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Marine stratigraphy and amino-acid geochronology of the Gubik Formation, western Arctic Coastal Plain, Alaska

by

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Dedicated in Memory of

Terry L. Brown (1954 - 1982) who was my assistant during my first field season on the North Slope in 1980

and

Fred Hemphill (1951 - 1983) a friend who influenced my life and my perspective on scientific endeavors during the course of this study.

Both of these individuals were employees of the U.S. Department of Interior.

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ABSTRACT

The Pliocene/Pleistocene Gubik Formation that mantles the Arctic Coastal Plain includes wedges of marine sediment that represent one of the most complete records of high sea level stands in Arctic North America. These marine sequences have been subdivided into six members, each of which corresponds to a specific aminozone representing a depositional time period when sea level was above present. The oldest marine units, the Nulavik and Killi Creek Members, are the most poorly preserved along Skull Cliff and represent high sea level stands that took place > 2.2 and between 1.7 and 2.2 m.y. ago, respectively. An early Pleistocene transgression represented by deposits up to 33-36 m a.s.l. of the Tuapaktushak Member occurred sometime between 0.7 and 1.7 m.y. B.P. Extralimital mollusk faunas in these latter deposits suggest that coastal conditions along northern Alaska were warmer than they are today and that the Arctic Ocean may have lacked perennial sea ice during a portion of the early Pleistocene. The Karmuk Member consists of marine sediments deposited when sea level reached about 23 m a.s.l. about 500 ka B.P. and is characterized by marine conditions similar to today. About 125 ka B.P. sea level reached 8 to 10 m a.s.l. and beach and lagoonal sediments of the Walakpa Member were deposited. Marine conditions in the Chukchi Sea at that time were similar to today: however, several mollusks that today reach their northern limits in the Chukchi Sea expanded their ranges into the Beaufort Sea at this time.

The youngest transgression onto the coastal plain took place 80 to 100 ka B.P. and deposited the Flaxman Member. These sediments are preserved only along the Beaufort Sea coast. At Barrow, Alaska, this transgression resulted in the formation of a curved spit about 7 m a.s.l. that is similar in form to the modern spit at Point Barrow. Eolian sand was periodically deposited and redistributed as a discontinuous sand sheet across portions of Coastal Plain throughout the Wisconsin. Mean annual air temperatures during the Happy and Duvanny Yar glacial intervals of the Wisconsinan averaged between -17°C and -23°C, or more than about 4.5°C lower than at present. Arctic summers throughout these periods were probably cooler. Finally, a dramatic warm interval, known as the birch period, initiated the active growth of thaw lakes and ice wedges and by 8.5 ka B.P. <u>in situ</u> peat began to accumulate.

CHAPTER I

INTRODUCTION

Throughout late Pliocene and Pleistocene time, world sea level oscillated repeatedly in response to the intermittent growth and decay of large ice masses on land. These changes in sea level are recorded by the often patchy occurrence of coastal sediments lying above present sea level and transgressive/regressive marine sequences laced by unconformities blanketing the continental shelves. Such deposits provide a means of subdividing the last 3 million years into climatically significant periods of regional and global extent. Moreover, they contain a faunal record of changes in the distribution and environmental conditions of contiguous water masses and the migration and evolution of vertebrate and invertebrate forms. This biostratigraphic history remains as a key to reconstructing late Cenozoic terrestrial climates and understanding past fluctuations in the ocean/atmosphere system. This study is concerned with the stratigraphy and Late Cenozoic history of the Gubik Formation in northwestern Alaska and represents a contribution toward understanding the paleogeography of the arctic regions during the last 3 million years.

Description of the Problem

The Arctic Coastal Plain and adjacent continental shelves of northern Alaska are mantled with unconsolidated deposits that record multiple sea-level events during Pliocene and Pleistocene time. This sedimentary complex, collectively recognized as the Gubik Formation (Payne and others, 1951; Gryc and others, 1951; Black, 1964) includes beach, nearshore, and shallow-marine shelf sediments, as well as fluvial, eolian and thaw lake facies.

Because this monotonously flat landscape was never glaciated throughout Pliocene and Pleistocene time, the transgressive/regressive marine sequences of the Gubik Formation probably comprise one of the most complete records of high sea level stands in Arctic North America. In addition, these deposits contain an invaluable almanac of paleoclimate in the high arctic, recording not only the migration and exchange of marine vertebrate and invertebrate faunal elements between the Pacific and Atlantic basins via the Arctic Ocean, but also the migration during low sea level intervals of terrestrial faunal and floral elements across the Bering Land Bridge between Asia and North America. Hopkins (1959, 1967a, 1967b, 1972, 1973, 1982a) has repeatedly emphasized the biogeographic significance of the entire region of Beringia, pointing out that changes in world sea level throughout the last 3 or 4 my have directly controlled both intercontinental and interoceanic dispersal routes (Hopkins, 1967a).

The purpose of this study is to describe the physical stratigraphy, determine the geochronology of the units, elucidate the sea level history, and synthesize the evidence concerning the paleoclimatic history of the Gubik Formation exposed across the western Arctic Coastal Plain, between Point Barrow and Cape Beaufort (Fig. 1.1). This region includes an exceptionally complete and well-exposed sequence of late Cenozoic marine sediments in the long bluff known as Skull Cliff. This bluff is now recognized as being the only area in Alaska, and probably in northwestern North America, where all of the known Pliocene and early and middle Pleistocene marine transgressive units are present in superposition. Thus, this stretch of coast represents a key section for understanding the marine fluctuations of the late Cenozoic in Beringia.

Organization

Chapter I provides a broad introduction to the physical characteristics of the northwestern Arctic Coastal Plain, the purpose and objectives of the dissertation, methods, and background information concerning previous Quaternary investigations in the area. Chapter II describes the lithostratigraphy of the Gubik Formation as it is exposed in the bluffs southwest of Barrow, followed by a discussion of the aminostratigraphy of these units in Chapter III. Chapter IV integrates the morphology of the landscape with the stratigraphy of marine sediments, particularly those exposed in riverbanks and coastal bluffs south and west of Peard Bay. Chapter V introduces the vertebrate and invertebrate paleontology of the aminostratigraphic units and extracts the paleoclimatological evidence associated with each superposed marine deposit. Chapter VI presents paleotemperature reconstructions for the Coastal Plain during the last 125,000 years based upon the thermodynamic constraints of the amino acid data. Chapter VII draws upon the information in previous chapters and presents a composite stratigraphic and geochronologic subdivision of the Gubik Formation and a synthesis of the sea level history it records. Finally, Chapter VIII introduces the correlation of the Gubik Formation to similar chronologies in adjacent areas of Alaska and Canada as well as the northeastern coasts of the Kamchatka and Chukotka peninsulas, USSR.

Background

The unconsolidated deposits of the North Slope of Alaska were first described by explorers who ventured into the region during the late 1800's (outlined by Leffingwell, 1919, p. 69-92; Black 1964, p. 62). The first systematic geologic descriptions were made by Dall and Harris (1892, p. 260-268) and Schrader (1904, p. 91-93) who reported the presence of massive ground ice and widely distributed clay, sand, and gravel containing the remains of Pleistocene mammals, notably along Skull Cliff and Colville River (reviewed by Black 1964, p. 62). Dall and Harris (1892) proposed the name "Kowak clays" for deposits of clay overlying the ground ice formations. Schrader (1904), on the other hand, proposed the name "Gubik sand" for surficial deposits of Pleistocene age along the Colville River; the name was evidently derived from "Kupik", the Eskimo name for the Colville River (Leffingwell 1919, p. 150). Schrader (1904, p. 93) spoke of the Gubik sand as

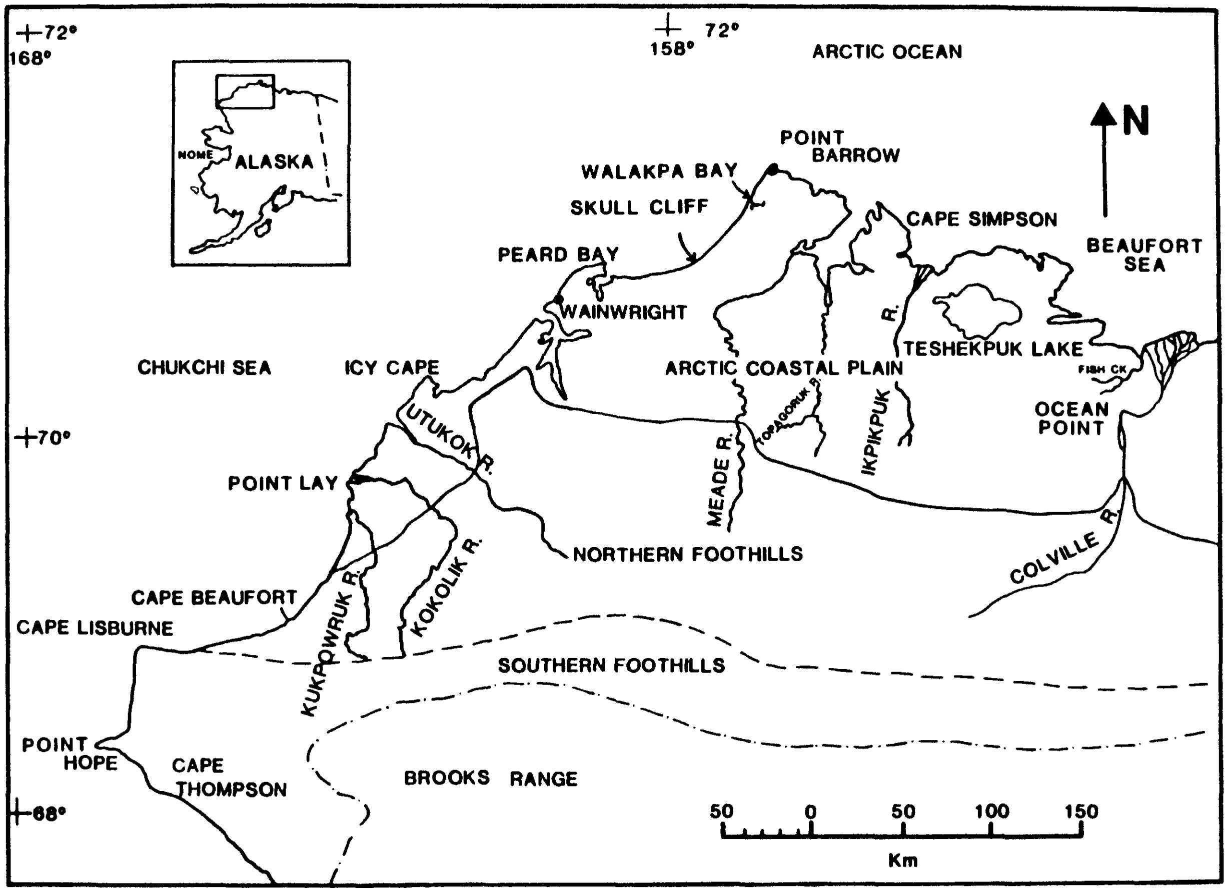


Figure 1.1. Location of the field area in northwestern Alaska, between Point Barrow and Cape Beaufort. Solid, dashed and dot-dashed lines indicate the boundaries of physiographic provinces (Wahrhaftig, 1965).

> . . .[a] deposit [that] seems to be distinct from the Colville series and to extend over a large area of country. It not only forms the surficial terrane of the coastal plain along the Colville, but seems to occur at some localities along the coast from the mouth of the Colville westward, in some instances apparently overlying the ground ice and probably the Kowak clay formation, while its inland margin probably overlaps the coastal edge of the Upper Cretaceous of the Nanushuk series along the Anaktuvuk nearly to the mouth of the Tuluga, where, in certain localities, judging from topographic criteria, it also appears to overlie gravels which are very tentatively referred to as glacial, but to which its relation is not definitely known.
>
> The deposit consists of fine sand, with apparently an admixture of considerable silt. In some localities it seems to be more sandy toward the base, and more earthy toward the top, where it terminates in from one to several feet of dark brown or black humus, clothed at the surface with moss and a little grass. It is ordinarily free from gravel, but in several instances subangular cherty pebbles ranging from mere sand grains to fragments as large as one-fourth inch in diameter were found. These occur very scatteringly and are sometimes roughened, as if wind worn. In some localities a fine gravel seems to intervene between the base of the deposit and the underlying Tertiary beds, as if representing the basal part of the deposit.

