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Pliocene Marine Transgressions of Northern Alaska: Circumarctic Correlations and Paleoclimatic Interpretations[1]

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ABSTRACT. At least three marine transgressions of Pliocene age are recorded by littoral to inner-shelf sediments of the Gubik Formation, which mantles the Arctic Coastal Plain of northern Alaska. The three recognized transgressions were eustatic high sea levels that, from oldest to youngest, are informally named the Colvillian, Bigbendian, and Fishcreekian transgressions. The geochronology is based upon amino acid geochemistry, paleomagnetic studies, vertebrate and invertebrate paleontology, and strontium isotope age estimates. Pollen, plant macrofossils, and marine vertebrate and invertebrate remains indicate that these trangressions occurred when the Arctic was at least intermittently much warmer than it is now. The Colvillian transgression took place at sometime between 2.48 and 2.7 Ma, when adjacent coastal areas supported an open boreal forest or spruce-birch woodland with scattered pine and rare fir and hemlock. The Bigbendian transgression occurred about 2.48 Ma. Climatic conditions were probably slightly cooler than during the Colvillian transgression, but probably too warm for permafrost and too warm for even seasonal sea ice in the region. Nearby vegetation was open spruce-birch woodland or parkland, possibly with rare scattered pine. The Fishcreekian transgression took place sometime between 2.14 and 2.48 Ma and was also characterized by warm marine conditions without sea ice. During the waning stages of this transgression, however, terrestrial conditions were relatively cool, and coastal vegetation was herbaceous tundra with scattered larch trees in the vicinity.

Other marine units from this time period occur around the Arctic Basin. The three oldest trangressions recognized from the Seward Peninsula may be broadly correlated with the three Pliocene transgressions of the Arctic Coastal Plain. The Tusatuvayam beds in Kamchatka possibly correlate with one of the two younger transgressions of northern Alaska. The non-marine Worth Point Formation of Banks Island may be younger than all three of the transgressions of the Arctic Coastal Plain, and marine sediment of the Beaufort Formation on Meighen Island is slightly older than the Colvillian transgression. None of the Pliocene marine units on Baffin Island can be confidently correlated with the high sea level events of northern Alaska. The upper Kap Kobenhavn Formation and the upper Loden Elv Formation of Greenland most likely correlate with the Fishcreekian transgression.

Key words: Arctic, amino acids, Pliocene, Pleistocene, paleoclimate, marine transgressions, sea level, Alaska, Gubik Formation

RÉSUMÉ. Au moins trois transgressions marines datant du pliocène sont inscrites dans les sédiments allant du littoral à l'intérieur de la plateforme de la formation Gubik, qui recouvre la plaine côtière arctique de l'Alaska septentrional. Les trois transgressions reconnues correspondent à des fortes remontées du niveau de la mer et ont reçu, dans l'ordre chronologique, les noms informels de formations «colvillienne», «bigbendienne» et «fishcreekienne». La géochronologie s'appuie sur la géochimie des acides aminés, des études paléomagnétiques, la paléontologie de vertébrés et d'invertébrés ainsi que sur des estimations de datation à l'isotope du strontium. Les pollens, les macrofossiles végétaux ainsi que les restes de vertébrés et d'invertébrés marins indiquent que ces transgressions se sont produites alors que l'Arctique était, pour le moins de façon intermittente, beaucoup plus chaud que maintenant. La transgression colvillienne a eu lieu à un moment donné entre 2,48 et 2,7 Ma, alors que les zones côtières adjacentes supportaient une forêt boréale ouverte ou des bois d'épinettes-bouleaux avec quelques pins éparpillés et de rares sapins et pruches. La transgression bigbendienne a eu lieu aux alentours de 2,48 Ma. Les conditions climatiques étaient probablement un peut plus froides que durant la transgression colvillienne, mais aussi probablement trop chaudes pour le pergélisol et en tout cas trop chaudes pour permettre la création d'une banquise – même saisonnière – dans la région. La végétation proche consistait en des bois ou des forêts-parcs d'épinettes-bouleaux avec peut-être quelques pins éparpillés. La transgression fishcreekienne a pris place à un moment donné entre 2,14 et 2,48 Ma et a aussi été caractérisée par des conditions marines chaudes sans banquise. Durant le déclin de cette transgression cependant, les conditions climatiques terrestres étaient relativement froides et la végétation côtière se composait de toundra herbacée semée de mélèzes aux alentours.

D'autres unités marines datant de cette période se trouvent autour du bassin de l'Arctique. Les trois plus anciennes transgressions établies dans la péninsule Seward peuvent être dans l'ensemble corrélées avec les trois transgressions du pliocène de la plaine côtière arctique. Les couches Tusatuvayam dans la Kamchatka sont peut-être à corréler avec l'une des deux transgressions les plus jeunes de l'Alaska septentrional. La formation non marine Worth Point de l'île de Banks est peut-être plus jeune que les trois transgressions de la plaine côtière arctique et les sédiments marins de la formation de Beaufort dans l'île Meighen sont légèrement plus anciens que la transgression colvillienne. On ne peut avec certitude corréler aucune des unités marines du pliocène sur l'île de Baffin avec les événements eustatiques qui ont amené une élévation du niveau marin dans l'Alaska septentrional. La partie supérieure de la formation Kap Kobenhavn et celle de la formation Loden Elv du Groenland sont probablement à corréler avec la transgression fishcreekienne.

Mots clés: Arctique, acides aminés, pliocène, pléistocène, paléoclimat, transgressions marines, niveau de la mer, Alaska, formation de Gubik

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РЕФЕРАТ. По крайне мере три трансгрессии моря Плиоценого возраста отмечаются литоральными к внутренными-шельфовыми осадками формации Губика, которой покрывает Арктическую береговую равнину северной Аляски. Эти три трансгрессии были евстатическими высокими уровнями моря, неофициально названными, от старого до молодого, Колвильская, Бигбендская, и Фишкрикская. Геохронология основана на геохимике аминовой кислоты, палеомагнетических исследованиях, позвоночной и непозвоночной палеонтологии, и стронций-изотопных оценках возраста. Пыльца, крупные ископаемые растений, морские позвоночные и непозвоночные остатки показывают, что эти трансгрессии происходили во время, когда темпуратура арктики была, по крайне мере прерывисто, более теплая чем на данное время. Колвильская трансгрессия случалась между 2,48 и 2,7 млн. лет, когда примыкающие береговые районы поддерживали открытый бореальный или ель-берёзовый лес с разбросанными соснами и редкими пихтами и болиголовами. Бигбендская трансгрессия происходила около 2,48 млн. лет. Климатические условия были вероятно слегка холоднее чем во время Колвильской трансгрессии, но вероятно слишком тепло для вечной мерзлоты, и слишком тепло для даже сезонного морского льда. Растительность была открытой, ель-берёзовой лесистой местностью или парковой местностью, может быть даже с редко разбросанными соснами. Фишкрикская трансгрессия случалась около 2,14 до 2,48 млн. лет и была характеризована теплыми морскими условиями без морского льда. Во время убываемых этапов трансгрессии, тем не менее, земные условия были относительно холоднее, и береговая растительность была травянистой тундрой с разбросанными лиственницами.

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Другие морские формации с этого времени всречаются около Арктического бассейна. Может быть, что три самые старые трансгрессии узнаваемые из полуострова Сюарда широко соотносительны к трем Плиоценовым трансгрессиям Арктической береговой равнины. Тусатуваемские толщи Камчатки возможно соотносятся к одной из двух молодых трансгрессий северной Аляски. Неморская формация Ворт Пойнта на острове Байнкса может быть моложе всех трёх трансгрессий арктической береговой равнины, и морской осадок формация Бофорта острова Мейгена слегка старше Колвильской трансгресии. Не одной из Плиоценовых трансгрессий острова Баффина не может быть определённо устанавливатся соотношение к случаям высокого уровня моря северной Аляски. Верхняя Кап Кобенхавенская формация и верхняя Лоден Елвская формация Гренландии наверно соотносятся к Фишкрикской трансгрессии.

Ключевые слова: арктика, аминовая кислота, плиоцен, плейстоцен, палеоклимат, морские трансгрессии, уровень моря, Аляска, формация Губика

Перевела для Арктика Соня Бенсон

INTRODUCTION

Three marine transgressions that occurred between 2.14 and 2.7 Ma in northern Alaska have been recognized through stratigraphic studies and analyses of amino acid diagenesis in marine mollusk shells collected from the lower part of the Gubik Formation of the Alaskan Arctic Coastal Plain (Fig. 1; Brigham, 1985; Carter *et al.*, 1986a). Sediments deposited during these transgressions contain information about relatively warm marine and terrestrial paleoenvironments that prevailed in northern Alaska following the establishment of essentially modern patterns of oceanic circulation that developed upon the opening of Bering Strait and the formation of the Isthmus of Panama. All of these transgressions have recently been interpreted as predating the first major late Cenozoic glaciation of the Northern Hemisphere *ca.* 2.4 Ma (Carter *et al.*, 1986b; Repenning *et al.*, 1987). However, the youngest of these transgressions may have occurred during interglacial conditions shortly after the first major glaciation.

In this paper, we discuss the paleoenvironmental conditions and paleogeographic significance of the deposits of these transgressions and attempt to relate them in time to late Neogene and early Pleistocene sediments at localities in Greenland, the Canadian Arctic Archipelago, western Alaska, and the Soviet Far East. These circumarctic correlations provide important information regarding the evolution of the climate and biota of the Arctic. As background for our discussion of the northern Alaska record, we 1) present information on

