lowered base level caused by lowering sea level. A similar lack of deposits during glacial periods is evident in the Atlantic coastal area of the United States (T.A. Ager, written commun., 1992).

The first significant late Cenozoic continental glaciation of the Northern Hemisphere, presumably as extensive as any later North American glaciation, reached a peak between 2.2 and 2.0 million years ago; the 2.1-million-year-old first normal-polarity excursion of the Reunion event has been located within till in Iowa (Easterbrook and Boellstorff, 1984). This glaciation at the end of the first cooling cycle terminated abruptly, possibly within 50,000 years; the 2.0-million-year-old second excursion of the Reunion event is present in interglacial deposits overlying the till in Iowa (Easterbrook and Boellstorff, 1984). By 2.0 million years ago, climates much warmer than today were present in the Arctic Ocean Borderland, and treeline moved northward to within about 500 m (800 km) of the North Pole. The temperate to subtropical microtine genus *Allophaiomys* was then able to move as far north as the Bering Land Bridge, enter North America, and return southward to the United States.

The warming of the Arctic Ocean Borderland to conditions similar to modern temperate climates at Beringian latitudes is also indicated by (1) the appearance of mixed conifer treeline forests on the northernmost shores of the Arctic Ocean at Kap København, lat 82°30' N. (Funder and others, 1985); (2) the presence of much warmer water ostracodes (cold temperate to warm subfrigid) living at this locality on the northern shores of Greenland, considerably farther north than their latitudinal limits today (Brouwers and others, 1991); (3) the great reduction in activity of the Scandinavian ice sheet (Jansen and others, 1988); (4) the lack of further evidence of glacial activity in the midwestern United States until the end of the second cooling cycle (Easterbrook and Boellstorff, 1984); (5) the beginning of the warm Tiglian floras of northwestern Europe; and (6) the entrance of the microtine genus *Allophaiomys* into North America, which indicates temperate conditions in the Bering Strait region of Beringia. Subtropical sea-surface temperatures were indicated for this time in the vicinity of Cornwall, England (Jenkins and Houghton, 1987), and

subtropical temperatures and greater rainfall are indicated, possibly for a somewhat later time, in north-central South Dakota (Holman, 1977). In Holland the warm Tiglian flora was present from 2.0 to about 1.7 million years ago (van Kolfschoten, 1990a).

This unusually warm period marked the beginning of the second cooling cycle. The warmth persisted for about 300,000 years with only one modest cold period at the beginning of the Olduvai Subchron (1.9 m.y. ago) when *Allophaiomys* entered the United States on the eastern side of the Rocky Mountains. About 1.7 million years ago ice activity of the Scandinavian ice sheet increased, the cold Eburonian flora became established in Holland, and the heather vole *Phenacomys* migrated down the western side of the Rocky Mountains from Beringia to Idaho in the United States.

Allophaiomys persisted in the United States for more than a million years, until shortly after the beginning of the Ice Ages at the end of the second cooling cycle, about 830,000 years ago (Repenning, in press). For at least part of this time, most of the United States appears to have had a subtropical climate. Jansen and others (1988) reported the presence of a steep temperature gradient between the Norwegian Sea and the North Atlantic that may have accentuated the warmth of this period and caused a northward shift of the subtropical North Atlantic Bermuda High, bringing warmth and greater rainfall to the United States east of the Rocky Mountains and to the Atlantic regions of Europe. In both Beringia and the contiguous United States, the genus *Lasiopodomys* also became extinct at this time. *Phenacomys* still lives, as several more advanced species, in Canadian forest environments and humid coastal forests of the Pacific Northwest of the United States.

The return of stronger and more frequent peaks of ice-rafted debris from the Scandinavian ice sheet marks the beginning of the Eburonian cold floral age of Holland, which began late during the Olduvai Subchron (van Kolfschoten, 1990a). From the end of the Olduvai Subchron until the Jaramillo Subchron there is no paleomagnetic clue available to form a basis for correlation, and it is an assumption, made only from the record of core 644 and its temporal coincidence with those events that are dated elsewhere, that the pattern of variation in ice-rafted debris after the end of the Olduvai Subchron corresponds to the floral ages of Holland. With this assumption, the Waalian warm floras began about 1.58 million years ago (the Eburonian being shorter than generally portrayed) and the Menapian cold flora began about 1.19 million years ago. The peaks of Scandinavian ice influx decreased after 1.1 million years ago (Cobb Mountain Subchron), and this time probably marks the end of the Menapian cold floras and the beginning of the Bavelian warm floras, which straddle the Jaramillo Subchron (Zagwijn and de Jong, 1984).