Leffingwell (1919) later spent several years between 1906 and 1914 mapping and describing all aspects of the bedrock, surficial materials and ground ice in the valley of the Canning River and westward along the coast to Point Barrow. Notably, he was the first to recognize and describe ice wedges and to explain their origin (Leffingwell, 1919, p. 206-214). Moreover, he proposed the name "Flaxman Formation" for what he described as "a deposit of foreign glacial till, possibly containing glacial ice, scattered along the Arctic coastline of America" (Leffingwell, 1919, p. 142). He later added (Ibid., p. 147) that

> the only workable hypothesis [for the origin of this deposit] is that the foreign material was brought into its present location either by an ice sheet that extended over the Arctic Ocean or by floating icebergs. It seems scarcely possible that an ice sheet could have extended along the coast of Alaska without leaving conspicuous marks of its presence. The material was probably transported by icebergs.

It is interesting to note that the arguments concerning the hypothesis of an Arctic Ocean ice sheet and the origin of the Flaxman deposits still continue in the literature (Hughes and others, 1977; Williams and others 1981; Hopkins, 1982; Clark, 1982).

Leffingwell's (1919) reconnaissance resulted in some of the first fossil mollusk collections from the Skull Cliff area (Dall, 1920) and expanded the known distribution of the Gubik Formation southwestward from Barrow along the Chukchi Sea. Meek (1923) examined the stratigraphy and fossil mollusks of the Gubik at the southwestern end of Skull Cliff near Peard Bay, and concluded that all of the unconsolidated sediments in that area were of Pleistocene age and represented a single period of deposition. During reconnaissance work in northwestern Alaska from 1924 to 1926, Smith and Mertie (1930, p. 238) examined unconsolidated marine sediments between Cape Beaufort and Peard Bay and noted

> evidence of former lower stands [of the land] which have since been uplifted is now recognizable through the marine deposits then formed, which are seen at many places and clearly indicate relatively recent uplift. The whole coastal plain province thus appears to have been at one time under the sea. . .in no place were very thick deposits of marine origin laid down, as is shown by the fact that the underlying hard bedrock is exposed at many places where these deposits are trenched by the present streams.

They also noted the presence of erratic boulders on the coastal plain.

> In close relation to some of the marine deposits. . .are numerous fragments of rocks of apparently foreign origin. At many places along the beach from Point Lay northward to Barrow and up some of the inlets pieces of red granite, gray granite, red quartzite and some gneiss or schist were found. Fragments of this sort were particularly abundant in Peard Bay. . . . Foran notes the presence of large angular blocks of red granite and pink quartzite as much as 5 feet in each dimension and many smaller ones as much as 8 inches in diameter at many points along Wainright Inlet along the way to the head of the Kuk River(Smith and Mertie, 1930, p. 242).

Payne and others (1951) and Gryc and others (1951) renamed the "Gubik sand" the Gubik Formation and defined it to include all of the unconsolidated deposits on the Arctic Coastal Plain of Pleistocene between Demarcation Point and Cape Beaufort. MacNeil (1957, p. 100) expanded the definition of Gubik Formation to include fossiliferous marine sediments at Ocean Point along the lower Colville River which Schrader (1904) had assigned to the Upper Colville Series of Pliocene age. MacNeil (1957) reassigned these beds to the Pleistocene but more recent studies indicate that the marine silt yielding the faunas examined by Dall (in Schrader, 1904) and MacNeil (1957) overlies a fossiliferous beach gravel that enclosed a fossil sea otter estimated to date between 1.7 m.y. and 2.2 m.y. BP (Repenning, 1983; Nelson, 1979, p. 39; Carter and Galloway, 1981; Carter, unpublished). Thus, the base of the Gubik Formation as redefined by MacNeil (1957) includes sediments of late Pliocene age.

Both O'Sullivan (1961) and Black (1964) attempted to subdivide the Gubik Formation on the basis of stratigraphic sections in scattered areas of the coastal plain and by mapping major topographic breaks which they assumed to be paleoshorelines. These studies included the first published grain-size analyses of different components of the Gubik Formation. Black (1964) subdivided the Gubik Formation into three informally-named lithologic units, the Skull Cliff, the Meade River, and the Barrow units. Black (1964, pg. 88-89) was unsure of the age of these individual units, but he hypothesized that

> the Barrow unit must represent much, if not all, of the Wisconsin Glaciation. One might then correlate the Meade River unit with the Sangamon Interglaciation and the Skull Cliff unit with the Illinoian Glaciation.

Recent (1976-1981) fieldwork by U.S. Geological Survey field parties led by L.D. Carter, D.M. Hopkins, J.R. Williams, and J.K. Brigham, has shown that Black's lithologic units are, in fact, sedimentary facies that are of different ages in different areas spatially across the coastal plain (Hopkins and others, unpublished manuscript).

J.R. Williams studied the stratigraphy and surface sediments of the Gubik Formation along the Chukchi sea coast and inland to the Meade River during the summers of 1975 to 1979 (in Williams and others, 1977; Williams, 1979; 1983a, 1983b). Hopkins and assistants worked along portions of Skull Cliff, Kuk River, the coast near Icy Cape and the Alaskan coast of the Beaufort Sea in 1976 through 1983. Fieldwork by Carter has focused largely on the stratigraphy of the Gubik Formation east of the Meade River (Carter and others, 1979; Carter and Robinson, 1980; Carter and Galloway, 1981; Carter, 1981).

My field work was conducted on the Arctic Coastal Plain west of the Meade River during the summers of 1980 and 1981. Major emphasis of the study was placed upon bluff exposures along Skull Cliff between Barrow and Peard Bay. Reconnaissance studies of bluff exposures in Wainwright Inlet and along the lower reaches of the Kuk River were conducted using an inflatable boat and studies inland along the Utukok, Kokolik, Epizetka, and Kukpowruk Rivers were conducted by U.S. Geological Survey helicopter.

Hopkins' (1967a) classic work, describing a sequence of late Cenozoic marine transgressions in western Alaska, was the first systematic documentation of sea level changes across Beringia. It was in this paper that he proposed names for these repetitive marine transgressions (Table 1.1) and attempted to correlate these events with the geographically scattered data known at that time concerning the stratigraphy of marine transgressions across the Arctic Coastal Plain. He tentatively correlated the older units at Ocean Point and the marine section at Skull Cliff with the Beringian transgression as defined at Nome, Alaska

Table 1.1. Quaternary Marine Transgressions in Alaska (Hopkins, 1967, 1973).

Transgression	Type Locality	Altitude of Shoreline	Climate as Compared with the Present	Archaeological or Radiometric Dating
Krusensternian	Recent beach ridges at Cape Krusenstern	Within 2 meters of present sea level for deposits <4,000 yrs. old	Same	<5,000 yrs. at Cape Krusenstern; up to 10,000 yrs. for terraces along Gulf of Alaska coast
Woronzofian[1]	Bootlegger Cove Clay near Point Woronzof, Anchorage area (Miller and Dobrovolny, 1959)	Probably a few meters below present sea level	Water colder Air colder	<48,000 yrs.: >25,000 yrs.
Pelukian	Second Beach at Nome (Hopkins et al., 1960)	Two distinct high-sea-level stands at +7-10 meters	Water warmer Air slightly warmer	Ca. 100,000 yrs.
Kotzebuan	Marine beds below Illinoian drift along eastern shore of Kotzebue Sound (McCulloch et al., 1965)	Probably ca. +20 meters	Water same Air unknown	170,000 yrs.; 175,000 yrs.
Einahnuhtan[2]	Eianhnuhto Bluffs, St. Paul Is. (Cox et al., 1966)	Probably ca. +20 meters	Water same Air unknown	<300,000 yrs.; >100,000 yrs.
Anvilian	Third Beach-Intermediate Beach at Nome (Hopkins et al., 1960)	Probably much higher than Kotzebuan and Einahnuhtan: <+100 meters; >+20 meters	Water warmer Air warmer	Probably <1,900,00 >700,000
Beringian	Submarine Beach at Nome (Hopkins, et al., 1960)	Two distinct episodes during which sea level was higher than at present but probably lower than Anvilian sea level	Water much warmer Air much warmer	Last episode ca. 2,200,000 yrs. on St. George Is.

[1]Because the Bootlegger Cove Clay has been shown by Schmoll and others (1972) to be only 12,000 to 14,000 years old, Hopkins (1973, p. 521-522) tacitly abandoned the name Woronzofian Transgression, but he continued to believe that uplifted beach deposits at an altitude of 7 m near Point Barrow (Sellmann and Brown, 1973) record a mid-Wisconsonian transgression. More recently, these uplifted beach deposits have been thought to represent the shoreline facies of the Flaxman Formation which Hopkins believes to be between 80,000 and 105,000 years old (D.M. Hopkins, written communication, May 1983).

[2]The stratigraphy of the Einahnuhan Bluffs on St. Paul Island has been reevaluated on the basis of field work and potassium-argon dating carried out during 1975 and 1976 by M.L. Silberman and D.M. Hopkins. The older marine beds which comprised the type locality for the Einahnuhtan Transgression are evidently of Kotzebuan age, and higher beds previously thought to be Kotzebuan are evidently Pelukian (see Fig. 8, Hopkins, 1973). Oune sand of late Wisconsinan or Holocene age to the north of the illustrated section had previously been mistaken for Pelukian beach sand. This reevaluation indicates that the term "Einahnuhtan transgression is synonymous with the Kotzebuan transgression and that the term "Einahnuhtan transgression" should be abandoned (D.M. Hopkins, written communication, May, 1983).

(Hopkins, 1967a, p. 58). It has recently become clear, however, that the Ocean Point sediments probably represent the Anvilian transgression (Carter and Galloway, 1981) and that the Skull Cliff section includes evidence of at least 4 marine transgressions, the oldest of which may be correlative with the Beringian transgression (this study). Despite these minor modifications, Hopkins' (1967, 1972, 1973) studies remain the basic framework for evaluating coastal marine sequences all around Alaska.

In a companion paper, McCulloch (1967) described evidence for a series of paleoshorelines inland from Point Lay and traceable northward into the Wainwright area at the western edge of the coastal plain. Based upon their elevation and some paleontological evidence, he hypothesized that beach deposits reaching 33 m asl were older than Sangamon deposits and were probably correlative with "pre?-Illinoian" (Kotzebuan) sediments of the Kotzebue sound area. Older marine sediments were found inland from these deposits.

It was, in part, McCulloch's renconnaisance (1967) that drew my interests to the North Slope. From 1976 to 1979, D.M. Hopkins attempted to refine the transgressive history along the shores of Alaska using amino acid geochronology to determine the relative ages of superposed units. The analytical work was carried out under a contract between the U.S. Geological Survey and G.H. Miller at the then newly established Amino Acid Geochronology Laboratory at the Institute of Arctic and Alpine Research, University of Colorado. As a student research assistant, I became involved in handling the Alaska contracts, and thus my association with Hopkins began. It was Hopkins who actually proposed that I enter a dissertation project concerning the aminostratigraphy of the Gubik Formation along Skull Cliff. Later after reading McCulloch's (1967) paper, I proposed to expand the project to a description of the Gubik Formation between Barrow and Cape Beaufort, thus realizing a broader, more three-dimensional view of the marine transgressive history.

Objectives

The objectives of this study are:

1) to describe the physical stratigraphy and develop a geochronology for the Gubik Formation along the western border of the Arctic Coastal Plain between Cape Beaufort and Barrow and to establish a stratigraphic framework for future work in the area;

2) to elucidate the history of Pliocene and Pleistocene high sea level events recorded by the Gubik Formation;

3) to assemble and interpret paleoclimatic evidence based on micro- and macrofossil collections and pollen analysis. This study realistically represents only a systematic start on the biostratigraphy of the Gubik Formation;

4) to provide a more detailed assessment of the distribution of gravel and sand resources that may be needed for future development and construction in the area of the coastal plain and Chukchi Sea;

5) to increase the available knowledge concerning the use and application of amino acid geochronology at high latitudes;

6) to correlate the marine events documented on the western Arctic Coastal Plain with similar stratigraphic sequences in adjacent areas of Alaska and along the Bering Sea coast of northeastern Siberia.

7) to outline constraints concerning the paleotemperature history of the coastal plain since the last interglacial period.