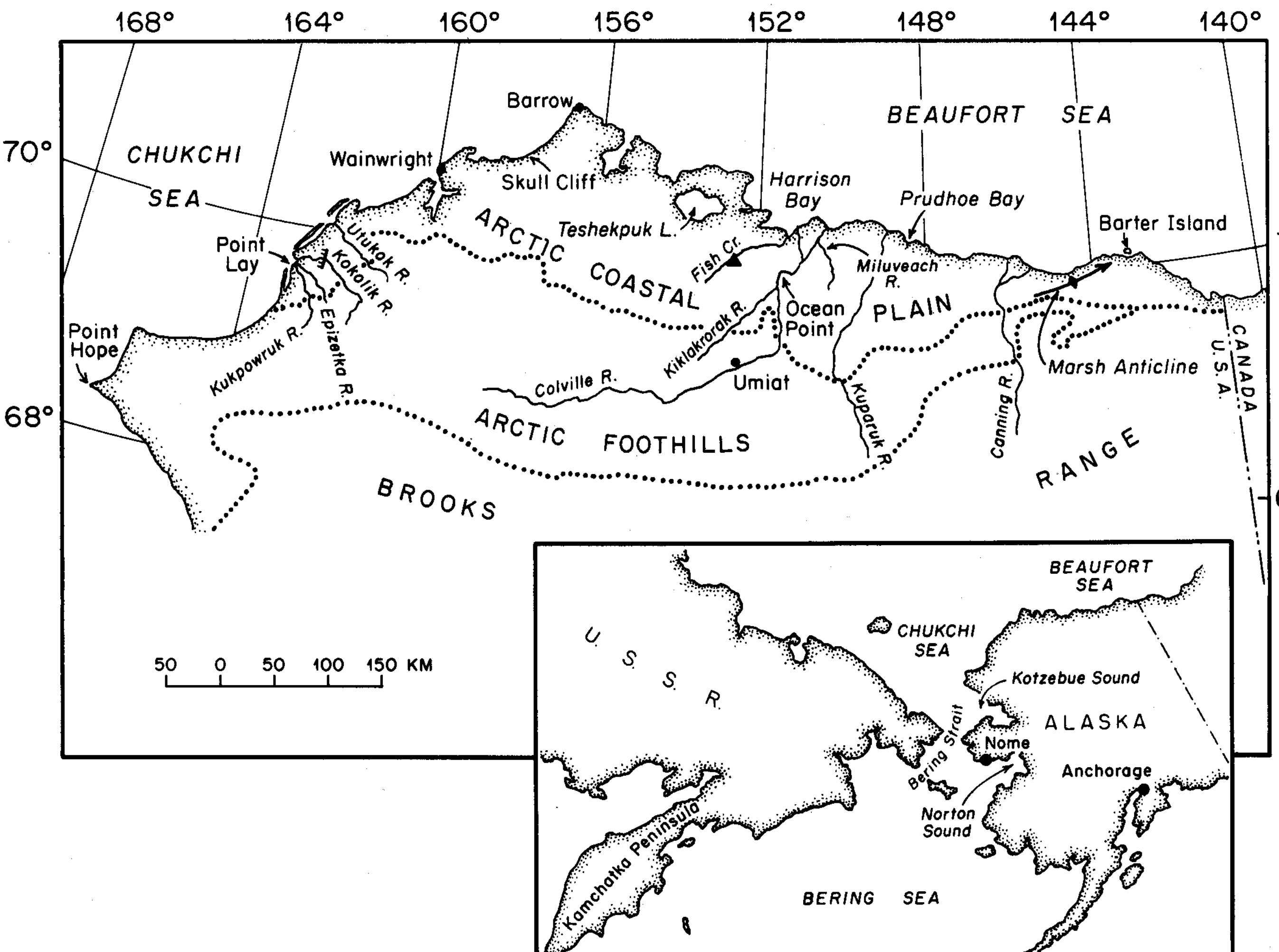

FIG. 1. Map showing localities in northern Alaska mentioned in the text. Triangle indicates locality on Fish Creek. The Gubik Formation covers the Arctic Coastal Plain. Physiographic provinces according to Wahrhaftig (1965).

the modern climate of the Arctic Coastal Plain as background for later discussions of paleoclimate, and 2) present the methods used to differentiate and correlate separate transgressions at individual sites.

NORTHERN ALASKA

Modern Climate and Permafrost

The climate of the Arctic Coastal Plain is characterized by long, dry, cold winters and short, moist, cool summers. Details of the climate can be found in Haugen and Brown (1980), Dingman *et al.* (1980), and Searby and Hunter (1971). Mean annual temperatures across the coastal plain range between −10°C at Point Lay at the western edge of the coastal plain to −12.7°C at Barter Island on the eastern coastal plain (Nelson, 1979). The annual total precipitation, as measured by standard precipitation gauges, ranges from 10 to 18 cm, of which about 35-50% falls as rain. However, because of the high winds common during snowfall events, precipitation amounts, especially for autumn and winter, may have been underestimated by 100-400% (Black, 1964; Benson, 1982).

The Arctic Coastal Plain lies in the region of continuous permafrost, the southern boundary of which is just south of the Brooks Range at approximately the −6°C isotherm (Gold and Lachenbruch, 1973). Permafrost beneath the coastal plain varies in thickness from 200 to 400 m near Barrow to about 650 m near Prudhoe Bay (Pewé, 1975; Osterkamp and Payne, 1981). Subsea permafrost, in part relict, is also present beneath the recently submerged continental shelf of the Beaufort Sea (Sellman and Hopkins, 1983). During the summer months, the active layer typically thaws to a depth of only 30-50 cm, which impedes drainage and produces the wet to moist tundra of the region. Ubiquitous ice wedges, thaw lakes, pingos, and localized solifluction lobes are manifestations of this frigid environment.

Most relevant for amino acid diagenesis are past and present ground temperatures. The mean annual temperature of the permafrost near Prudhoe Bay is about −10°C at a depth of 18.5 m, where the seasonal temperature wave is nearly completely damped out (Lachenbruch *et al.*, 1962). Brigham and Miller (1983) have shown that below a depth of 3 m the effect of the seasonal temperature wave on rates of isoleucine epimerization is essentially less than the analytical precision of the amino acid method. At the sites discussed here, the marine deposits of interest are buried by 5 m or more of younger sediments.

The Gubik Formation: Identifying and Correlating Marine Transgressions

The Gubik Formation is of Pliocene and Pleistocene age and includes all of the unconsolidated sediments (marine, fluvial, lacustrine, and eolian) of the Alaskan Arctic Coastal Plain (Black, 1964; Repenning, 1983; Dinter, 1985; Carter *et al.*, 1986a; Fig. 1). A brief history of investigations and interpretations of the Gubik Formation is given in Brigham (1985) and in Dinter *et al.* (1990). Sediments of the Gubik Formation overlie a marine abrasion platform cut across Cretaceous and Tertiary rocks that is an emergent extension of the continental shelves of the adjacent Beaufort and Chukchi seas. Gubik Formation sediments are also preserved on the Beaufort shelf but have not been thoroughly studied (Dinter, 1985; Dinter *et al.*, 1990; McDougall *et al.*, 1986; McNeil, 1990; Wolfe *et al.*, 1985, 1986; Smith, 1986). Recent investigations (e.g., Williams, 1979, 1983a,b; Brigham, 1985; Carter and Galloway, 1985) of the emerged marine sediments on the western part of the coastal plain resulted in the identification of six marine transgressions (Table 1; Carter *et al.*, 1986a).

Marine sediments of the Gubik Formation are distinguished in the field from those of other sedimentary environments by their physical character and macrofossil content. During our investigations, stratigraphic sections of marine deposits were carefully measured, and bounding unconformities were identified wherever possible. However, superposed shallow marine deposits of widely differing ages sometimes are lithologically indistinguishable, and shoreface erosional surfaces separating them are not always obvious as unconformities. Additionally, marine facies of contrasting physical appearance, but formed during the same marine incursion, may be separated by sharp boundaries that could be mistaken for significant unconformities. Because of this, we used amino acid geochronology in fossil mollusks to differentiate the sediments of successive marine transgressions at each locality. Also, due to the relatively uniform modern climate and ground temperatures across the coastal plain, and because sediments of the same age have been affected by the same sequence of climatic changes and ground temperature variations, amino acid geochronology was used to correlate sediments in disjunct stratigraphic sections across the entire coastal plain.

Amino acid geochronology is based on the extent of epimerization of L-isoleucine (Ile) to D-alloisoleucine (aIle), expressed as a ratio (aIle/Ile), in both the Total acid hydrolysate (free plus peptide-bound amino acids) and the Free (naturally hydrolized) amino acid fraction. Shells of the common pelecypods

TABLE 1. Marine transgressions and amino acid ratios, Alaskan Arctic Coastal Plain

	aIle/Ile in *Hiatella arctica*[1]						
	Skull Cliff			Colville River area			
Transgression	Total	Free		Total	Free		Age Estimate
Simpsonian	—	—		—	—		>50, <90 ka
Pelukian	0.014 ± 0.003	ND[2]	(20)[3]	0.0165 ± 0.0005	ND	(3)	120-130 ka
Wainwrightian	0.038 ± 0.007	0.40 ± 0.052	(28)	—	—		400-500 ka
Fishcreekian	0.090 ± 0.018	0.52 ± 0.040	(11)	0.085 ± 0.010	0.48 ± 0.06	(10)	>2.14, <2.48 Ma
Bigbendian	0.15 ± 0.025	0.58 ± 0.08	(8)	0.126 ± 0.016	0.67 ± 0.06	(23)	2.48 Ma
Colvillian	0.235 ± 0.017	0.75 ± 0.07	(4)	0.265 ± 0.025	0.70 ± 0.08	(21)	>2.48, <2.7 Ma

[1]Ratios expressed as mean ± 1 standard deviation.
[2]ND means not detectable; no measureable aIle.
[3]Number in parentheses is number of shells analyzed.

Hiatella arctica, Mya truncata, Astarte spp., *Macoma* spp., and *Portlandia arctica* were used because they have yielded consistent results in previous arctic studies and are relatively common in deposits along arctic and subarctic coasts (Miller, 1985; Miller and Brigham-Grette, 1989). Over 500 shells from the Gubik Formation have been analyzed over the last ten years at the Institute of Arctic and Alpine Research (INSTAAR), Boulder, Colorado. Recent reviews of the geologic application of this method include Wehmiller (1984, 1986, 1989), Miller (1985), and Miller and Brigham-Grette (1989).

Pliocene and Early Pleistocene Marine Transgressions

Of the six marine transgressions presently recognized, the three oldest occurred during late Pliocene time and, from oldest to youngest, are informally named the Colvillian, Bigbendian, and Fishcreekian transgressions (Table 1; Carter *et al.*, 1986a,b). The most important exposures of these sediments are in coastal bluffs along Skull Cliff, in bluffs along the lower Colville River, and at Fish Creek (Fig. 1). Deposits of all three transgressions occur in superposition at Skull Cliff (Fig. 2); Colvillian and Bigbendian beds are exposed in superposition in the Colville River bluffs; and Bigbendian and Fishcreekian strata are in superposition at an important exposure on Fish Creek. At each locality, amino acid ratios clearly distinguish the deposits of each transgression and allow correlation of the beds from site to site. Summary descriptions of the deposits of each transgression are presented here.

Colvillian: The Colvillian transgression is based on marine deposits in the Colville River area that contain *Hiatella arctica* shells with Total aIle/Ile ratios of 0.265 ± 0.025 (Table 1). Colvillian sediments are well exposed in bluffs along the Colville River from its confluence with the Kikiakrorak River north for about 13 km and upstream along the Kikiakrorak River for 5 km from the confluence (Fig. 1). Throughout these exposures the Colvillian deposits unconformably overlie Cretaceous strata. They are overlain locally by 1-8 m of Bigbendian deposits and everywhere by 11-12 m of much younger fluvial and eolian sediment. The Colvillian beds reach altitudes of at least 40 m and consist of gravelly sand up to 2.5 m thick that locally is overlain by up to 2.5 m of silty clay or clayey silt. Cobbles and small boulders from the Colvillian basal beds include sandstone, quartzite, and chert-pebble conglomerate clasts, some of which are striated. Similar-sized clasts of these rock types are common in the Kuparuk gravel unit (informal name) of Tertiary age east of the Colville River, and clasts in Bigbendian deposits could have been derived from this older unit by erosion (Carter and Galloway, 1985). Cobble- to boulder-sized clasts of metamorphic, intrusive, and volcanic lithologies occur as float on talus slopes below some exposures of Colvillian and Bigbendian beds, and an erratic of augen gneiss with an exposed long dimension of 1 m occurs in the bed of the Miluveach River adjacent to an exposure whose base is composed of Colvillian or Bigbendian sediments. Most of these rock types occur in Paleocene beds that underlie the Gubik Formation north of Ocean Point (Frederiksen *et al.*, 1988; Carter and Galloway, 1985), but no boulder-sized clasts have been observed in the Paleocene beds, and thus clasts of this size may be erratics.