Thus, we suggest that the Cobb Mountain Subchron, discussed earlier in connection with the cold climate in the Krestovka sections, marks the end of a cold, ice-active period that correlates with the Menapian flora of Holland, although the Cobb Mountain Subchron is not recorded in core 644 nor in Holland. The subchron is believed to be present in CESAR core 14 of the central Arctic Ocean, where it is associated with the second development of intermittent perennial ice cover on the ocean (Scott and others, 1988). Just before the Cobb Mountain event the microtine genus *Microtus* first appeared in the coldest early Pleistocene environment known from the Krestovka sections of Yakutia.

During the Jaramillo Subchron there probably was a slight warm spell, named the Bavelian in Holland, that is not clearly marked in oceanic cores 644 or CESAR-14. According to Zagwijn and de Jong (1984), the Bavelian consisted of a series of short warm and cold periods, and, although the ice-active periods are evident in core 644, the intervening warmer periods appear to have no time duration; they may have been too brief to be of recognizable duration in the plots of the core. *Microtus* first appeared in Holland, as advanced morphotypes, during the Bavelian; the genus represented a morphologically varied population in Yukon Territory during the Jaramillo Subchron (R.E. Morlan written commun., 1991).

Fifty thousand years after the end of the Jaramillo Subchron the first of several glacial advances that have been included in the "Nebraskan" glaciation, the first of the Ice Ages, spread across the north-central part of the contiguous United States, ending what is here called the second cooling cycle.

With this first ice advance of the Ice Ages the microtine genera *Lasiopodomys* and *Terricola* entered the United States from Beringia. Although *Lasiopodomys* had been a resident of Beringia for 250,000 years before the first advance of "Nebraskan" ice, similar to the genus *Allophaiomys* there is no Beringian record of *Terricola*. *Terricola* is and has always been a temperate-climate animal of Eurasian origin; thus it is presumed that at least one of the brief warm periods during or shortly following the Jaramillo Subchron must have approached a temperate climate in Beringia (Repenning, in press).

One further dispersal into the United States by microtine rodents has been mentioned, that of Beringian *Microtus*. This dispersal was believed to have occurred during isotope stage 12, about 450,000 years ago, on the basis of the climatic extreme suggested by isotope records, and this inference is weakly supported by age control of a similar dispersal in Eurasia (Repenning, 1984, 1987; Repenning and others, 1990); however, climatic extremes recorded in CESAR core 14 suggest that cold temperatures may have reached a maximum starting with isotope stage 8. Whether or not there is a correlation between these events has not been established. It might also be

noted that current information suggests that Beringian *Bison* did not enter the United States until isotope stage 6 (Repenning, 1987), so there may have been several immigrations during the late Ice Ages, stimulated by different cold periods in the Arctic Ocean Borderland.

This correlation of microtine biochronology, paleomagnetic signatures, and climatic trends as indicated by pollen and other records of the terrestrial flora and by marine fauna, as well as by the evidence of ice activity, has produced a chronology of events that we feel is convincing. There can be little question that between 3.0 and 2.0 million years ago the Arctic Ocean Borderland changed from a forested region to one in which forest-tundra and extensive tundra were dominant and in which permafrost first developed by 2.4 million years ago. During this same period the Arctic Ocean changed from cold-temperate to subfrigid; water temperatures became subfrigid about the time that tundra and permafrost developed in the latitudes of Beringia.

The end of this first cooling period is dramatically marked by the return of warm climate in the Northern Hemisphere. In the Arctic Ocean Borderland the climate was extremely warm, warmer than any other late Tertiary or Pleistocene climate so far recorded. With this warming the second cooling period began, and it culminated with the Ice Ages. The climate of the second cooling period fluctuated more than that of the first, with a return to warmer climates about every 100,000–300,000 years. Including the colder fluctuations, the warm period persisted, intermittently, for about a million years during most of the second cooling cycle. And the warmth that it began with was extreme in the conterminous United States, as well as in the Arctic Ocean Borderland.

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