The results of this study will be of interest to researchers across a number of Quaternary disciplines, including stratigraphers, biogeographers, and paleoclimatologists, as well as those interested in the evolution of both vertebrates and macro- and micro- invertebrates. It is hoped that this research will be of use as a stratigraphic and geochronologic reference for years to come.

Geography

> ...In the lower parts of the coastal plain where the relief is so slight that even an owl looms up with startling prominence, almost the only topographic relief is afforded by aeolian deposits . . . .
>
> Smith and Mertie (1930 p. 249)

Location

The North Slope of Alaska includes three broad physiographic provinces, the Arctic Coastal Plain, the Arctic Foothills, and the Arctic Mountains (Fig. 1.1; Wahrhaftig, 1965). These areas, which drain into the Arctic Ocean, are recognized largely on the basis of relief, elevation, geologic structure and drainage characteristics. My study was undertaken along the northwestern edge of the Arctic Coastal Plain province and includes an area of approximately 28,700 km^2 and a coastline approximately 400 km long,between Point Barrow (Lat. 71° 22' N; Long. 156° 30'W) and Cape Beaufort (Lat. 69° 4'N; Long. 163° 49'W) (Fig. 1.1).

The topography of the Arctic Coastal Plain is impressively flat and featureless, reaching a height of only 65 m asl at its widest point approximately 160 km south of Barrow. The area largely represents a marine-abrasion bedrock platform that is essentially an emergent extension of the adjacent Chukchi and Beaufort sea continental shelves. Actively eroding coastal bluffs along the Chukchi Sea average 12 m asl with bluffs as high as 23 m occurring locally. This vast tundra landscape is peppered with literally thousands of oriented and unoriented thaw lakes, low and high centered ice wedge polygons, solifluction lobes, and pingos--all related to the presence of continuous permafrost across Northern Alaska. Numerous meandering streams with broad flood plains drain the coastal plain and have eroded and reworked a great deal of the morphologic and stratigraphic record once deposited by repeated transgressions of the sea. Syntheses by Williams (1983a, 1983b) contain a thorough outline of past studies of periglacial geomorphic processes on the North Slope.

Eolian processes play an active role in shaping the local landscape (Black, 1951). Very-fine to fine-grained sand is actively excavated by wind from beaches, river bars, and eroding bluffs and redeposited locally as small dunes. A great deal of this material is redeposited once again in thaw lakes where it becomes interbedded with thin layers of organic material and clumps of peat which topple in from the thermally-eroding lake margins. Blowouts are common in well-drained areas where the wind has removed vegetation and older eolian deposits.

During the height of the late Wisconsin glacial stage, vegetation was sparse in contrast to today and the continental shelves were subaerially exposed (Carter, 1982). This allowed the intensive development of aeolian deposits, especially east of the Meade River where fields of longitudinal dunes are now stabilized by vegetation (Carter, 1981). In contrast, much of the western coastal plain is blanketed by only a thin cover of eolian sands. Nevertheless, small local stabilized dune fields are present, e.g. SW of the Kuk River estuary.

Bedrock

Across the Arctic Coastal Plain, the Gubik Formation rests on a wave-cut marine platform composed of Cretaceous and Tertiary bedrock (Fig 1.2; Dutro, 1981; Grantz and others, 1981, 1982). Cretaceous strata beneath the Gubik Formation include, in ascending order, the Torok Formation, composed of marine sandstone, shale, and siltstone; the Nanushuk Group, composed of marine and nonmarine shale, limestone, sandstone, coal and bentonite; and the Colville Group, composed predominately of nonmarine sandstone, siltstone and conglomeratic facies to the west and shallow-to deep-marine sandstone, siltstone and organic shale to the east. On the northeastern part of the coastal plain, several hundred km to the east of the area of this study, there are thick Tertiary strata, the Sagavanirktok Formation, composed of poorly consolidated marine and nonmarine shale, sandstone and conglomerate including the uppermost Nuwok member of marine silty shale, sandstone and mudstone (Nuwok Formation of MacNeil, 1957). These units represent the final infilling of a major east-west elongate trough known as the Colville Foredeep, a major structural feature bounded on the north by the anticlinal Barrow arch and on the south by the and uplifted Brooks Range. The infilling of this trough prograded

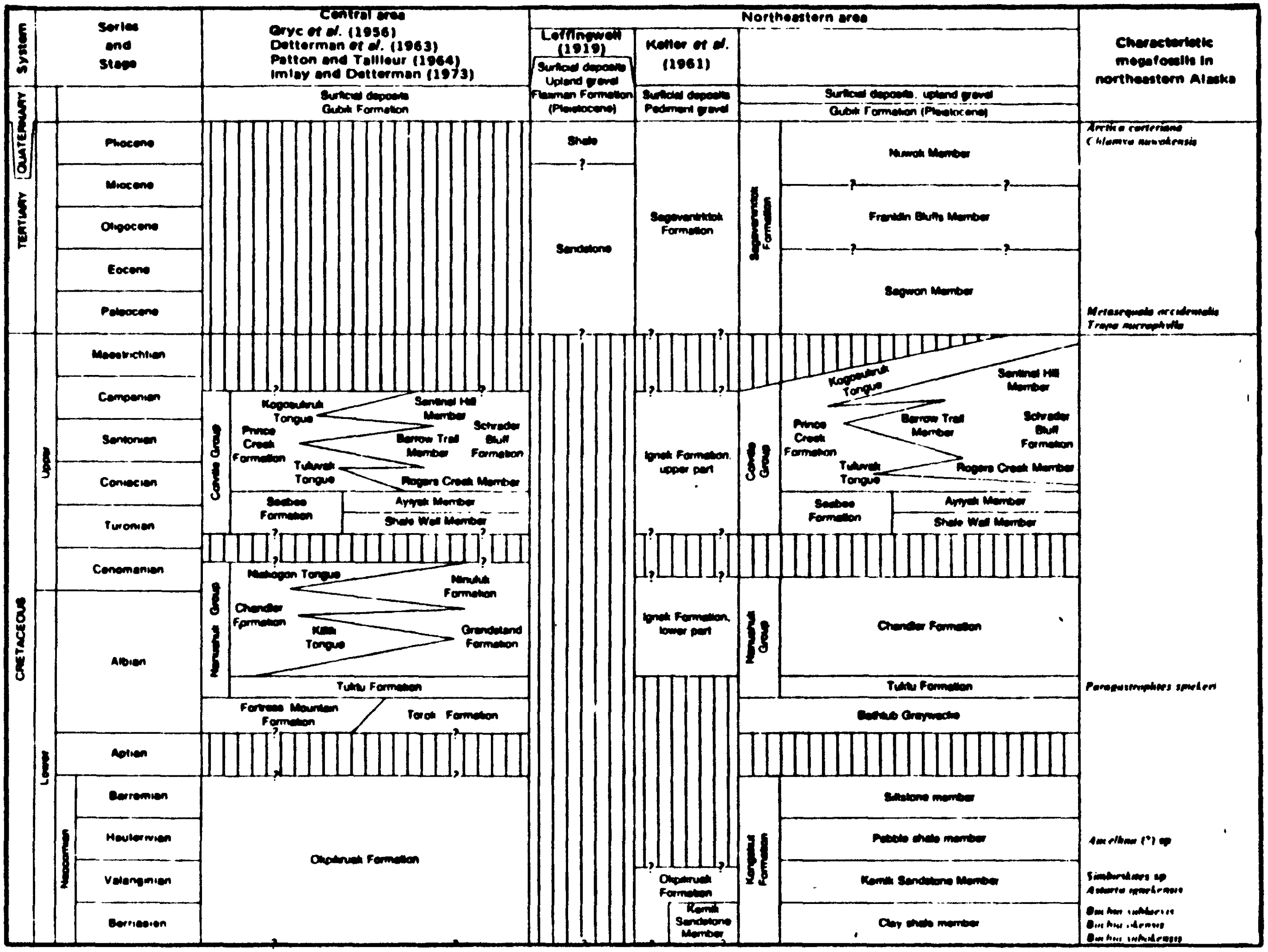


Figure 1.2. Bedrock stratigraphy of the North Slope (Neocomian-Recent) from Dutro, 1980).

northeastward from the Brooks Range throughout the upper Lower and Upper Cretaceous and Tertiary periods, and evidently breached the Barrow arch, allowing sediments of the upper Colville Group and the Sagavanirktok Formation to prograde onto the Beaufort shelf (Grantz and others, 1981, 1982). The western Arctic Coastal Plain appears to have undergone a great deal of erosion throughout Tertiary time, as Tertiary-age sediments are not known southwest of Peard Bay and are thin and discontinuous further north.

East of the Colville River, the predominant bedrock beneath the Gubik deposits includes portions of the Colville Group and the Sagavanirktok Formation. West of the Colville River, the basal Gubik sediments overlie a continuation of the Colville Group; however, at Barrow, erosion of the NW-SE axis of the Barrow Arch has exposed portions of the Nanushuk Group and the older Torok Formation (Fig. 1.3) (Grantz and others, 1982). From Walakpa Bay south to Peard Bay, the Nanushuk group crops out intermittently as nearly flat-lying, but gently undulating beds at the base of the coastal bluffs (Williams, 1983a). These are overlain by patches of shallow marine silty-clays of middle (?) Miocene age, which were discovered during the course of this study (see Plate I and II). Additional outcrops of the Nanushuk Group and minor constituents of the Colville Group were observed along the banks of the Kuk River near Wainwright and along the major reaches of the Utukok, Koklik, Epizetka, and Kukpowruk rivers inland from Icy Cape and Point-Lay (Chapman and Sable, 1960). Near Icy Cape and further southward, these beds become intensely folded into an east-west trending series of anticlines and synclines known as the Foreland Fold Belt which extends northwestward under the Chukchi Sea (Chapman and Sable, 1960, Plate I; Grantz and others, 1982).

Climate and Permafrost

The longest and most continuous climatic records for the Arctic Coastal Plain are those from Barrow and Barter Island where observations have been continuously maintained since 1949 and discontinuous records date back to approximately 1921. Supplemental data is available from a number of sites across the coastal plain including Wainwright and Point Lay at the western edge of the study area (Fig. 1.4).

The Arctic Coastal Plain can be generally described as an Arctic desert; winters are long, dry, and, cold; summers are short, moist, and cool. Winter conditions are fairly uniform across the north slope; however, in summer, the climate of the coastal regions is directly modified by the cooling effect of open water adjacent to the coast and the proximity of the arctic pack ice (Dingman and others, 1980). Consequently, mean daily summer temperatures dramatically increase with distance from the refrigerated coast (Haugen and Brown, 1980). Coastal fog is frequent throughout the summer when moisture is available from areas of open water and from puddles on sea ice. Fog is rare inland. A seabreeze from the open Arctic waters dominates the coastal climate in summer (Moritz, 1977).