Other sediments deposited during the Colvillian transgression, as determined by amino acid geochemistry and superposition, are exposed east of the Colville River along the Miluveach River, on the north and south flank of Marsh Anticline on the eastern part of the coastal plain, and at Skull Cliff along the Chukchi Sea coast (Figs. 3 and 4), where they have been called the Nulavik unit (Brigham, 1985). The Nulavik unit mostly represents an inner shelf facies and occurs as thin patches of basal cobbly gravel, cross-bedded ripply sand, and silty sand with interbedded sand and silt. The sediments lie unconformably on either sandstone of the Cretaceous Nanushuk Group or the Papigak clay unit (Brigham, 1985), the latter of which appears to correlate with the Seabee Formation of the Cretaceous Colville Group (Will Elder, U.S. Geological Survey, written comm. March 1990; Arthur Grantz, U.S. Geological Survey, written comm. March 1990). *Hiatella arctica* shells from these beds yield Total aIle/Ile ratios of 0.23 ± 0.017, nearly identical to those of similar shells from the Colville River exposures. Cobbles and boulders of sandstone, red quartzite, chert, and granite occur at the base of and within the Nulavik unit. There is no possible nearby source for the red quartzite and granite clasts, and we interpret them as erratics. Some of the granite clasts appear lithologically similar to granite that composes erratics of Canadian shield provenance in the late Pleistocene Flaxman Member of the Gubik Formation (Dinter, 1985; Rodeick, 1979; MacCarthy, 1958), but others do not, and the overall assemblage of rock types is unlike that of the Flaxman Member. The presence of erratics in Bigbendian sediments at Skull Cliff strengthens the possibility that the boulder-sized clasts in the Colville River area are also erratics.

Paleomagnetic measurements on Colvillian sediments at Skull Cliff and the Colville River area have not provided conclusive evidence for magnetic polarity at the time of deposition.

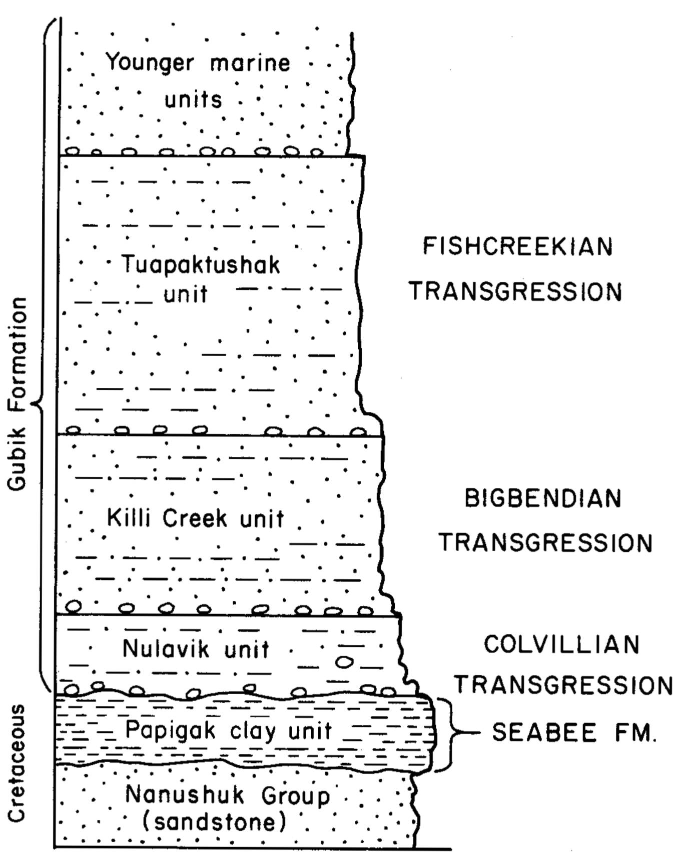

FIG. 2. Schematic stratigraphic column from Skull Cliff showing deposits of the Colvillian, Bigbendian, and Fishcreekian transgressions in superposition. Thickness of each unit varies from section to section. Textural variations are described in text.

Despite demagnetization, the measurements yield a mixture of steeply dipping negative (reversed) and positive (normal) inclinations, as well as very low-angle inclination values, characteristics of sediments that have been overprinted with chemical remanent magnetism (Brigham, 1985; V.L. Pease, U.S. Geological Survey, written comm. 1985).

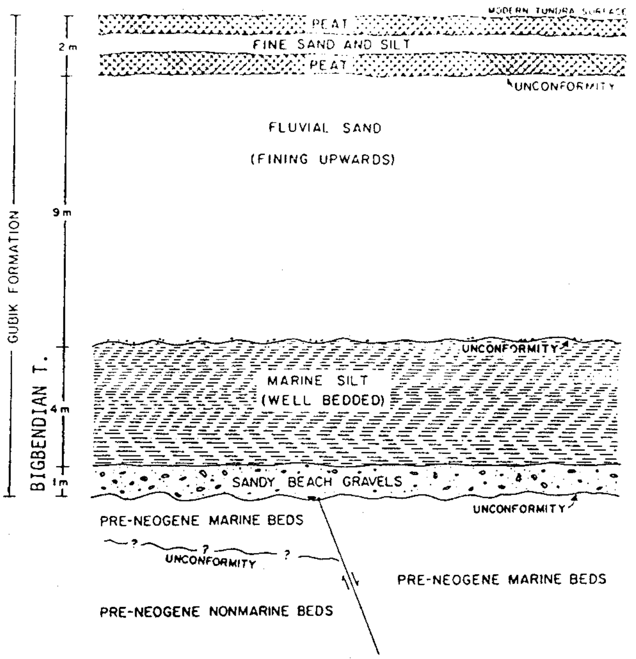

FIG. 3. Schematic stratigraphy of Bigbendian transgression sediments near Ocean Point (modified from Nelson and Carter, 1985).

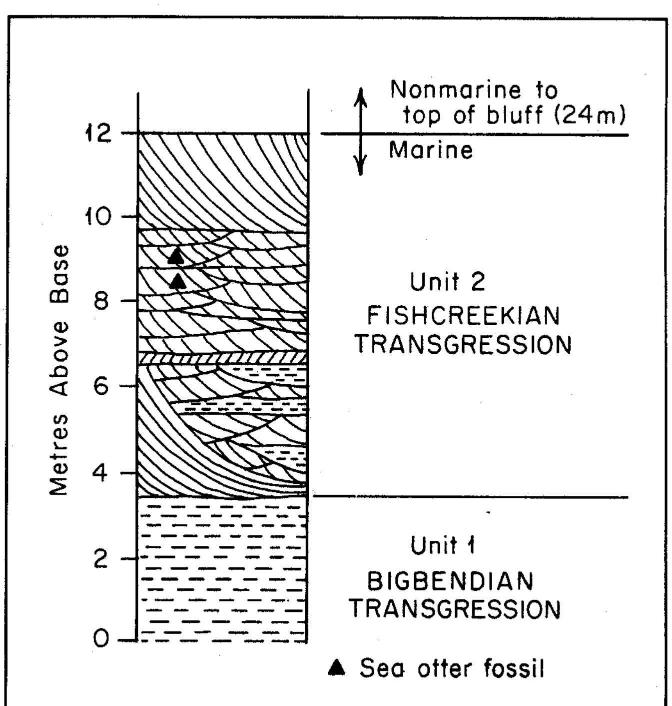

FIG. 4. Generalized marine stratigraphy at Fish Creek and location of sea otter fossils (modified from Carter *et al.*, 1986b).

Bigbendian: The Bigbendian transgression is based on deposits exposed for about 10 km in bluffs along the big bend of the Colville River near Ocean Point and the Big Bend benchmark, where they occur at altitudes up to about 35 m (Carter *et al.*, 1986a). The Bigbendian beds here generally consist of basal, transgressive, gravelly sand about 1 m thick overlain by 4-7 m of sandy to pebbly silt that ranges from well bedded to massive (Fig. 3). The basal bed contains cobbles and boulders like those described for Colvillian deposits in the Colville River bluffs. *Hiatella arctica* shells from these deposits yield Total aIle/Ile ratios of 0.126 ± 0.016 (Table 1). Bigbendian beds in places rest unconformably on Cretaceous and early Tertiary strata but elsewhere are separated from them by a few metres of Colvillian deposits.

Bigbendian beds near Ocean Point deposited in the early part of the transgression exhibit normal-magnetic polarity (Carter *et al.*, 1986b). A normal to reversed-polarity boundary occurs in Bigbendian beds about 10 km upstream from Ocean Point, and sediments of the basal unit at Fish Creek that can now be confidently correlated with the Bigbendian transgression on the basis of amino acid diagenesis (Table 2) are reversely magnetized (Carter *et al.*, 1986b: Carter and Hillhouse, 1991).

Bigbendian deposits are also exposed at altitudes up to about 20 m at Skull Cliff, where they are known as the Killi Creek unit (Fig. 2; Brigham, 1985), and along the Miluveach River. At Skull Cliff their distribution is patchy and discontinuous due to extensive erosion during younger transgressive events. The Killi Creek unit represents an inner shelf facies and consists of gravel, sand, silt, silty sand, and some basal gravel lags containing cobble-sized clasts. At one locality, fine-grained sediments at the base of the unit contain boulders of chert and quartzite that may be dropstones. Locally the horizontal bedding is contorted and deformed into large isoclinal and overturned folds with an amplitude over 4 m high, perhaps by iceberg scour or soft sediment slides. At present the Killi Creek unit can be distinguished from the Nulavik unit in the Skull Cliff section only by their contrasting aIle/Ile ratios from the enclosed mollusks. *Hiatella arctica* shells from the Killi Creek beds yielded Total aIle/Ile ratios of 0.150 ± 0.025 (based on eight individuals), significantly lower than those determined on similar shells from the Nulavik unit, but statistically indistinguishable from those for Bigbendian shells from the Colville River area.