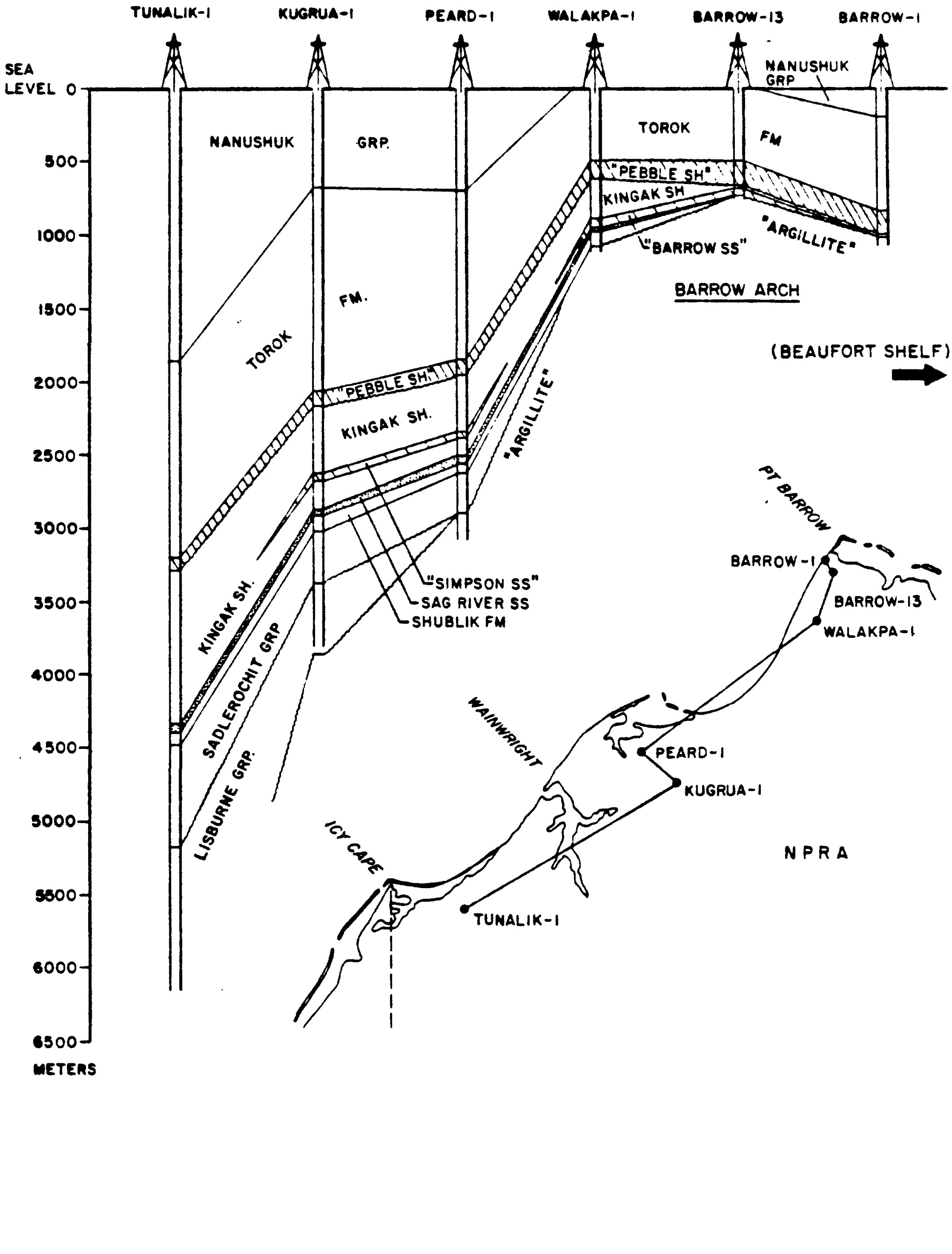


Figure 1.3. Cross-section of the bedrock stratigraphy based on test wells in National Petroleum Reserve in Alaska (NPRA) along the Chukchi Sea coast (from Grantz and others, 1982).

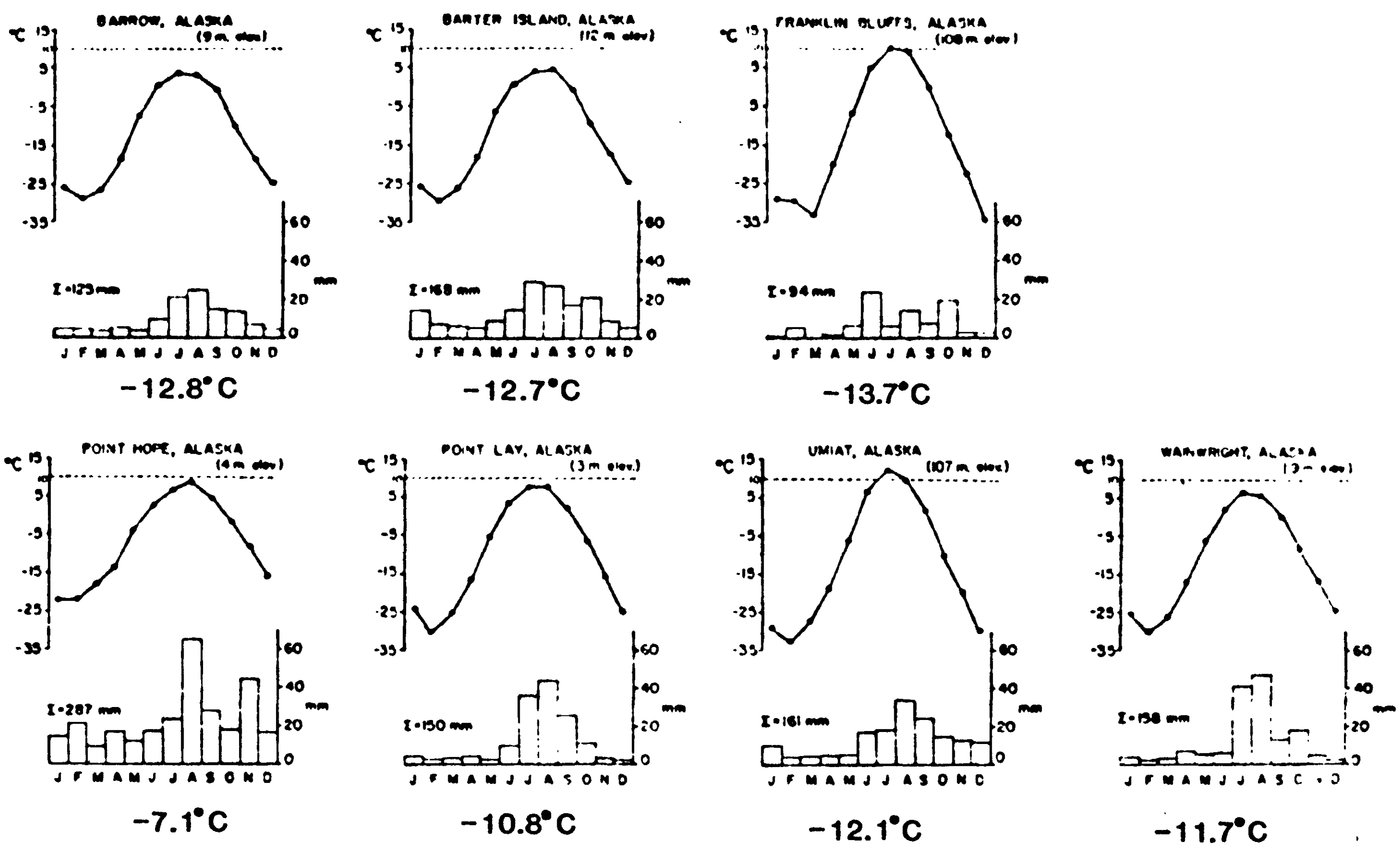


Figure 1.4. Climatographs for stations on the North Slope compiled by Nelson (1979). Diagrams illustrate the average Monthly temperatures and precipitation with snowfall converted to mm H_2O equivalent. Large number below each graph is mean annual air temperature (1949 - 1978).

The mean annual temperatures across the coastal plain range between -10.8° C at Point Lay at the western edge of the coastal plain to -13.7° C at Franklin Bluffs on the eastern coastal plain 50 km inland from Prudhoe Bay. Annual precipitation is somewhat uniform, averaging between 125 and 167 mm of water equivalent (w.e.) per year. Summer precipitation is greatest in August. Winter snowfall patterns, in contrast, peak twice; once in October and again in spring with monthly accumulations rarely exceeding 17 mm w.e. at Barrow.

The harsh temperatures of the coastal plain are accompanied by moderate wind speeds generally from the east to northeast, averaging 5 m s^{-1} (18 mph), which can cause blowing snow in winter (Dingman and others, 1980). Winds are generally less frequent inland. The coastal stations, Barrow and Barter Island, average only 1 percent and 4 percent calm days per year, respectively; Umiat which lies in a sheltered inland valley, averages 17.1 percent calm days per year (Skelkregg, 1975). Winds at Barrow rarely exceed 12 m s^{-1} (43.2 mph), even in January (Fig. 1.5).

Storms carrying significant precipitation usually enter the coastal plain region from the Soviet Arctic to the west or through the Bering Strait to the southwest. Storms moving northward from the Aleutian Islands through the Bering Strait during August deliver the greatest monthly amounts of precipitation (Skelkrey, 1974). Storms during August and September produce occasional storm surges along the Arctic Coast (Hume and Schalk, 1967; Wiseman and others, 1973).

Northern Alaska lies in the region of continuous permafrost, the southern boundry of which lies just south of the Brooks range at approximately the -6° C isotherm (Gold and Lachenbruch, 1973). Permafrost on the coastal plain varies in thickness from 200-400 m near Barrow to ca. 650 m near Prudhoe Bay (Péwé, 1975; Osterkamp and Payne, 1981) and occurs as relict permafrost beneath the recently submerged continental shelf of the Beaufort Sea (Barnes and Hopkins, 1978). During the summer months, the active layer typically thaws to a depth of only 30 cm which impedes drainage and produces the wet to moist tundra of the coastal plain. Ubiquitous ice wedges, ice-wedge casts, thaw lakes, pingos, and localized solifluction lobes, are all by-products of this frigid environment.

The depth and temperature gradient in permafrost varies across the coastal plain as illustrated by Gold and Lachenbruch (1973) and Osterkamp (1982) (Fig. 1.6, this study). Anomolies are readily apparent and illustrate that the correlation between permafrost depth and temperature is poor. For example, at Cape Thompson where the mean annual air temperature is -7° C, permafrost is 25% deeper than that at Cape Simpson where the mean annual air temperature is close to -12° C. Because geothermal heat flow is approximately the same at all sites, Gold and Lachenbruch (1973) and Osterkamp (1982) have shown that variations in permafrost

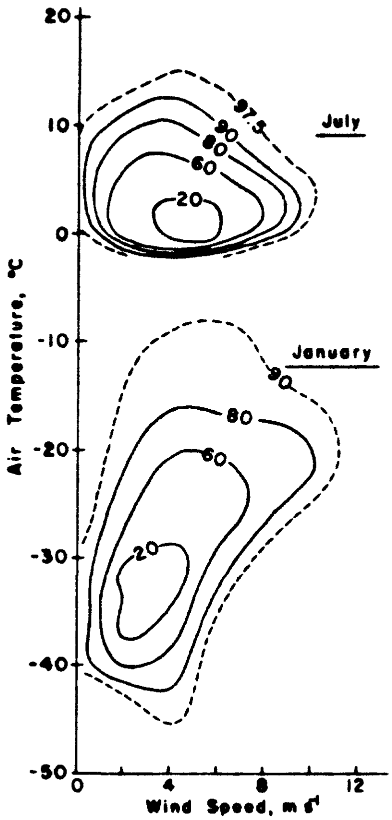


Figure 1.5. Cumulative frequencies of hourly air temperatures and wind speeds at Barrow from 1945-54. Values expressed as the cumulative percentage frequencies of the whole distribution (Dingham and others, 1980).

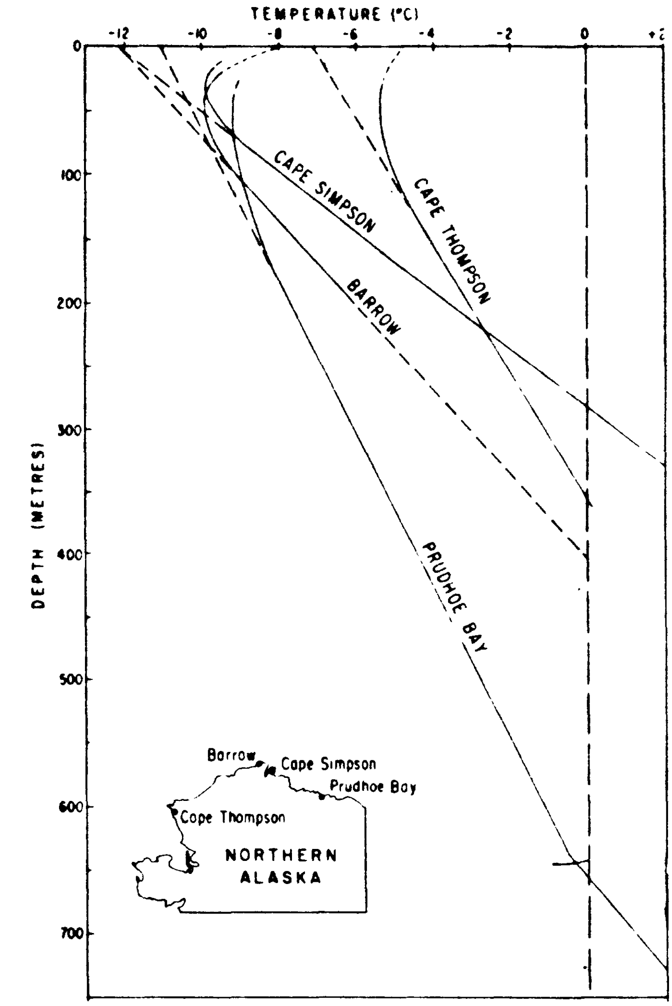


Figure 1.6 Profiles of measured temperature in permafrost on the Alaskan Arctic Coastal Plain (solid lines). Dashed lines are extrapolated (from Gold and Lachenbruch, 1973).

thickness depend largely upon the thermal conductivity of the sediments and, to a lesser extent, on the surface temperature.

Vegetation

The vegetation on the Arctic Coastal Plain is herbaceous tundra and can be subdivided into zones of predominantly wet, moist, and dry tundra vegetation types. These areas are typified by grass, sedge and moss communities mixed with lichens and prostrate woody shrubs (Britton, 1957). The wide variety of habitats including low and high-centered polygons, thaw lakes, riverbanks, and dune fields create a mosaic of complex plant communities characterized by parameters of microclimate temperatures, soil pH and soil moisture (Webber and others, 1980). Other detailed studies outlining interrelationships between vegetation types and micro-, meso-, and macro-relief habitats include those of Wiggins (1951), Cantlon (1961), Wiggins and Thomas (1962), and Britton (1967). A general list of common vegetation communities for the entire coastal plain appears on Table 1.2 (D.A. Walker, INSTAAR, oral communication, 1983; Walker, 1982).

Methods

Field work for this study was conducted during the open-water months of July and August, 1980 and 1981, to support a detailed analysis of the coastal stratigraphy by using a Zodiak inflatable boat. This work was carried out between Peard Bay and Barrow and within Wainwright Inlet and lower Kuk River (Fig. 1.1). Base camps for an assistant and myself were established at a series of sites spaced 20-30 km apart to allow quick access to nearby coastal cliff exposures. Field studies in the areas of the Utukok, Kokolik, and Kukpowruk Rivers south of Wainwright and eastward as far as the Meade River, were conducted by helicopter during 10 days in late August and early September, 1981.

Field strategy was largely aimed at the description and measurement of stratigraphic sections and the tracing and correlating of individual lithologic units and/or facies changes between sections in Skull Cliff and other bluffs between Barrow and Peard Bay. Seventy-four stratigraphic sections (a density of approximately 1 section km^{-1}) were studied between Barrow Village and Peard Bay. In addition, thirty-eight stratigraphic sections and less well-exposed surficial sediments were examined at many sites along river banks, shores of thaw lakes, and frost scars on the tundra surface scattered across the western coastal plain. At each section the physical characteristics of each lithologic unit were noted, especially with respect to grain-size, sorting, and bed forms. Where they occurred, fossil mollusks were collected from each lithologic unit for amino acid analysis and biostratigraphic studies. Mollusks were identified by Louie Marincovich and George Kennedy assisted by Dale Russell, all of the U.S. Geological Survey in Menlo Park. Similarly, sediment samples were taken for microfaunal analysis, grain-size analysis and pollen studies.

Table 1.2. Common plant forms on the Arctic Coastal Plain through a catena of wet aquatic to dry habitats (Walker, 1982; 1983, pers. comm.)

Habitat	Species	Common names
Aquatic Tundra (water 30 cm - 1 m deep)	Arctophila fulva	(pendant grass)
	Carex aquatilius	(aquatic sedge)
Wet Tundra	Carex aquatilus	(aquatic sedge)
	Eriophorum angustifolium	(common cotton-grass)
	E. russeolum	(russet cotton-grass)
	Pedicularis sudetica ssp. albolabiata	(Sudetan louse-wart)
	Saxifraga hirculus	(bog saxifraga)
	Caltha palustris	marsh marigold
	Potentilla palustris	(marsh five-finger)
	Drepanoclaudus spp.	(moss)
	Scorpidium scorpiodes	(moss)
	Campylium stellatum	(moss)
	Calliergon spp.	(moss)
	Sphagnum spp.	(moss)
Moist Tundra	Salix arctica	(arctic willow)
	S. pulchra	(diamond leaf willow)
	Dryas integrifolia	(arctic avens)
	Betula nana	(Dwarf birch)
	Vaccinium uliginosum	(bog blueberry)
	Arctastaphylos rubra	(Bearberry)
	Carex aquatilus	(aquatic sedge)
	Carex bigelowii	(Bigelous sedge)
	Luzula arctica	(arctic wood-rush)
	Eriophorum vaginatum	(sheathed cottongrass)
	E. angustifolium ssp. tristi	
	Arctagrostis latifolia	(wideleaf arctogrostis)
	Petasites frigidus	(lapland butterbur)
	Chrysanthemum integrifolium	(entireleaf Chrysanthemum)
	Tomenthypnum nitens	(moss)
	Dicranum elongatum	(moss)
	Sphagnum spp.	
Dry Tundra	Salix rotund folia	
	Dryas integrifolia	
	Arctostaphylos rubra	(bearberry)
	A.alpina	
	Ledum palustre ssp. decumbens	
	Luzula arctica	(arctic wood-rush)
	Saxifraga oppositfolia	
	Polytrichum sp.	(moss)
	Rhacomitrium languinosum	(moss)
	Thamnoli spp.	(lichen)
	Ochrolichia frigida	(lichen)
Shrub communities	mostly Salix spp.	
	Betula nana spp. exilis	(somewhat inland)

Foraminifera were identified by Kristen McDougall, U.S.G.S., Menlo Park, and ostracodes by Elizabeth Brouwers, U.S.G.S., Denver. Organic material and peat were sampled for additional pollen studies and plant microfossil identifications; several samples were examined by Robert E. Nelson, Colby College. Finally, wood (including, rare in situ growth forms and fossil driftwood) was collected for radiocarbon dating and also for amino acid analysis at the University of Alberta, Edmonton. Preliminary maps of remnant beach features and paleoshorelines were made using aerial photographs and Side-Looking Radar imagery (SLAR) provided by John Cady, U.S.G.S., Denver, and these were subsequently field checked during the helicopter reconnaissance. At several sites, oriented samples of fine-grained sediment were collected for paleomagnetic studies.

The primary objective of this study was to define the stratigraphy and geochronology of the Gubik Formation across the western coastal plain. Due to the disjunct nature of inland exposures, the size of the field area, and most importantly, the fact that successive transgressive units are not lithologically distinct, stratigraphic control was provided by establishing an amino acid chronology for the deposits. The stratigraphic potential of this technique is well documented (see Schroeder and Bada, 1976; Williams and Smith, 1977; Hare and others, 1980; Wehmiller, 1982a, for reviews) and in this study provided the most useful means of differentiating superposed marine-transgressive deposits. Because the kinetic parameters associated with shell protein diagenesis are not entirely understood, absolute age estimates are limited to first approximations or "ball park" age estimates. Nevertheless, the method is invaluable as an index of relative age. Indirectly this dissertation is intended to supplement what is currently known concerning the applications and limitations of amino acid geochronology to Quaternary problems at high latitudes.

Stratigraphic Nomenclature

The North American Stratigrapic Code (North American Commission on Stratigraphic Nomenclature or ACSN, 1983) now recognizes six material categories for defining stratigraphic units; lithostratigraphic, biostratigraphic, lithodemic, magnetopolarity, pedostratigraphic and allostratigraphic units. The first two categories appeared in the previous code (ACSN, 1970) and are the only descriptive units recognized by the International Guide (ISSN, 1976). The remaining categories, in contrast, are new units introduced by the ACSN (1983) to accomodate a number of inadequacies that emerged through the use of the International Guide and the previous code (ACSN, 1970). Of particular concern was the lack of adequate nomenclature for subdividing deposits of Quaternary age that are biostratigraphically similar and lithologically indistinguishable (Brigham, 1980, p. 10-16).

The allostratigraphic unit (allo meaning "other, different") developed and introduced by the ACSN (1983, Article 58) is intended, in part, to facilitate the subdivision of unconsolidated sediments. It is defined as a "mappable stratiform body of sedimentary rock that is defined and identified on the basis of its bounding discontinuities." Its specific purpose is "to distinguish superimposed, contiguous, and/or geographically separated, discontinuity-bounded deposits of similar lithology." Geologic history, genesis, age, or inferred depositional time spans can not be used to define an allostratigraphic unit but "they may influence the choice of the unit's boundaries." Technically then, age, or the chemical expression of age in the context of amino acid geochronology, may be used as a criterion for choosing relevant boundaries to define allostratigraphic units.

In principle, the subdivision of unconsolidated sediments on allostratigraphic criteria is a useful addition to the code. In practice, however, the additional nomenclature of alloformations and allomembers seems redundant and confusing especially when used adjacent to the traditional nomenclature. If applied to the Gubik Formation, it would be necessary to establish a new alloformation name and subdivide the deposits into allomembers. Members of my committee and colleagues within the U.S. Geological Survey agreed that they did not favor the adoption of these terms.

In this study, the lithostratigraphic Gubik Formation has been subdivided into members based on allostratigraphic criteria, i.e. they are distinguished by bounding transgressive disconformities or erosional unconformities and can be differentiated by the extent of amino-acid diagenesis observed in fossil mollusks from each unit. In a stratigraphic sequence where transgressive marine units lie disconformably, the extent of amino acid diagenesis in enclosed fossil mollusks has been used to chose relevant boundaries (Fig. 1.7). Formal adoption of the new Code by the U.S. Geological Survey will require that the nomenclature used here be altered to strictly accommodate the guidelines of the Code.

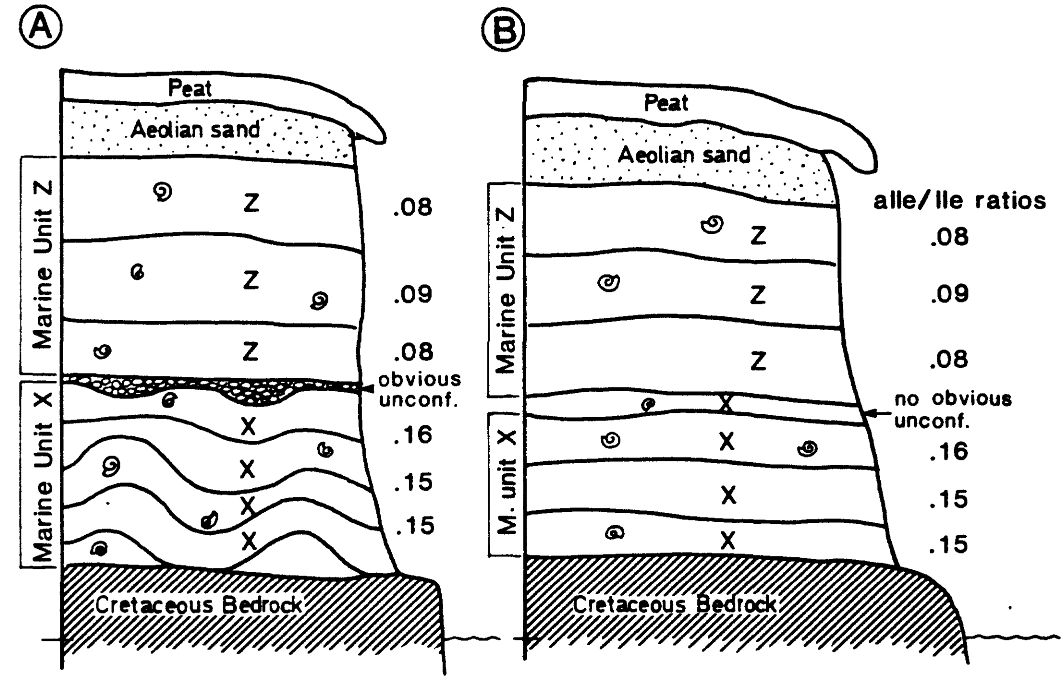


Figure 1.7. Differentiation of members of the Gubik Formation by allostratigraphic criteria. In case A, the identification of the discontinuities is somewhat obvious and aIle/Ile values of $x^{\pm}$ for unit x and $z^{\pm}$ for unit z substantiate the age difference between units X and Z. In Case B, the disconformity between units x and z is not obvious on conventional stratigraphic grounds alone. Here, the aIle/Ile values (as a chemical measure of age) influence the choice of the units boundaries.