Fishcreekian: The Fishcreekian transgression is based on fossiliferous marine beds exposed along the north side of Fish Creek (Fig. 4: unit 2; Carter *et al.*, 1986b). This unit, interpreted to be a tidal channel deposit, consists of 5-9 m of predominantly trough cross-bedded, fossiliferous, brown to gray sand, pebbly sand, and silt, containing detrital wood and organic debris. *Hiatella arctica* shells from those sediments yield Total aIle/Ile ratios of 0.085 ± 0.010 (Table 1). Fishcreekian deposits at this location are overlain by up to 17 m of younger nonmarine sediments. The tidal channel deposits

TABLE 2. Amino acid ratios for *Portlandia arctica*, Ocean Point and Fish Creek

	Total aIle/Ile			No. of shell
Locality	Min	Max	Mean & sd	analyses
Bigbendian sediments near Ocean Point	0.120	0.160	0.136 ± 0.015	5
Lower marine unit Fish Creek (Fig. 1)	0.143	0.162	0.155 ± 0.007	5

record a paleo-sea level that now lies about 25 m higher than modern sea level (Carter and Galloway, 1985). This is also the altitude of the base of a scarp 30 km east of Fish Creek and 5 km west of the Colville River delta (Fig. 1). The scarp truncates a terrace beneath which Bigbendian and Colvillian deposits are preserved (Terrace I of Carter and Galloway, 1982) and was interpreted to have formed during the Fishcreekian transgression (Carter *et al.*, 1979). However, some reworked marine mollusk shells from fluvial sediments that overlie Colvillian and Bigbendian beds on this terrace yield amino acid ratios characteristic of the Fishcreekian transgression, as do some mollusks from the inner edge of the coastal plain at an altitude of about 70 m (L.D. Carter, unpubl. data). Thus the scarp seems unlikely to represent the limit of the Fishcreekian transgression, and the tidal channel deposits at Fish Creek appear to have formed during the regressive phase of this marine incursion.

The sparsely fossiliferous lower unit exposed at Fish Creek (Fig. 4: unit 1) was initially interpreted as Fishcreekian in age (Carter and Galloway, 1985) and later as Colvillian (Brouwers, 1987). However, recent amino acid analyses on *Portlandia arctica* from these deposits indicate instead that they are Bigbendian (Table 2).

Sediments correlated with the Fishcreekian transgression on the basis of their amino acid geochemistry also occur at Skull Cliff on the western coastal plain, where they are known as the Tuapaktushak unit (Brigham, 1985). These beds consist largely of inner-shelf facies of interbedded sand, silty sand, silt, and some clay and pebbly layers overlying a basal gravel lag. Erratic cobbles and boulders, commonly of the same rock types as erratics of Canadian Shield provenance in the Flaxman Member of the Gubik Formation, are dispersed throughout the unit but are most common at the base. Severely contorted bedding, similar to that seen in the Nulavik unit, is common in these deposits (Fig. 5). The origins of these structures are unknown, but possible causes are iceberg scouring and low-gradient sliding. *Hiatella arctica* shells from the Tuapaktushak unit yield Total aIle/Ile ratios of 0.090 ± 0.018 (Table 1).

Also on the western coastal plain, Fishcreekian beds have been mapped south of the Kokolik River to an elevation of 33-36 m above sea level, where they intersect the base of a prominent break in slope that is presumably a wave-cut scarp (Fig. 1). Four to five km seaward of this break in slope, between the

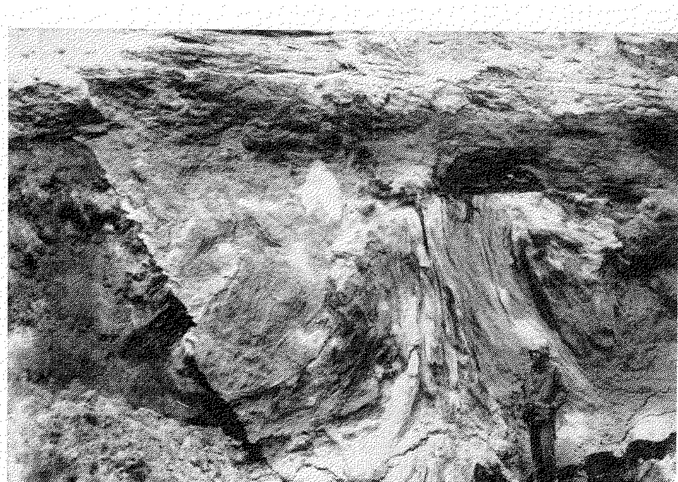

FIG. 5. Complex isoclinal folding that occurs locally in Fishcreekian transgression sediments at Skull Cliff. Man is 1.9 m tall.

Kukpowruk and Epizetka rivers, are beach ridges (*cf.* McCulloch, 1967) thought to represent contemporaneous barrier islands (Brigham, 1985). Fishcreekian sediments also occur as uplifted beds on the north flank of Marsh Anticline on the eastern coastal plain (Fig. 1).

Measurements of paleomagnetism on Fishcreekian sediments at Fish Creek indicate reversed polarity (Carter *et al.*, 1986b). The original magnetic signature of Fishcreekian sediments at Skull Cliff, however, apparently has been overprinted by chemical remanent magnetization; measurements there yield a mixture of reversed and normal polarities, as well as very low-angle inclination values that cannot be used to determine polarity (Brigham, 1985). At least part, and probably all, of the Fishcreekian transgression occurred during a period of reversed-magnetic polarity.

Age Estimates

Colvillian: Colvillian mollusk faunas include taxa of Pacific origin and thus post-date the opening of Bering Strait, which occurred about 3.2 Ma (Hopkins, 1967, 1972; Gladenkov, 1981). A minimum age for the Colvillian transgression is provided by the younger Bigbendian sediments, thought to span the Gauss Normal-Polarity Chron-Matuyama Reversed-Polarity Chron boundary (Carter and Hillhouse, 1991). The age of this boundary is 2.48 Ma, according to the Geomagnetic Polarity Timescale of Mankinen and Dalrymple (1979). The difference in the amount of isoleucine epimerization between Colvillian and Bigbendian mollusks suggests a significant difference in age, perhaps as much as 1 m.y., if permafrost was present during the interval between the two transgressions so that ground temperatures were similar to modern values. However, if permafrost was absent, this amount of epimerization could have been accomplished in less than 0.1 m.y. Inasmuch as the earliest unequivocal record of permafrost in lowland continental areas of the Northern Hemisphere is between 2.0 and 2.5 Ma (Sher *et al.*, 1979), the Colvillian transgression most likely occurred no more than 2.6 or 2.7 Ma. Thus, we estimate that the Colvillian transgression occurred sometime between 2.48 and 2.7 Ma. This estimate could be refined with better paleomagnetic data and further analyses of foraminifer faunas.

Bigbendian: Strontium isotope analyses suggest a minimum age of 1.9 Ma for the Bigbendian transgression. Repenning (1983) suggested a possible age of between 1.7 and 2.6 Ma for the Bigbendian deposits near Ocean Point based on the stage of evolution exhibited by the fossil sea otter remains from that locality. However, inasmuch as the sea otter indicates markedly warmer water in the Arctic Ocean than today, Repenning considered the period from 2.2 to 2.6 Ma an unlikely time for the Bigbendian transgression, because the first major continental ice accumulation in the northern hemisphere apparently occurred between about 2.2 and 2.4 Ma (Shackleton and Opdyke, 1977; Raymo *et al.*, 1989). Carter and Galloway (1985) considered it more probable that the Bigbendian transgression preceded this climatic deterioration, a conclusion now also accepted by Repenning *et al.* (1987). If this conclusion is correct, then the polarity change of normal to reversed that occurred during the Bigbendian transgression is most likely the 2.48 Ma Gauss-Matuyama boundary (Carter and Hillhouse, 1991).

Fishcreekian: Controversy surrounds discussion of the age of the Fishcreekian transgression. Age estimates have ranged from 1.0 to 2.41 Ma.

Based upon broad assumptions concerning the Pleistocene temperature history of the Arctic Coastal Plain, Brigham (1985) suggested that the extent of isoleucine epimerization in Fishcreekian mollusks indicated an age of between 1 and 1.4 Ma, an age consistent with reversed magnetic polarity. Carter *et al.* (1986b) proposed an age of about 2.41 Ma for three reasons:

1) They accepted Repenning's initial interpretation that microtine rodents in the Fishcreekian fauna indicated an older age than that of the Cape Deceit fauna, which Repenning placed at 2.0-2.4 Ma (C.A. Repenning, U.S. Geological Survey, written comm. 1985).

2) A maximum limiting age of 2.48 Ma was indicated by the reversed magnetic polarity when coupled with the fact that the earliest known occurrence of lowland tundra flora, according to Wolfe (1985), is in the upper part of the Kutuyakh beds of the Kolyma lowland, northeastern Siberia. The Kutuyakh beds are thought to span the Gauss-Matuyama boundary (Sher *et al.*, 1979).

3) The Fishcreekian transgression probably corresponds to some ^{18}O minimum in deep-sea sediment cores. Carter *et al.* (1986b) therefore suggested an age of 2.41 Ma, because that is the age of the only significant ^{18}O minimum within the Matuyama Chron older than 2.4 Ma (Shackleton *et al.*, 1984).

Repenning *et al.* (1987) also suggested an age of about 2.4 Ma for the Fishcreekian beds but acknowledged that the microtines from the Fishcreekian beds do not in themselves provide definitive evidence for an age even as great as the Cape Deceit fauna (Repenning *et al.*, 1987:23-24). Instead they used arguments based on climatic history and geopolarity to support their age assignment. Moreover, the Cape Deceit fauna is now thought by Matthews and Ovenden (1990) to be much younger than 2.0 Ma, based upon interpretation of floral assemblages and the fact that *Phenacomys gryci*, one of the microtines from the Fishcreekian beds thought to indicate great antiquity, is now known from sediments in the Old Crow region of Canada (C.A. Repenning, U.S. Geological Survey, written comm. to J.V. Matthews, 1990). These sediments can be no older than the Jaramillo Normal-Polarity Subchron (0.97-0.90 Ma) of the Matuyama and may have formed during the Brunhes Normal-Polarity Chron, which began about 0.73 Ma (Matthews and Ovenden, 1990).

Strontium analyses of Fishcreekian mollusks suggest an age of 0.5-1.7 Ma (Kaufman *et al.*, 1990), which brackets Brigham's (1985) original age estimate based on the amino acid data. More recent analyses indicate that the strontium content of some Fishcreekian mollusk shells has been diagenetically altered, and so we cannot be confident of these age limits (L.D. Carter and G.L. Farmer, unpubl. data).

A tooth of a primitive sea otter collected from the Fishcreekian beds at Fish Creek suggests a pre-Olduvai age for the transgression. Repenning *et al.* (1987) state that the tooth does not resemble that of the living sea otter, *Enhydra lutris*, and is more like the ancestor *Enhydriodon* than is the single known tooth of *Enhydriodon? reevi* from the Norwich Crag in East Anglia, which was deposited during the Olduvai Normal-Polarity Subchron of the Matuyama (Zalasiewicz and Gibbard, 1988). If this difference is of evolutionary significance, and if the tooth from the Fishcreekian beds is not redeposited, it suggests that the reversely magnetized Fishcreekian sediments were deposited prior to the Olduvai, or before 1.87 Ma. Furthermore, teeth of the microtine *Plioctomys mimomiformis* occur in these same beds, and the youngest occurrence of this genus elsewhere is in beds older than the oldest part of the Reunion Normal-Polarity Subchron of the Matuyama (C.A. Repenning, U.S. Geological Survey, written comm. 1990), which began 2.14 Ma. The Fishcreekian transgression most likely occurred prior to the Reunion Subchron, between 2.14 and 2.48 Ma.