CHAPTER II

GENESIS OF LITHOFACIES

Introduction

The Gubik Formation consists of a complex of interbedded marine, fluvial, eolian, thaw lake and in situ peat deposits. West of the Meade River, offshore marine and beach facies comprise a significant portion of the Gubik deposits. These sediments are best known between Peard Bay and Barrow, where sea bluffs over 70 km in length expose the most complete sequence of marine deposits known on the North Slope. These bluffs, in fact, contain the most complete sequence of superposed late Cenozoic marine units known in northern and western Alaska and consequently provide the stratigraphic key for the lithological and chronological subdivision of the Gubik Formation. This chapter describes the marine and non-marine lithofacies exposed in the bluffs southwest of Barrow. In addition, the discussion includes observations concerning the nature and preservation of marine transgressive deposits on the Arctic Coastal Plain.

Black (1964) made the only previous attempt to subdivide the Gubik Formation. From oldest to youngest, his subdivision (Black, 1964, p. 65) includes:

the Skull Cliff unit - sticky or greasy, generally poorly sorted, blue-black to dark-gray, marine clay-silt-sand-cobble unit; possibly glacially derived, in part, and deposited unconformably on the Upper Cretaceous rocks in much of the coastal plain west of the Colville River.

the Meade River unit - white, yellow, buff, or light-tan sand; generally clean and well sorted; marine; conformable to discomformable on the Skull Cliff unit but in places deposited on the Cretaceous rocks. Locally, whipped into surface dunes; in the south and southeast commonly loesslike, and

the Barrow unit - poorly sorted to well-sorted mixtures of clay, silt, sand, and gravel; unit grades from yellow, tan, and brown to black; in part, contemporaneous with and grades laterally into the Meade River sand, and, in part, younger; rarely deposited directly on Upper Cretaceous rocks; mostly marine; locally, in upper part, fluvial deposits and lacustrine deposits are characteristic.

Black recognized and correlated the units outlined above largely on the presence or absence of clay and gravel. Although his lithologic motif describes many general aspects of the Gubik Formation, particularly along the Skull Cliff bluffs; it is an oversimplification. In many exposures the Gubik Formation is composed of superposed offshore shelf sediments, consequently marine events of different ages are commonly lithologically indistinguishable. Moreover, transgressive shoreface erosional surfaces are not always obvious as unconformities. This study indicates that Black's lithologic units do not recognize the complex lithofacies associated with successive transgressions and that his units are of different age in different places across the coastal plain.

During the course of this study seventy-four stratigraphic sections (a density of approximately 1 section km^{-1}) were studied, between Barrow village and Peard Bay (Fig. 2.1a). At each section, the physical characterictics of each lithofacies were described including observations on the nature of disconformaties, and the lateral extent of these features (Plate I and II). Sediment and microfaunal collections were made from most units represented at each site. Thirty-eight additional disjunct sections were described from inland sites across the coastal plain west of the Meade River (Fig. 2.1b). Mollusks were collected for amino acid analysis from all sections (Chapter III) and to examine the biostratigraphic, as well as, paleoclimatic significance of the fossil assemblages (Chapter V). All of the grain-size information contained therein are qualitative in nature.

Modern Shallow Shelf and Nearshore Sedimentation

Modern nearshore and offshore sediments of the Chukchi and Beaufort sea shelves provide the only useful analogues for evaluating the marine stratigraphy of the Gubik Formation. Although the sedimentary processes operating on arctic shelves are similar to those observed at lower latitudes, the persistence of sea ice nine to ten months of the year markedly retards sediment movement. Most sediment transport occurs during the open-water months of July, August, and September at which time storm waves and wind-driven currents dramatically influence the coast and shallow-shelf environments (Short and others, 1974; Barnes and Reimnitz, 1974). During the winter months, in contrast, the sea is ice-covered and stream and river discharge is reduced, resulting in a shelf environment nearly closed to sedimentation. Slight currents persist (McManus and others, 1969), and some fine-grained sediment moves across the arctic shelves throughout the winter.

Investigations of sediment movement on the arctic shelves of Alaska have, until recently, been confined to the Beaufort Sea where interest was first stimulated by petroleum exploration.

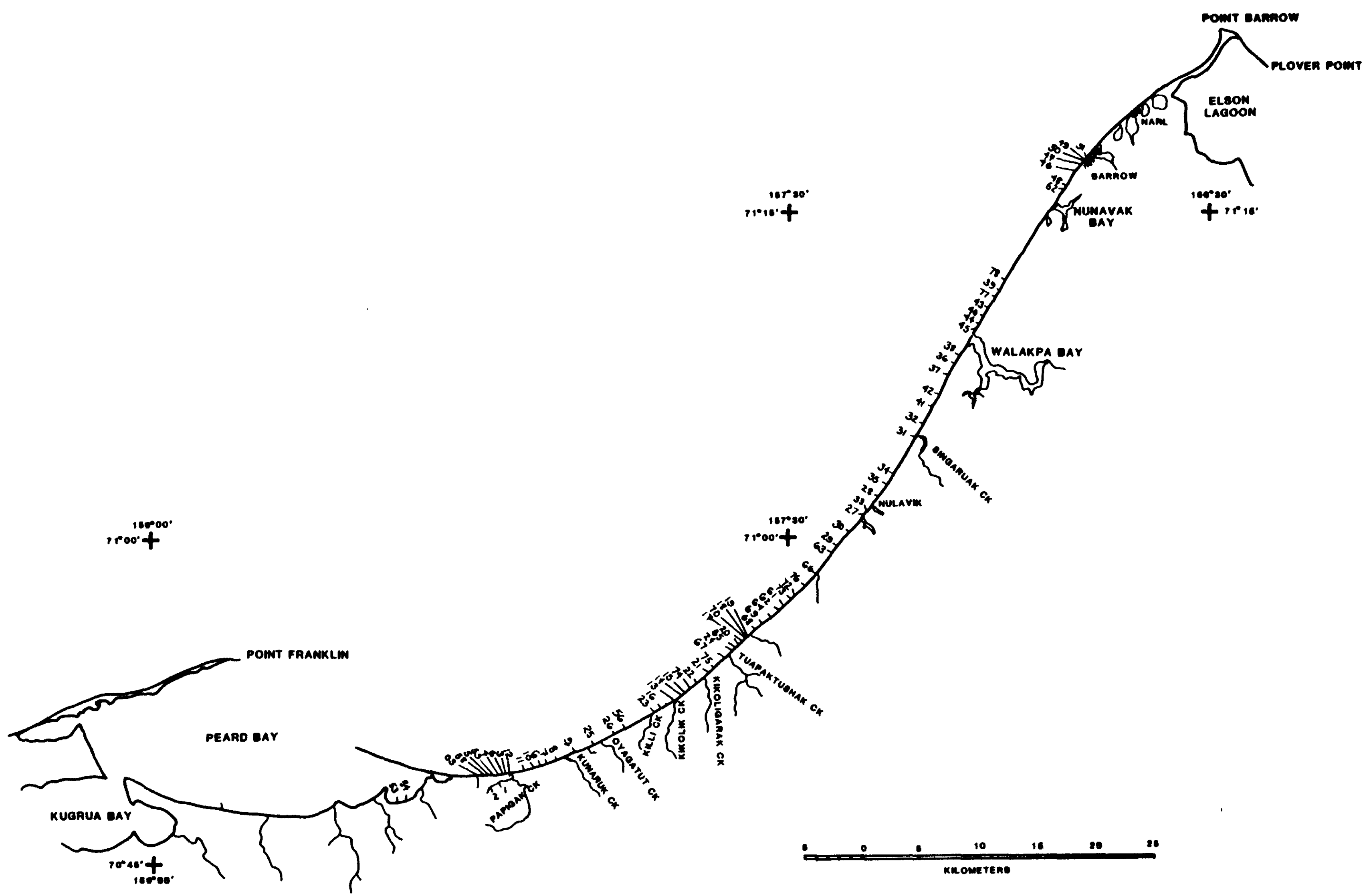


Figure 2.1 a. Location of stratigraphic sections shown on Plate I and II.

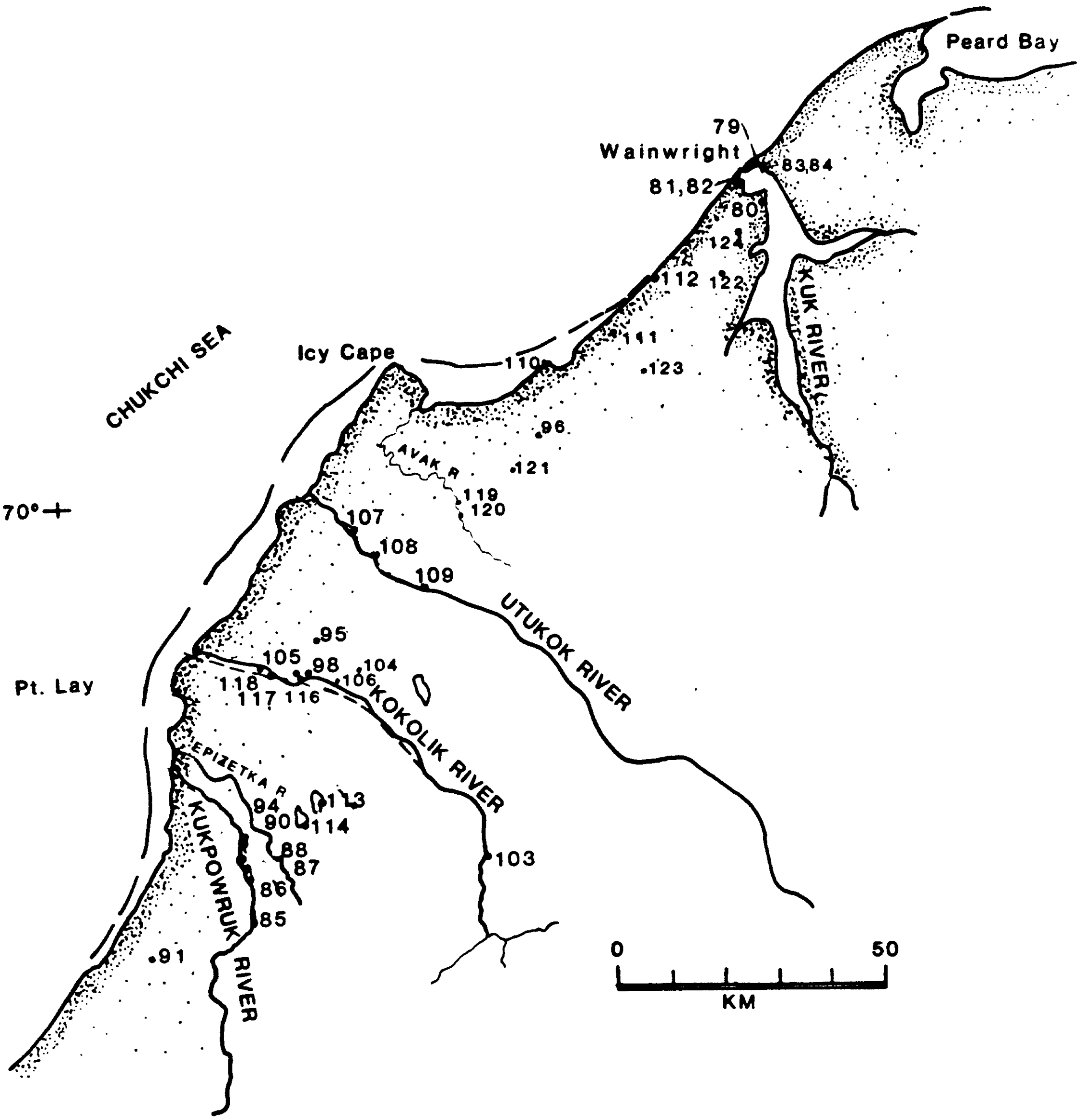


Figure 2.1b. Location of stratigraphic sections southwest of Peard Bay. (All sections not shown.

Studies of sediment movement on the Chukchi shelf are limited and it is difficult to extrapolate the sediment patterns in the Beaufort Sea to the Chukchi Sea. Extreme diversity across short distances characterizes the sediment of both arctic shelves (Barnes and Hopkins, 1978).