Paleoclimatic Interpretations

Colvillian: The few data available regarding climatic conditions during the Colvillian transgression indicate that climate was much warmer than at present. Pollen analyses of five samples from Colvillian sediments near Ocean Point and on the Kikiakrorak River (Fig. 1) suggest that nearby vegetation was open boreal forest or spruce-birch woodland with scattered pine and rare fir and hemlock (Nelson and Carter, 1991). Hemlock is not present in Bigbendian pollen floras (Nelson and Carter, 1985), and the Colvillian forest had a more closed aspect than the Bigbendian forest or parkland. These differences in the composition and character of the vegetation suggest that climate during the Colvillian transgression may have been slightly warmer than during the Bigbendian transgression, which was characterized by an absence of sea ice and probably also permafrost (see below). Warm marine conditions are confirmed by the general character of the ostracode fauna, which includes *Pterygocythereis vannieuwenhuisei* Brouwers (Brouwers, 1987; Table 3), an extinct species of a genus whose modern northern limit is the Norwegian Sea but in the northwestern Atlantic Ocean does not occur north of the southern cold temperate zone (Brouwers, 1987). Erratics in Colvillian beds at Skull Cliff indicate that glaciers were present somewhere around the margin of the Arctic Basin during this transgression.

At the Colville River, a femur of the North Atlantic harp seal, *Pagophilus groenlandica* (Erxleben), was found as float (Repenning, 1983) and could have been derived from either Colvillian or Bigbendian deposits. The harp seal fossil may have paleoenvironmental significance because modern harp seals give birth only on sea ice. The presence of sea ice seems inconsistent, however, with the warm terrestrial and marine conditions indicated by the pollen flora and ostracode fauna of

TABLE 3. Occurrence of some extinct and extralimital taxa of the lower part of the Gubik Formation

	Status or modern northern limit	Cv	Bb	Fc
Mollusks				
Neptunea lyrata leffingwelli	extinct	X	X	X
Neptunea cf. N. beringiana	S. Chukchi Sea		X	X
Astarte leffingwelli	extinct	X	X	X
Aforia circinata	S. Bering Strait		X	X
Littorina squalida	S. Bering Strait		X	X
Ademete regina	S. Chukchi Sea		X	X
Natica (Tectonatica) janthostoma	N.W. Pacific			X
Clinocarium californiense	S. Bering Strait		X	X
Zirfaea pilsbryi	S. Bering Strait			X
Ostracodes				
Rabilimus paramirabilis	extinct	X	X	X
Pterygocythereis vannieuwenhuisei	extinct	X		
Vertebrates				
Enhydriodon? fossil sea otter	extinct		X	X
Synaptomys (Plioctomys) bog lemming	extinct			X
Phenacomys gryci, heather vole	extinct			X

Cv = Colvillian, Bb = Bigbendian, Fc = Fishcreekian.

the Colvillian transgression and is certainly inconsistent with the strong faunal evidence for an absence of sea ice at the height of the Bigbendian transgression (see below). Climatic deterioration probably accompanied the regressive phase of each marine incursion and was the most likely time for the formation of sea ice.

Bigbendian: Evidence for paleoclimate during the Bigbendian transgression was presented in R.E. Nelson (1981), Repenning (1983), Nelson and Carter (1985), Carter *et al.* (1986b), and Repenning *et al.* (1987) and is summarized here. A relatively mild climate during the Bigbendian transgression is suggested by the presence of sea otter remains near Ocean Point (Repenning, 1983) and by a mollusk fauna that is richer than the presently known Colvillian fauna and includes the gastropod *Littorina squalida* and the bivalve *Clinocardium californiense* (Deshayes) (Carter *et al.*, 1986b). The modern northern limit of both these mollusk taxa is south of Norton Sound. Moreover, modern sea otters cannot tolerate severe seasonal sea ice conditions (Schneider and Faro, 1975); hence, the presence of sea otter remains suggests that the limit of seasonal ice on the Beaufort Sea was north of the Colville River fossil site during the Bigbendian transgression (Carter *et al.*, 1986b). Erratic clasts at Skull Cliff, if not reworked from underlying Colvillian beds, indicate that glaciers were present somewhere around the margin of the Arctic Basin.

Permafrost probably was not present, or was at most discontinuous and limited to north-facing slopes. This is based on a pollen suite from the Colville River section, which indicates that the nearby coastal plain probably supported an open spruce-birch woodland, or even parkland, with scattered pine possibly present in favorable sites (R.E. Nelson, 1981; Nelson and Carter, 1985, 1991).

Fishcreekian: Paleoclimate during the Fishcreekian transgression was discussed in Carter *et al.* (1986b) and Repenning *et al.* (1987). Sea-surface temperatures during this transgression were warmer than at present, as indicated by the presence in the Fish Creek beds of the bivalve *Clinocardium californiense*, whose modern northern limit is Norton Sound, and the gastropods *Aforia circinata* and *Littorina squalida*, which presently range no farther north than the Bering Sea, and also by the presence in the Tuapaktushak unit of the gastropod *Natica (Tectonatica) janthostoma* (Broderip and Sowerby), which is presently limited to the waters adjoining Japan, Kamchatka Peninsula, and the Commander Islands (Carter *et al.*, 1986b). Collections from the Tuapaktushak unit also include the extralimital elements *Clinocardium californiense* and *Aforia circinata*, along with extinct forms including *Neptunea lyrata leffingwelli* (Dall) and *Astarte leffingwelli* (Dall) and the extinct ostracode *Rabilimus paramirabilis* Swain (E.M. Brouwers, U.S. Geological Survey, written comm. 1986). Sea otter remains and the intertidal gastropod *Littorina squalida* at Fish Creek suggest that perennial sea ice was absent or severely restricted during the Fishcreekian transgression (Carter *et al.*, 1986b).

In contrast to the relatively warm marine conditions, the late Fishcreekian terrestrial climate was apparently harsh. Pollen assemblages from the tidal-channel deposits of unit 2 (Fig. 4) at Fish Creek, which represent the regressive phase of this high sea level stand, are similar to those from the late Pliocene or early Pleistocene Cape Deceit Formation of Matthews (1974) of the northern Seward Peninsula (Giterman *et al.*, 1982) and suggest that the nearby coastal plain then supported wetland herbaceous tundra with larch trees in the vicinity (Repenning *et al.*, 1987). Dropstones of probable Canadian shield provenance observed in the Tuapaktushak unit at Skull Cliff suggests a moderate buildup of Laurentide ice during or immediately prior to the Fishcreekian transgression.

CIRCUMARCTIC CORRELATIONS

Several sites around the arctic basin contain evidence relevant to our understanding of the Pliocene and Pleistocene evolution of arctic climate (Fig. 6). The gradual late Tertiary decline in global temperatures, as inferred from isotopic trends in benthic and planktonic foraminifers (Lloyd, 1984), is reflected in the Arctic by a decrease in vegetation diversity and the southward shift in the forest-tundra ecotone (Matthews and Ovenden, 1990).

In addition to paleobotanical and biostratigraphic data, sites containing marine mollusks offer an additional means of evaluating the relative ages of these widely separated sites through amino acid geochronology. The biggest challenge to this method, however, is trying to interpret temperature-dependent data across more than 20° of latitude and from sites with modern mean annual temperatures that differ by as much as 17°C. As shown by Brigham and Miller (1983), Miller and Mangerud (1985), and Miller and Brigham-Grette (1989), rates of isoleucine epimerization are severely retarded at temperatures below 0°C, decreasing the stratigraphic resolution of the method but at the same time increasing the geologic time span over which it is useful for relative age estimates.

In the following section we have compiled all of the published amino acid data on shell material from arctic and subarctic sites believed to date between 1 and 3.5 Ma (Table 4). Stratigraphic relationships, paleomagnetic data, micro- and macrofaunal characteristics, and radiometric numerical ages provide a basis for comparing these data to amino acid data from the three oldest marine transgressions of the Gubik Formation.

Seward Peninsula and Bering Sea

The marine transgressions of the Gubik Formation may be correlative with the transgressions documented by Hopkins (1967, 1973) along the western coast of Alaska. This was first suggested on the basis of paleontological criteria (Hopkins, 1967) and then by preliminary amino acid work in the late 1970s by G.H. Miller and Hopkins (summarized in Brigham, 1985). More recently, revision of the local stratigraphy has shown that the Pliocene Beringian transgression of Hopkins (1967) can be subdivided on the basis of amino acid data into three marine transgressions of different ages (Kaufman *et al.*, 1989; Kaufman, 1991; Table 4). Using these data, the transgressions, recognized as Beringian I, II, and III at Nome by Kaufman *et al.* (1989), are tentatively correlated with the Colvillian, Bigbendian, and Fishcreekian transgressions respectively (Brigham-Grette and Kaufman, 1990; unpubl. data). However, it is important to note that the aIle/Ile ratios are much higher at Nome than on the Arctic Coastal Plain, reflecting the warmer Pleistocene thermal history of these deposits at only 64°N latitude. Also, due to the nonlinearity of isoleucine kinetics beyond a ratio of about 0.3-0.4, it is difficult to accurately model the age/temperature relationships in the deposits at Nome. However, the relative differences in the extent of isoleucine epimerization among the three transgressions at Nome are similar to the differences measured for the three transgressions of the Arctic Coastal Plain, suggesting that the transgressions in the two areas were separated by similar time intervals. The tentative correlations between the Alaskan Arctic Coastal Plain and the Seward Peninsula are shown on Figure 7 as solid lines. The slopes of these lines are parallel, suggesting that the temperature gradient between these two areas has been essentially constant for the past 2-3 m.y. Initial attempts to use strontium isotopes to date the deposits at Nome have been unsuccessful because of post-depositional replacement of strontium (Kaufman *et al.*, 1990).

Deposits that may be correlative with the two oldest transgressions on the northern Alaskan Coastal Plain crop out south of Tolstoi Point on St. George Island, one of the Pribilof Islands, southeastern Bering Sea. Here, two sequences of marine sediments are separated in places by an unconformity and in other

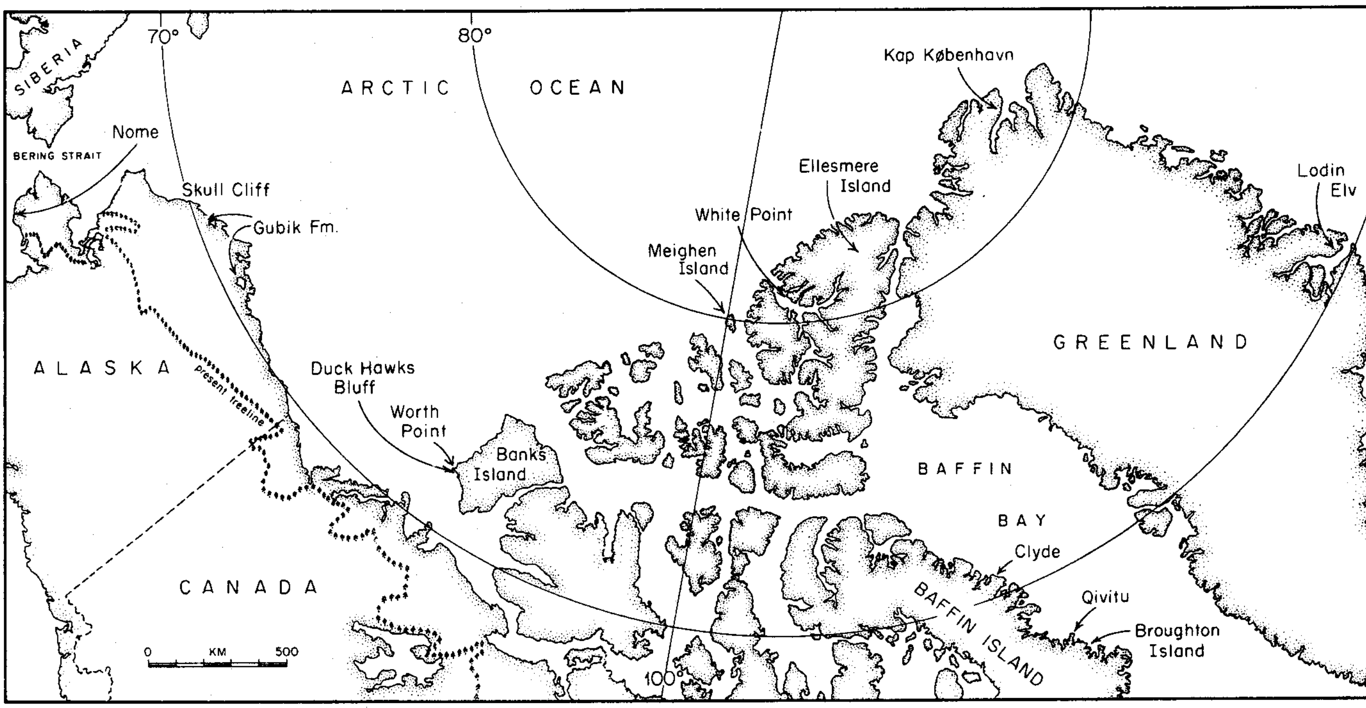

FIG. 6. Pliocene sites discussed in the text.

TABLE 4. Extent of isoleucine epimerization in mollusks of Pliocene to early Pleistocene age from the arctic region

	Estimated age	Stratigraphic unit	Species*	aIle/Ile (no. of shells) Free	Total	References
Kamchatka, U.S.S.R.	Pliocene to Pleistocene	Tusatuvauam beds	Mt	1.160 ± .03	0.577 ± .004 (2)	Brigham, 1983, 1985
	Late Pliocene	Lower Olkhovaya Suite	Ha	1.380 ± .03	1.040 ± .06 (2)	
			Mt	1.280 ± .03	1.040 ± .05 (2)	
Seward Penisula, AK	Late Pliocene	Beringian III	Ha	1.000 ± .050	0.422 ± .03 (31)	Kaufman *et al.*, 1989
	Late Pliocene	Beringian II	Ha	1.068 ± .044	0.540 ± .03 (23)	Kaufman, 1991
	Late Pliocene	Beringian I	Mt	1.138 ± .030	0.635 ± .03 (13)	
Meighen Island	Pliocene	Beaufort Fm. (75-2)	*Arctica*	0.782 ± .062	0.213 ± .028 (3)	Brigham-Grette *et al.*, 1987
		Beaufort Fm. (FG-61-144c)	Ha	0.906 ± .044	0.221 ± .014 (3)	This paper
Ellesmere Island	Pliocene	White Point (FG-61-149a)	Ha	0.530 ± .444	0.054 ± .0014(3)	This paper
N.E. Greenland	Pliocene to Pleistocene	Kap København Fm. B2	Mt	0.727 ± .032	0.140 ± .018 (3)	Funder *et al.*, 1985
		A	Ha	0.790 ± .028 (2)	0.118 (1)	
East Greenland	Pliocene to Pleistocene	Lodin Elv Fm. (GGU No. 231420)	Ha	0.950 ± .044 (2)	0.520 ± .023 (3)	Feyling-Hanssen *et al.*, 1983
Baffin Island	Late Pliocene	Clyde-Tupirtalik	Ha	0.90	0.18	Mode, 1985
	Late Pliocene	Clyde-Pinguruluk	Ha	0.89	0.24	
	Late Pliocene	Clyde-Akulaanga	Ha	1.00	0.33	
	Late Pliocene	Clyde-oldest Pre-Cape Christian	Ha	—	0.55	Miller, 1985
	Early Pleistocene	Qivitu AZ 5**	Ha	—	0.29 ± .06	A.R. Nelson, 1982
	Pliocene	Qivitu AZ 6	Ha	—	0.36 ± .04	
	Pliocene	Qivitu AZ 7	Ha	—	0.47 ± .08	
	Pliocene	Qivitu AZ 8	Ha	—	0.60 ± .06	

* Mt, *Mya truncata*; Ha, *Hiatella arctica*; Arctica, *Arctica islandica*.
** AZ = aminozone.

places by a thick sequence of basaltic lava flows. The younger marine sequence is overlain by basaltic pillow lava K/Ar dated at 2.2 Ma (Hopkins, 1967:Fig. 3; Hopkins, unpubl. data). The sparse molluscan faunas contained in these Pliocene sediments unfortunately provide no useful basis for correlation with faunas in the three Beringian marine units at Nome or with faunas from the Arctic Coastal Plain. Amino acid geochronology has not been attempted, but we anticipate that transient heat produced during eruption of the lava flows would preclude its use for correlation. However, it seems likely that the younger transgressive marine deposits near Tolstoi Point correlate with the Bigbendian deposits in northern Alaska, and the older transgressive deposits probably correlate with the Colvillian transgression.

Kamchatka Peninsula

Marine beds representing Pliocene and Pleistocene transgressive events have been described from the western shores of the Bering and Chukchi seas on Kamchatka and Chukchi peninsulas (Petrov, 1967, 1982; Khoreva, 1974). Samples of *Hiatella arctica* and *Mya truncata* collected by O.M. Petrov, Geological Institute of Moscow, from many of these marine units were analyzed to determine the progression of amino acid diagenesis in the chronostratigraphic series (Brigham, 1985). The oldest unit analyzed from Kamchatka, the Lower Olkhovaya suite, is magnetically reversed and directly underlies the oldest known glacial diamictons in the region (Arkhipov *et al.*, 1986; Fig. 1). Shells of *Hiatella arctica* from this unit are racemic (that is, at chemical equilibrium) in the Free fraction (average aIle/Ile ratio of 1.38) and nearly racemic in the Total acid hydrolysate (average 1.04) (Table 4). Nearly 70% of the amino acids in these samples are non-peptide bound, further

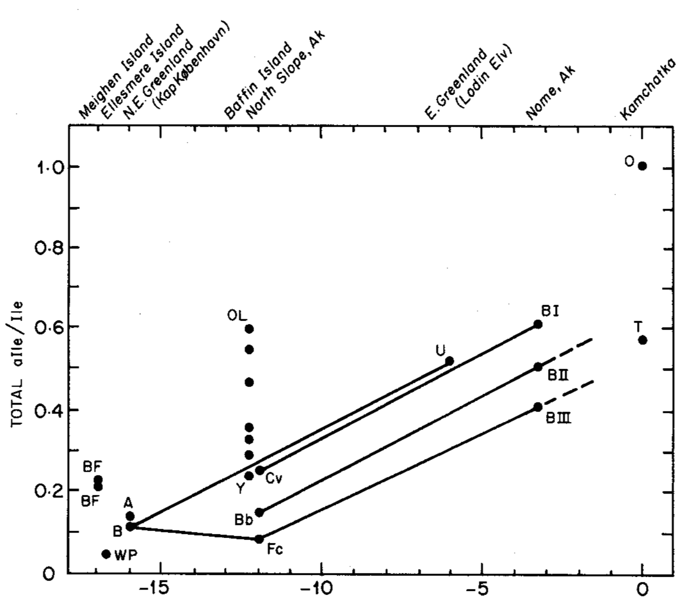

FIG. 7. Average aIle/Ile ratios in the Total fraction of Pliocene and some Pleistocene sites throughout the Arctic. Solid lines indicate provisional correlations discussed in text. Meighen Island, BF = Beaufort Formation; Ellesmere Island, WP = White Point; Northeast Greenland, Kap Kobenhavn Formation, A = Member A, B = Member B; Baffin Island, Ol = oldest, Y = youngest; Arctic Coastal Plain, Alaska, Cv = Colvillian transgression, Bb = Bigbendian transgression, Fc = Fishcreekian transgression; Eastern Greenland, Lodin Elv Formation, U = upper part; Nome, Alaska, BI = Beringian I transgression, BII = Beringian II transgression, BIII = Beringian III transgression; Kamchatka, O = Lower Olkhovaya suite, T = Tusatuvayam beds.

reflecting their antiquity. These results suggest that these sediments are much older than the Beringian I of Kaufman *et al.* (1989) in western Alaska and the Colvillian transgression of the North Slope, especially when compared with data from other Pliocene stratigraphic units (Fig. 7). Specimens of *Mya* sp. from the Coralline Crag of Pliocene age in the British Isles, an area where the current mean annual air temperature (CMAT) is about +10°C, gave aIle/Ile ratios of 1.20 ± 0.002 in the Total hydrolysate (Miller *et al.*, 1979). The higher ratios of the Lower Olkhovaya suite, where the CMAT is only about O°C, and average Pleistocene temperatures were doubtless even lower, suggest that the Lower Olkhovaya suite is considerably older than the Coralline Crag.

Our interpretation conflicts with the correlation proposed on paleontological grounds by Gladenkov (1981:20), who equated the entire Lower Olkhovaya suite with marine deposits younger than the Anvilian transgression of Hopkins (1967). The Anvilian transgression was thought to be late Pliocene in age, but recent studies (Kaufman, 1991; Huston *et al.*, 1990) have shown that Anvilian deposits at the type section in Nome are correlative with deposits hitherto assigned to the Kotzebuan transgression of the Kotzebue Sound area and that both are middle Pleistocene in age and on the order of 0.4-0.6 Ma, perhaps correlative with ^{18}O stage 11 (Kaufman *et al.*, 1991).

The Tusatuvayam beds on Karagin Island, Kamchatka, with an average Total aIle/Ile ratio of 0.58, possibly may be correlative with one of the younger Beringian transgressions (II or III) of Kaufman *et al.* (1989), based on the broad similarity of amino acid ratios. It is not possible to clearly evaluate this correlation using the amino acid ratios, however, due to non-linearity of the epimerization reaction beyond a ratio of about 0.3 and our lack of adequate age calibration in the region. Based on the mollusk faunas, Gladenkov (1981) equated this unit with the Anvilian transgression, which he believed was late Pliocene in age.