Barnes and Reimnitz (1974) recognize five facies zones across the Beaufort shelf, which have characteristic textural properties and sedimentary structures resulting from the influence of currents, waves, and the action of ice as follows: (1) The nearshore zone from the beach to 3-4 m depth is complexly effected by waves and ice. Its sediments of sand, silt, and gravel are derived from river effluents (cf. Naidu and Mowatt, 1983) and the coastal erosion of older sediment. (2) The inner shelf area from the nearshore zone to the winter ice shear zone at about the 20 m bathymetric contour is influenced by grounded ice, but is dominated to a greater extent by waves and longshore currents. Sediments in this zone are better sorted and consist of layers of silt and sand affected by the seasonal wave regime. (3) The central shelf of the Beaufort Sea from -20 m depth to the shelf break is influenced by ice-gouging that results in the deposition and reworking of unstructured gravelly muds. Much of the fine-grained material in this area is derived from local rivers or ice-rafting (Naidu and others, 1971), whereas the striated coarse-grained material is interpreted as relict ice-rafted debris that originated from glaciers calving into the eastern Beaufort Sea from the Canadian Arctic Archipelago (Rodeick, 1975). The amount of coarse-grained material on the Beaufort seabed diminishes westward.

The final two facies zones recognized by Barnes and Reimnitz (1974) subdivide the deeper portions of the shelf. (4) The shelf edge is similar to the central shelf in being dominated by relict surficial gravel apparently ice-rafted from the same source as that on the central shelf. In contrast to the central shelf, however, the fine-grained sediments are scarce and appear to be continually removed, perhaps by strong currents (Barnes and Reimnitz, 1974). Below the shelf edge at depths greater than 200 m, the surficial sediment of the continental slope consists of bioturbated "silt-clay", often with faint laminations. A distinct lack of coarse-grained material is characteristic of this zone and remains unexplained, although this zone may be the main depot for fine-grained sediments swept from the central shelf and shelf edge, effectively burying relict ice-rafted sediments (Barnes and Reimnitz, 1974, pg. 467).

Knowledge of the sediment distribution on the shallow Chukchi Sea shelf (average depth, ca. 50 m) is derived primarily from the work of Creager and McManus (1967), McManus and others (1969, 1983), Phillips and others (1984) and is summarized by Sharma (1979). Creager and McManus (1967) and McManus and others (1969) provide a generalized description of modern sedimentation across the northeastern Chukchi Sea (Fig. 2.2). Sediments at the surface are largely composed of sand that becomes finer-grained northward toward Barrow with minor admixture of sandy silt and

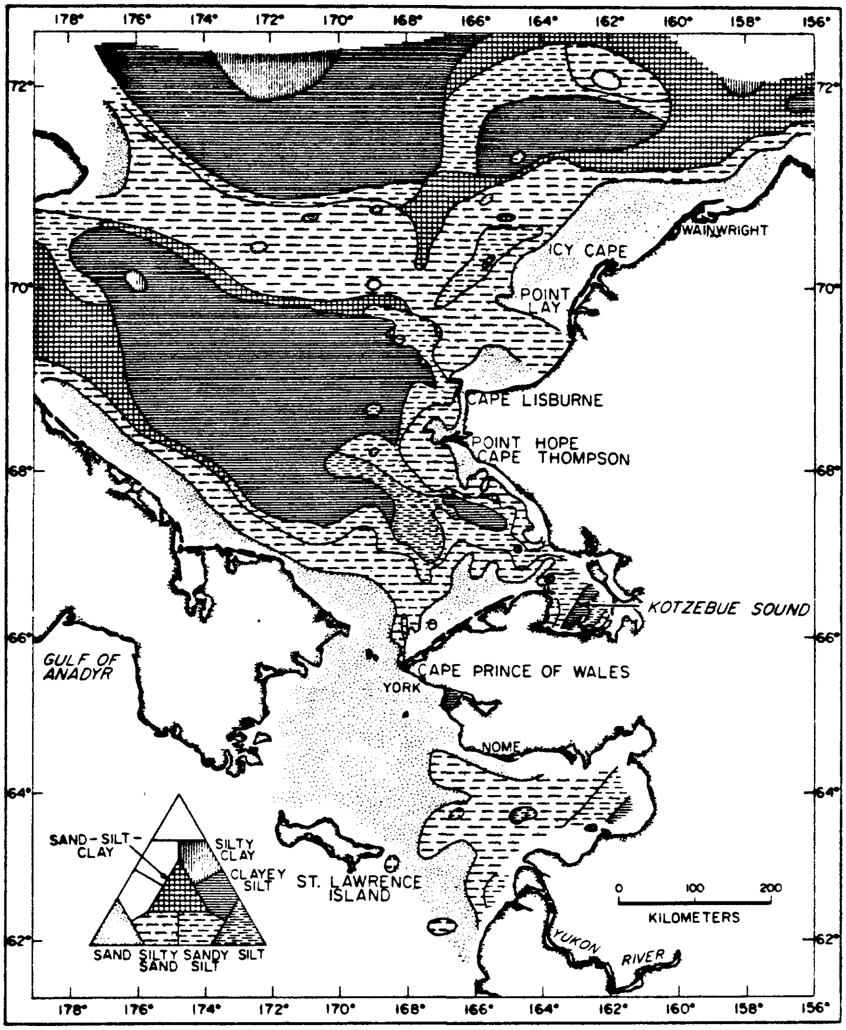


Figure 2.2. Textural character of seabed sediments in the Chukchi Sea (reproduced from Creager and McManus, 1967).

silty sand, becoming even finer-grained more than 100 km offshore. Well-sorted relict sand and some gravel exist at water depths of 25 to 35 m offshore from Pt. Lay and Icy Cape. As in much of the Chukchi and Bering seas, unconsolidated Quaternary sediment across the northeastern Chukchi Sea is less than a few meters thick and may be absent, but can locally reach thicknesses of 12 m (Moore, 1964).

Phillips and others (1984) conducted a detailed study of the unconsolidated marine sediment shallower than 20 to 30 m water depth on the inner Chukchi Sea shelf between Wainwright and Skull Cliff. Their study substantiated that surficial sediment in this area is extremely thin. The deposits thin to less than 2 m 5 km offshore of Pt. Belcher and to only 1 m within 0.5 km offshore of Skull Cliff (Fig. 2.3). Only in areas of converging currents where velocities diminish, e.g. off Point Franklin or Icy Cape, does sediment accumulate to thicknesses of 10-15 m. These large capes are formed where wave and current patterns converge causing progradation at the points parallel to the northeasterly longshore current (Alaskan Coastal Current) (Short, 1975, 1979; Sharma, 1979).

Textures of bottom sediment between Skull Cliff and Wainwright range from the silt and clay-dominated lagoonal deposits in Peard Bay to pure gravel, probably relict or lag, at 30 m depth along the coast. The extreme local diversity is well illustrated near Skull Cliff where grab samples varied from 21% to 96% sand and from 4% to 77% silt and clay (Phillips and others, 1984).

The paucity of Pleistocene sediment just offshore in the Chukchi Sea is also reflected by the thin blanket of marine sediments perserved on shore across the western coastal plain. The Gubik Formation reaches a maximum thickness of only 17.5 m over Cretaceous bedrock at Skull Cliff where as many as 3 transgressive episodes are present in superposition. More typical are sediment thicknesses of 8 to 10 m representing one, and often two, marine transgressive episodes; these are typically overlain by 0.5 to 2 m of eolian sand or thaw lake facies. The erosional capacity of the most recent Holocene transgression probably provides a suitable analogue to the amount of shoreface erosion that took place during previous transgressions, leaving at best only a thin veneer of older sediments preserved over the undulating bedrock surface. In addition, this transgression demonstrates that the initial sediment cover following submergence can be extremely thin, in places limited to a lag of gravel on scoured bedrock.

Lithostratigraphic subdivision of the Gubik Formation

The sediments of the Gubik Formation can be subdivided into numerous depositional facies of marine and non-marine origin; I recognize eight primary lithofacies, each of which represents a distinct depositional environment or mode of deposition but is independent of chronology. Each facies is distinguished on the basis of textural characteristics and primary sedimentary

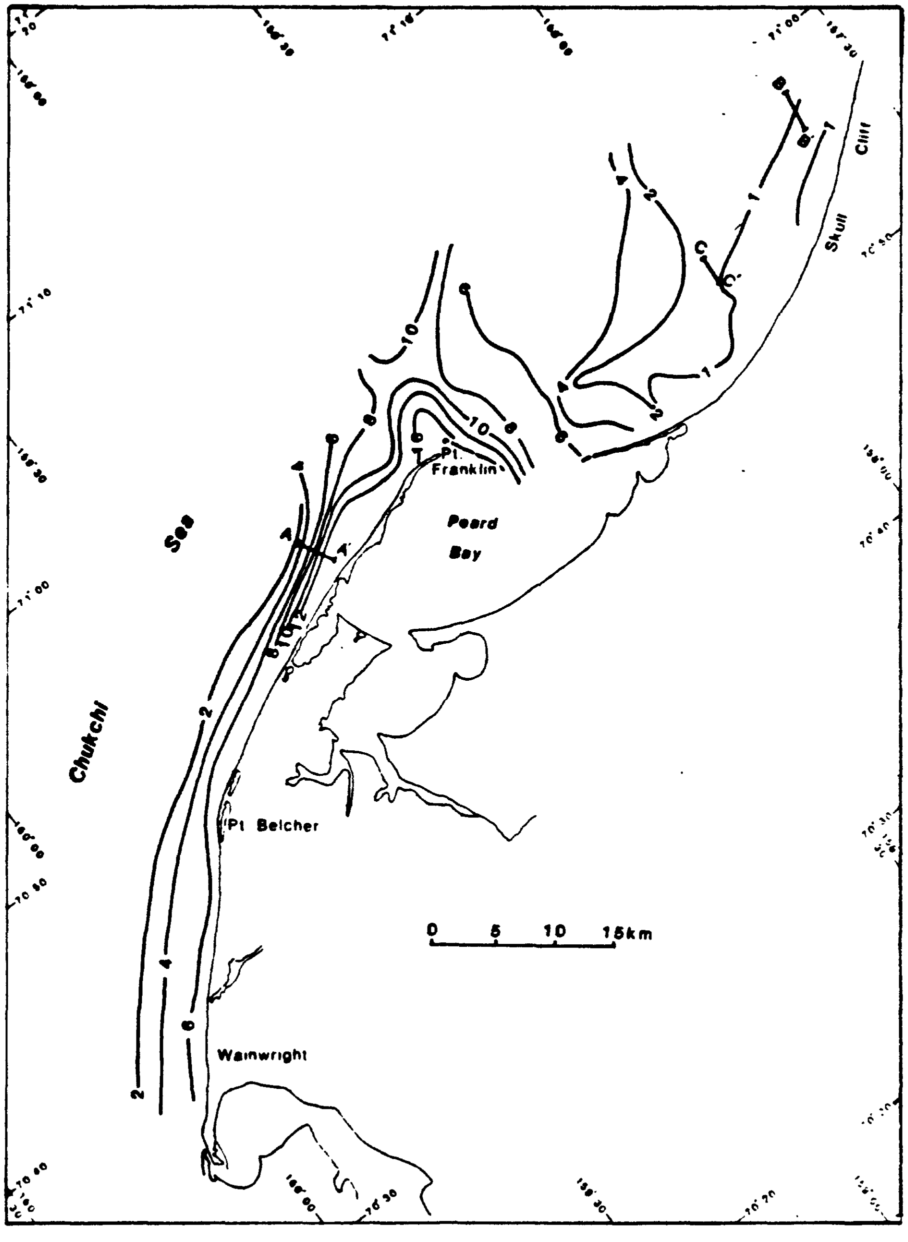


Figure 2.3. Isopach map (in meters) of Quaternary sediments overlying Cretaceous bedrock between Skull Cliff and Wainwright (Figure from Phillips and others, 1984). Unconsolidated sediments thin rapidly offshore.

structures, in addition to information assembled concerning the habitat requirements of enclosed micro- and macrofaunas (Chapter V) and analogues with the modern shelf and nearshore environments. The following discussion, that includes a brief description of the bedrock exposed along Skull Cliff, is keyed to the stratigraphic sections shown on Plates I and II attached.

Bedrock exposures

Cross-bedded arenaceous sandstone and mud shale of the Nanushak Group crop out at the base of bluffs from Peard Bay to Sec. 28 near Nulavik. These consolidated rocks of lower Cretaceous age project up to 7 m above sea level (a.s.l.) and where present severely retard the rate of cliff retreat and subsequent wave-induced thermal-erosion of the overlying sediments. The top of bedrock is marked by an undulating erosional surface which is superimposed onto broad structural undulations.