Banks Island

Vincent (1982, 1990) and Vincent *et al.* (1983, 1984) have reported on the Pliocene and Pleistocene glacial and sea level history of unconsolidated deposits on Banks Island and adjacent areas of the Canadian Arctic Archipelago (Fig. 6). The oldest preglacial unconsolidated deposits on Banks Island, the Worth Point Formation, do not include marine sediments, but we mention them here because they were recently correlated with sediments of the Fishcreekian transgression on the basis of similar floristic characteristics and the fact that both deposits are reversely magnetized (Matthews and Ovenden, 1990). Both the Worth Point beds and the Fishcreekian sediments contain *Larix* macrofossils and other floristic elements suggestive of a forest-tundra environment near the tree line (Vincent, 1990; Matthews and Ovenden, 1990). If this age estimate is correct, and if our Pliocene age assignment for the Fishcreekian sediments is correct, then the two deposits are not correlative.

Meighen Island

The Beaufort Formation on Meighen Island (Fig. 6) consists of approximately 220 m of unconsolidated sand, silt, and clay. The clay-rich sediment is marine and extends to nearly 100 m asl. Botanical and insect remains from deposits overlying the marine sediment have been extensively reported and found to be one of the earliest records of some elements of forest tundra in the Northern Hemisphere (Matthews, 1987;

Matthews and Ovenden, 1990). Age estimates based upon these fossils range from the middle Miocene to the early Pliocene (Hills, 1975; Matthews, 1987, 1989). McNeil (1990) reports the foraminifer *Cibicides grossa* in the marine unit, suggesting a Pliocene age of greater than 2.4 Ma. Moreover, strontium isotopes on shells of the boreal mollusk *Arctica islandica* from the same marine sediments indicate a calibrated-age between 2.8 and 5.1 Ma (Kaufman *et al.*, 1990; Matthews and Ovenden, 1990; Matthews *et al.*, 1990). This age of the marine beds is further constrained by the co-occurrence of *Arctica islandica*, presently an Atlantic endemic, and *Mya truncata* and *Hiatella arctica*, mollusks of Pacific origin, indicating that the deposits must closely postdate the submergence of Bering Strait about 3.2 Ma (Fyles *et al.*, 1991).

During an earlier study to document racemic conditions in arctic marine shells (Brigham-Grette *et al.*, 1987), hinge fragments of *Arctica islandica* from Meighen Island were analyzed for their amino acid content. The aIle/Ile ratio in the Total fraction of these samples averaged 0.213 ± 0.028 (Sample 75-2, Table 5). More recently, *Arctica* shells from a nearby site produced similar Total ratios averaging 0.221 ± 0.014 (FG-61-144c). Both samples (Table 5) yielded overlapping strontium isotopic ratios (Ken Miller, Lamont-Doherty, written comm. to J.G. Fyles, 1989; and *cf.* 2.8-5.1 Ma, Kaufman *et al.*, 1990), and the similarity of the amino acid ratios would suggest that they are probably the same age or do not differ in age by more than 100 000 years.

Brigham-Grette *et al.* (1987) pointed out that the low amino acid ratios from deposits older than 2.4 Ma require that ground temperatures dropped below 0°C shortly after deposition and subsequent emergence of the marine sediments and have remained as cold or colder since that time. If these *Arctica* had experienced ground temperatures of roughly 4°C for 1-2 m.y., then the measured ratios should be much higher than 0.213. In fact, *Mya*, which racemizes at about the same rate as *Arctica*, reaches equilibrium values of about 1.3 in *ca.* 2 m.y. at lower latitudes around the North Sea. This interpretation is consistent with the forest-tundra flora and fauna in deposits overlying the marine sequence and implies the presence of permafrost beginning shortly after deposition of the Beaufort marine beds on Meighen Island.

Although the amino acid ratios on *Arctica* at Meighen Island are similar to those on *Hiatella* and *Mya* from the Colvillian transgression on northern Alaska (Fig. 7), the higher latitude and inferred cooler thermal history, along with the strontium ages and paleobotanical evidence, suggest that the deposits on Meighen Island are probably older.

Ellesmere Island

At White Point, Ellesmere Island, marine shells of *Hiatella arctica* from a deposit consisting of gravel and sand yielded Total aIle/Ile ratios of only 0.054 ± 0.0014 (FG-61-149a, n = 2), significantly lower than the shells from deposits on Meighen Island. These same shells yielded a preliminary strontium ratio of 0.709101, much higher than those from Meighen Island (Table 5). We suggest that these deposits represent a marine transgression younger than the Meighen Island deposits but possibly correlative with the Colvillian or Bigbendian transgressions recorded on the Alaskan Arctic Coastal Plain. Samples from White Point contain *Cibicides grossa*, and the foraminiferal assemblage led McNeil (1990) to suggest that these deposits must be of late Pliocene age. Most important, they are not the same age as the deposits on Meighen Island.

Baffin Island

The broad forelands at Clyde, Qivitu and, to a lesser extent, Broughton Island along the northeast coast of Baffin Island (Fig. 6) contain a rich record of Pliocene and Pleistocene glacio-isostatic and eustatic events. Stratigraphic work by Miller *et al.* (1977), A.R. Nelson (1981), Brigham (1983), and Mode (1985) collectively outline evidence for at least eleven marine or glaciomarine units, three or four of which are late Pliocene or early Pleistocene in age (Table 3). Summaries of the regional stratigraphy include discussions by Miller (1985) and Andrews (1988), who suggest age ranges for these deposits based upon amino acid data and assumed thermal histories. Based upon foraminiferal biostratigraphy, Feyling-Hanssen (1976, 1980, 1985) suggests that the four oldest deposits are late Pliocene in age. Moreover, all of the marine units on Baffin Island include mollusks of Pacific origin (Mode, 1985: Table 17.3), suggesting that all of the oldest units must be younger than the opening of Bering Strait (*cf.* Durham and MacNeil, 1967).

The oldest aminozones at Clyde and Qivitu forelands are defined by amino acid ratios that average 0.55 and 0.60 respectively (Table 4). Using an effective diagenetic temperature (EDT) of −9°C (*cf.* CMAT is *ca.* −12°C), Miller (1985) suggested the beds at Clyde could be as old as 3.5 Ma; however, a higher EDT of −5°C would suggest they are as young as 1.6 Ma. Because these units underlie deposits rich in *Cibicides grossa* (*cf.* McNeil, 1990), the older age estimate may be most accurate and would concur with Feyling-Hanssen's (1980, 1985) interpretation of the biostratigraphy. Preliminary paleomagnetic data from the deposits at Clyde indicate that the

TABLE 5. Comparison of amino acid and strontium results from northern Queen Elizabeth Islands

Sample ID[1]	Genera	aIle/Ile ratio total (n)	Sr-isotope[3] mean value	Sr age range	Sr reference
Beaufort Formation, Meighen Island					
75-2	*Arctica*	0.213 ± 0.028 (3)	0.709038		Rutgers[2]
			0.709027	2.8-5.1 Ma	Kaufman *et al.*, 1990
FG-61-144c	*Arctica*?	0.221 ± 0.014 (3)	0.709065		Rutgers
White Point, Ellesmere Island					
FG-61-149c	*Hiatella*	0.054 ± 0.0014 (2)	0.709101		Rutgers

[1]Samples collected by John Fyles and John V. Matthews, Jr., Geological Survey of Canada.
[2]Unpublished results from Kenneth G. Miller and M.D. Feigenson, Rutgers University; age ranges not given.
[3]Rutgers lab isotopic standard is NBS-987 = 0.710252 (2σ standard deviation 0.000026, n=35) normalized to $^{86}Sr/^{88}Sr$ of 0.1194 (Miller *et al.*, 1991); Kaufman *et al.* (1990) lab isotopic standard is EN-1 = 0.701980. According to Hodell *et al.* (1989), EN-1 is lower than NBS-987 by 0.001060. Hence, EN-1 should be 0.701980 + 0.001060 = 0.710240 relative to NBS-987.

sediments are normally magnetized (Jacobs *et al.*, 1985), which, together with the amino acid and foraminiferal evidence, suggests that the sediments were deposited during the Gauss Normal-Polarity Chron.

Younger aminozones including the Tupirtalik, Pinguruluk, and Akulaanga aminozones are estimated to be 2.0-0.80 Ma, according to Andrews (1988:Fig. 7). However, these units fall within or below Feyling-Hanssen's (1985) *Cibicides grossa* zone, suggesting that they may be older. The paleomagnetic data through these units is too sparse to make firm interpretations, but the data include measurements indicating reversed magnetic polarity in beds below the Pinguruluk aminozone at Clyde (Jacobs *et al.*, 1985; Andrews, 1988).

Based upon a regional comparison, it appears that aminozones 6 and 7 at Qivitu (Table 4) are not present in the stratigraphy at Clyde. On the other hand, Mode (1985:Table 17.1) correlates the Tupirtalik and Pinguruluk at Clyde with aminozone 6 at Qivitu and the Akulaanga at Clyde with aminozone 7 at Qivitu, despite the distinct difference in ratios from the two areas, and does not discuss the basis for this.

Like Banks Island, the complex glacial isostatic and eustatic history confounds attempts to correlate marine and glaciomarine units of northeastern Baffin Island with deposits of the lower part of the Gubik Formation. Although the present mean annual temperature is nearly the same, longer periods of submergence in response to isostatic loading along the Baffin coast would yield higher amino acid ratios as compared to units of similar age on the North Slope of Alaska, and this accounts for the position of these units on Figure 7. However, we suggest that parts of the older record on Baffin Island are probably correlative with the lower part of the Gubik Formation. As mentioned before, future work to determine whether *Cibicides grossa* occurs in the Colvillian sediments will contribute to direct comparisons.

Greenland

The Kap Kobenhavn Formation (Funder *et al.*, 1985; Bennike and Bocher, 1990) and the Lodin Elv Formation (Feyling-Hanssen *et al.*, 1983) contain some of the oldest late Cenozoic paleoclimatic information on Greenland. Both are thought to be late Pliocene to early Pleistocene in age.

The Kap Kobenhavn Formation, in Independence Fjord along northeast Greenland (Fig. 6), consists of two marine units: 1) a lower glaciomarine facies, Member A, of laminated silt and clay with isolated pockets of ice-rafted debris, and 2) an upper nearshore high-energy facies, Member B, of sand and silt. In contrast to the permanently ice-bound coast that exists today in the area, sedimentary structures and *in situ* mollusks in Member B indicate that these sediments were deposited when the coastal waters were at least seasonally ice free and wave-generated sedimentary structures could form. Paleobotanical evidence suggests that the climate in the region was probably similar to present-day Labrador (Funder *et al.*, 1985). This reconstruction indirectly implies that the Greenland Ice Sheet was probably not present or was much reduced in size.

Present age estimates for the Kap Kobenhavn Formation are based on faunal, paleobotanical, and paleomagnetic evidence. The presence of mollusks of Pacific origin in both members requires that the entire sequence is younger than the opening of Bering Strait at about 3.2 Ma. Member A, which contains the foraminifer *Cibicides grossa*, can be no younger than 2.4 Ma (McNeil, 1990). The co-occurrence of *Lepus* (modern hare) and a large species of *Hypolagus* (an extinct rabbit) in Member B suggests an age of about 2.0 Ma for this unit (Repenning *et al.*, 1987; Bennike and Bocher, 1990). Paleobotanical evidence also suggests an age of about 2.0 Ma (Matthews and Ovenden, 1990), and paleomagnetic measurements suggest deposition during a period of reversed-magnetic polarity, which supports the paleontological age estimates and indicates deposition during the early part of the Matuyama Reversed-Polarity Chron.

These age estimates suggest a correlation between Member B and sediments of the Fishcreekian transgression, which we estimate occurred sometime between 2.14 and 2.48 Ma. The paleobotanical assemblages from Member B suggest warmer climatic conditions than do the palynological analyses of Fishcreekian sediment, but this sediment formed during the regressive phase of the Fishcreekian transgression, which was almost certainly accompanied by climatic cooling. Member A may correlate with either the Bigbendian or Colvillian transgressions. Amino acid data are available only for Member B, where Total amino acid ratios average 0.13 ± 0.025 (Funder *et al.*, 1985; 0.156 ± 0.033, Funder, 1987), similar to ratios from the Bigbendian transgression. Considered alone, this implies that Member B is older than the Bigbendian transgression, because CMAT at Kap Kobenhavn is –16°C, 7°C colder than CMAT at Barrow. However, the Kap Kobenhavn Formation was covered by glacier ice for an unknown period, and epimerization rates were certainly accelerated when the ice cover was present. Although the amino acid data are not directly comparable because of uncertain thermal histories (Fig. 7), they do not preclude our suggested correlation.

The Lodin Elv Formation, Jameson Land, east Greenland (Fig. 6; Feyling-Hanssen *et al.*, 1983) consists of prodeltaic deposits overlain by glacial diamicton. The lower unit includes *Cibicides grossa* and the upper unit commonly contains *Cassidulina cf. teretis*, the later of which Feyling-Hanssen (1985) believes is common in deposits that straddle the Pliocene-Pleistocene boundary. By biostratigraphic comparison, he suggested that the Lodin Elv and the Kap Kobenhaven are similar in age (Feyling-Hanssen, 1987, 1990). Although amino acid ratios on *Hiatella arctica* from the Lodin Elv diamicton average 0.520 ± 0.023 ($n = 3$) (Feyling-Hanssen *et al.*, 1983), nearly four times higher than those on shells from the Kap Kobenhaven Formation, Funder (1987) also suggested that the two deposits are of similar age, the higher amino acid ratios in the more southern deposit resulting from higher mean ground temperatures throughout the Pleistocene (CMAT at Scoresbysund is –6.0°C).

Arctic Basin

Recent analyses of Arctic Ocean deep sea cores suggest that although perennial sea ice was initiated between 2.1 and 2.48 Ma (Scott *et al.*, 1989) it was not permanent until just before 1.65 Ma. This contrasts with statements by Herman *et al.* (1989), who maintain that permanent sea ice did not develop until *ca.* 0.9 Ma. Although "dropstones" occur in sediments as old as 4.26 Ma in one of the CESAR cores from the Alpha Ridge (Scott *et al.*, 1989), the significant occurrence of glacially ice-rafted debris does not begin until *ca.* 1.5 Ma (Darby *et al.*, 1989; Glen Jones, Woods Hole Oceanographic Institute, pers. comm. March 1990).

None of these studies has yet clearly identified the warm interglacial intervals documented on the Alaskan Arctic Coastal

Plain or elsewhere in the arctic region (*cf.* Funder *et al.*, 1985; Feyling-Hanssen *et al.*, 1983). Perhaps this is because of the slow sedimentation rates in the central arctic basin. Ice-rafted erratics seen at only a few localities in Colvillian and Bigbendian deposits along Skull Cliff may corroborate the occurrence of a few dropstones in late Pliocene sections of the deep basin cores (Scott *et al.*, 1989; Darby *et al.*, 1989).

DISCUSSION

Data from circumarctic sites provide compelling evidence for climatic conditions much warmer than at present at several times during the late Pliocene. Although we generally lack accurate knowledge of the chronology of these warm events, which commonly are bracketed by broad numerical age ranges, it is now possible to attempt circumarctic correlations based on paleontological criteria (see papers in *Arctic* 43 [4]). Amino acid geochronology also provides an important means of comparison for marine deposits, but differences in thermal history hinder confident correlations. Our correlations based on amino acid geochronology are shown in Figure 7, and our chronology and correlations based on a variety of evidence are shown in Figure 8.

These warm events have a direct bearing on whether the Arctic Ocean has had a continuous cover of perennial sea ice for the past 3.5 m.y. (Clark, 1982), and the evidence indicates that there were at least several periods when sea ice did not exist or was at most seasonal and in the central basin. Sea otter fossils from sediments of the Bigbendian and Fishcreekian transgressions, for example, suggest by analogy with modern ecological tolerances of sea otters that even seasonal sea ice did not form along the coast of northern Alaska during these high sea level events. Palynological and micropaleontological evidence indicating that terrestrial and marine conditions during the Colvillian transgression were at least as warm as those of the Bigbendian transgression suggest that sea ice was absent during the Colvillian as well. Evidence at Kap Kobenhaven suggests further that at least once during the late Pliocene, probably coincident with the Fishcreekian transgression, the north shore of Greenland was ice free, larch and

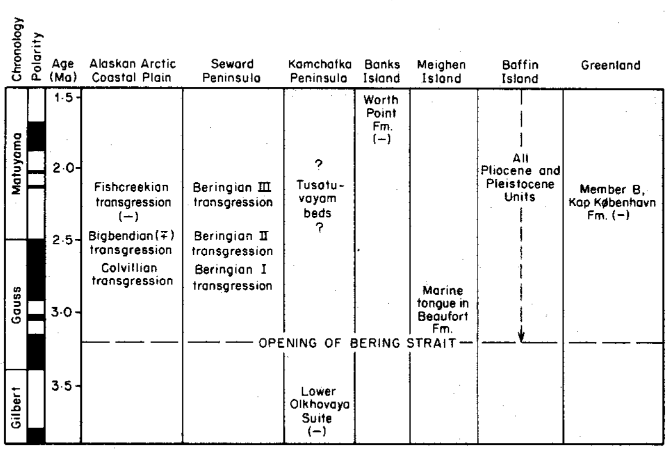

FIG. 8. Tentative relative age and correlation of some Pliocene and Pleistocene arctic and subarctic marine units. Paleomagnetic time scale (Mankinen and Dalrymple, 1979) along the left axis. (–) indicates reversed-magnetic polarity, (+) indicates normal-magnetic polarity. "?" indicates queried age ranges. Gilbert, Gilbert Reversed-Polarity Chron; Gauss, Gauss Normal-Polarity Chron; Matuyama, Matuyama Reversed-Polarity Chron. Shaded = normal-magnetic polarity; unshaded = reversed magnetic polarity.

cedar grew at the coast, and the Greenland Ice Sheet did not exist in its present form. These conclusions may support the suggestion of Raymo *et al.* (1989) that the climatic deterioration of the late Pliocene was complex rather than uni-directional.

If the coastal regions of Alaska and Greenland are ice free even in winter, can a seasonal or perennial ice pack be maintained over the central Arctic Ocean? Attempts to model the stability of the present arctic sea ice cover and to model the transitional states between an ice-free and perennially ice-covered Arctic Ocean have had ambiguous results. However, most modelers agree that an open Arctic Ocean would in itself produce considerable warming north of 65°N latitude (see Barry, 1989, for review). We assume that the formation of even seasonal winter ice over the central Arctic Ocean during a warm interglacial would require that the ocean be stratified with a freshwater lid, probably from the drainage of Russian rivers. Moreover, if the ocean was ice-covered only in winter, as in Baffin Bay and the Sea of Okhotsk today, seasonal sea ice probably would form initially over the shallow shelf areas before extending over deeper water (Barry, 1989). Therefore, we suggest that the lack of coastal sea ice during the Bigbendian and Fishcreekian transgressions makes it unlikely that seasonal sea ice formed over the central Arctic Ocean at these times.

CONCLUSIONS

The lower part of the Gubik Formation, consisting of sediments deposited by the Colvillian, Bigbendian, and Fishcreekian transgressions, represents the only depositional sequence in the circumarctic region where at least three Pliocene high sea level events can be found in superposition. The oldest transgression, the Colvillian, which occurred sometime between 2.48 and 2.7 Ma, is slightly younger than the initial late Cenozoic submergence of the Bering Strait. Limited biostratigraphic data suggest that marine temperatures during this transgression were warmer than at present. In fact, the rapid migration of Pacific mollusks of boreal affinity to the shores of the North Atlantic suggests that little or no sea ice may have been present.

During the Bigbendian transgression, which occurred about 2.48 Ma, marine temperatures were warm enough to preclude even seasonal ice along northern Alaska and coniferous forests extended to the arctic coast.

The Fishcreekian transgression occurred at sometime between 2.14 and 2.48 Ma and was also accompanied by warm marine conditions lacking seasonal sea ice; however, conifer forests were some distance south of the Brooks Range foothills and a mosaic of larch forest tundra and wetland herbaceous tundra occupied emergent parts of the coastal plain.

Based upon biostratigraphic evidence and amino acid geochronology, we believe the marine tongue of the Beaufort Formation on Meighen Island was deposited during or just after the initial submergence of Bering Strait *ca.* 3.2 Ma (Fyles *et al.*, 1991). Member B of the Kap Kobenhaven Formation on Greenland is likely contemporaneous with the Fishcreekian transgression on the Alaskan Arctic Coastal Plain. Correlations between arctic Alaska and Baffin Island remain unclear; however, the presence of mollusks of Pacific origin suggests that the oldest recognized marine units at Clyde and Qivitu forelands were deposited after the opening of Bering Strait.

Floral remains recovered from terrestrial deposits of the Worth Point Formation on Banks Island are most similar to the plant cover interpreted for the Arctic Coastal Plain during the Fishcreekian transgression (Matthews and Ovenden, 1990), but the Fishcreekian transgression is evidently older. Tentatively, we believe that the three Beringian marine units now recognized at Nome are correlative with the three Pliocene transgressions in arctic Alaska.

Ongoing work by the U.S. Geological Survey and the Geological Survey of Canada will eventually clarify the micropaleontology of these units. Moreover, future work planned by U.S. and Russian paleontologists and stratigraphers on either side of the Bering Strait will provide needed insights for correlating the North American Arctic with the rich record in northeast Siberia.

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