A discontinuous unit of massive, plastic, blue-gray clay to silty clay overlies the Cretacous bedrock between Peard Bay and Sec. 42, some 5 km south of Walakpa Bay (Fig. 2.4). The unit is internally structureless and macrofossils have not been found. Because of the very stiff, putty-like consistency of the clay, exposures are difficult to excavate. The best exposures of the clay are from Section 7 to Section 5 between the mouth of Kunarak Creek and Peard Bay (Plate I) where the unit reaches its maximum exposed thickness of 7 m and between Section 32 and Section 35 north of Nulavik (Plate II) where it is thinner. This unit forms part of Black's (1964 p. 77) original Skull Cliff unit and blankets the erosional surface of the underlying Cretaceous bedrock. Foraminifera washed from the clay at Sec. 10, 22, and 64 (Fig. 2.5 and Table 5.4) indicate that the sediment is of Miocene age and probably equivalent to the middle Miocene Nuwok Member of the Sagavanirktok Formation exposed at Carter Creek and elsewhere in northeastern Alaska (Detterman and others, 1975) (the foraminifera were examined by Kristin McDougall, U.S.G.S.). Paleoecologically, the fauna is interpreted as a marginal marine facies, characteristic of shallower water environments than those found in the Nuwok Member at Carter Creek (located some 490 km east of the sites at Skull Cliff). This represents the first documentation of Miocene marine sediments beneath the coastal plain west of the Colville River.

It may be reasonable to refer to the Miocene clay at Skull Cliff as the Nuwok member, because they are lithologically and faunally similar. The two areas are, however, separated by several hundred kilometers and further study will be necessary before the correlation can be confirmed. Because the Skull Cliff clay is so widespread, I have given it an informal provincial name, the Papigak clay, named after Papigak Creek near Peard Bay where it is well exposed.

Blue-gray clay similar to the Papigak clay exposed at Skull Cliff was also found beneath unconsolidated sediments near Karmuk Point in the Kuk Inlet (Fig. 2.5, see also Fig. 4.17). A single

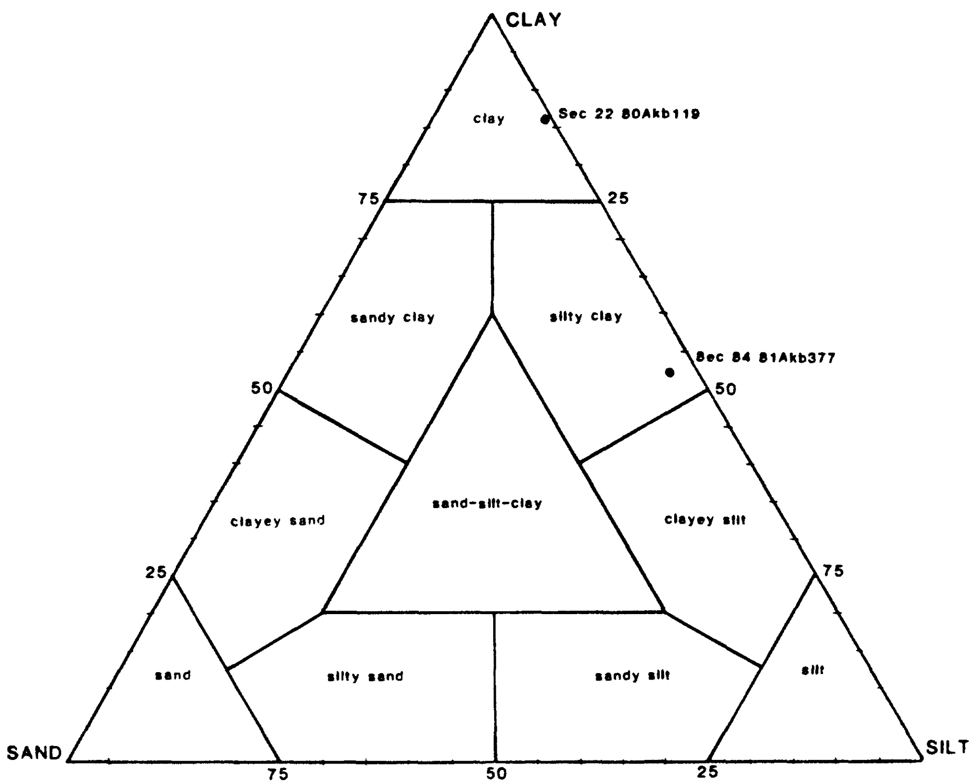


Fig. 2.4. Grain size distribution of two samples of the blue-gray clay thought to be of Miocene age.

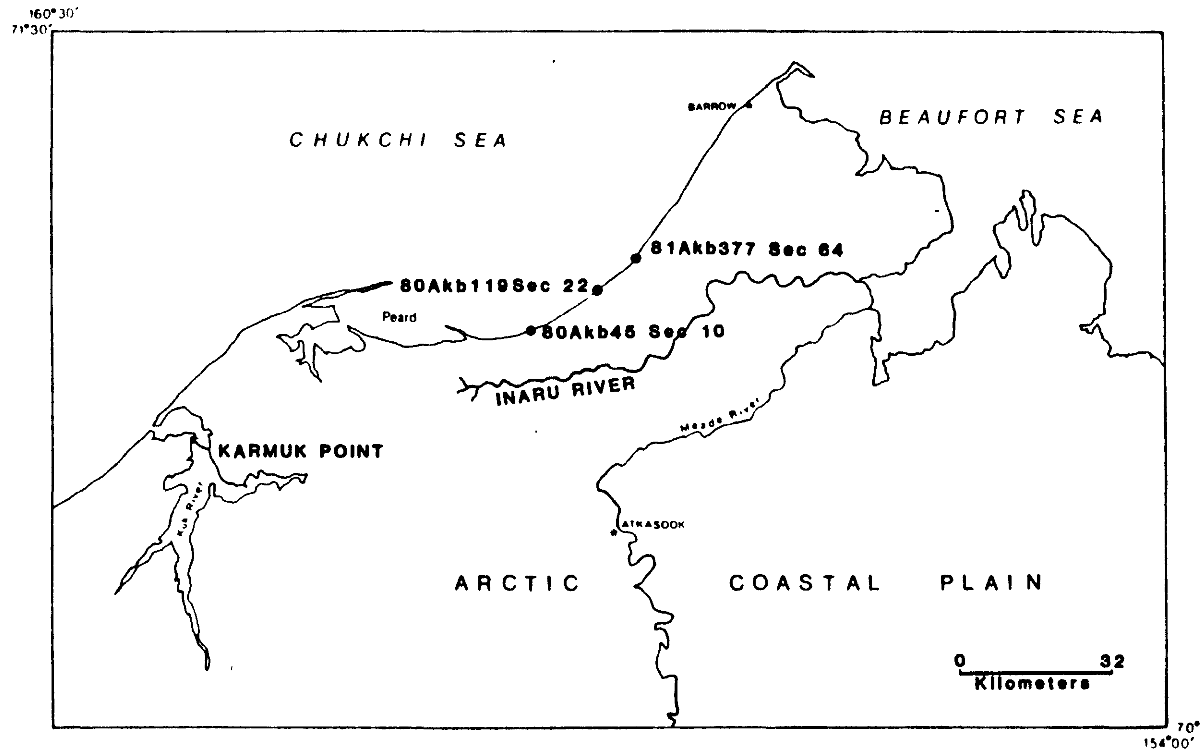


Figure 2.5. Locations of blue-gray clay containing a Miocene age foraminiferal assemblage. Numbers above indicate my field numbers for samples and the sections (Plate I and II) from which the samples were taken.

sediment sample from the Karmuk Point section was washed for microfossils, but like many samples of the clay along Skull Cliff, it was barren. The blue clay in the area of the Kuk Inlet was first described by Webber (1947 p.9) who hypothesized that the sediment represented "a post-Upper Cretaceous soil, possibly of Tertiary age, formed by long continued weathering at or near sea level." James Gilluly described a similar clay occurring along the Inaru River (Fig. 2.5), some 12-15 km inland from Skull Cliff (Smith and Mertie, 1930, p. 241). Possibly all of the clayey sediments in these areas are correlative.

Miocene sediments perhaps marking a contemporaneous marine transgression are known from two other areas surrounding the Arctic Basin: the Beaufort Formation on Meighan Island in the Canadian Arctic Archipelago (Matthews, 1976) and the Mackenzie Bay Formation of the Mackenzie Delta region (Young and McNeil, 1982). The paleogeographic importance of these Miocene sites is discussed by Hopkins (in press) who suggests that they represent the peak Miocene transgression of Vail and others (1977).

Gravel/Cobble Lag Lithofacies

A lag deposit composed of sandy silty gravel or cobble gravel is commonly found along portions of Skull Cliff where it directly overlies Cretaceous sandstone and, less commonly, the Papigak clay. In this position, it marks a major unconformity between the regional bedrock and deposits of the unconsolidated Gubik Formation. Elsewhere, similar lag deposits are found marking unconformities between successive marine units suggesting episodes of marine erosion between accretionary events.

The nature of the lag gravel varies from section to section depending upon the erodibility of the underlying material, i.e. whether the underlying material is Cretaceous sandstone, Papigak clay, or older marine sediment. Above Cretaceous sandstone, (for example, at Sec. 61, 62, and 17-19) the lag gravel ranges in thickness from 5 to 35 cm and consists of cobbly, sandy gravel commonly including subangular or subrounded clasts 10 to 15 cm in diameter. The cobbles include pink and white granite, dolomite, red sandstone and mafic igneous erratics as well as cobbles of Cretaceous sandstone of local origin. Locally these materials are stained with iron-oxides and in some sections the matrix sand and gravel is firmly cemented.

Lag gravel positioned over Papigak clay, in contrast, forms a more diffuse unit commonly consisting of more clayey, silty gravel and sand. The lithologic contact between the clay and the gravel is most often gradational, although some exceptions can be found (e.g., section 16, near Killi Creek, and section 10 between Kunarak and Papigak creeks). In a few places, the contact is marked by a layer 1-2 cm thick of cherty pebbles.

Discontinuities between successive marine units are also commonly marked by lag gravel and sand. The unconformity at the base of these lag deposits can be abrupt (e.g., as in sections 20 and 65 at ca. 15.5 m a.sl., or Sec. 77 at 6.5 m a.sl.) and accompanied by gentle relief (i.e. <1 m). Lag deposits may also be missing or obscure and unconformities are marked only by abrupt to gradational changes in lithology (e.g. as in Sec. 25 and 26 at 9 and 10 m a.s.l., or Sec. 76 at 5.80 m and Sec. 78 at ca. 7 m a.s.l.). Such unconformities are commonly marked by less than a scattering of cherty pebbles. Elsewhere, deposits along unconformities consist of interbeds of pebbles and clay, sometimes containing considerable amount of organic material. These deposits are thought to represent remnants of the upper shoreface of the transgressive beach.

Lag gravel and sand thicker than a few cm is generally fossiliferous and commonly contains paired valves of Cardita, Hiatella, or Macoma suggesting minimal reworking. Broken and whole single valves, however, are found most commonly.

Based upon their nature and stratigraphic setting, I interpret the lag deposits and associated unconformities exposed in the Gubik Formation as shoreface erosional surfaces, developed during episodes of rising sea level when sediment was removed from the upper shoreface and dispersed into the lower shoreface and offshore areas or moved laterally along the coast by currents (Elliott 1978, pg. 162 and references therein). This process may or may not have been followed by the landward migration of a barrier bar system in different areas.

Presumably as the shoreface and foreshore environments migrated inland with each sea level rise, the underlying facies deposited during the initial stages of the transgression (i.e., lagoonal deposits) were eroded along with portions of older marine sequences deposited during earlier transgressions. Therefore, the base of each transgressive sequence should include a nearly planar erosional surface. It is apparent along some sections of Skull Cliff that during each subsequent transgression this erosive process partially removed older sediments and left only remnants between unconformities.

Rarely are remnants of a former tundra surface, such as that presumably developed during intervening periods of low sea level, found sandwiched between marine units. Such surfaces must have been almost completely eroded during the transgressive phase. Remnants of former tundra surfaces are found (1) at Sec. 98 along the Kokolik River, where middle Pleistocene lagoonal sediments overlie ice-wedge pseudomorphs developed in fluvial gravels (Chapter IV) and (2) offshore in boreholes of the Beaufort Sea where thaw lake sediments enclosing fresh-water ostracodes are found bracketed by marine units (Peggy A. Smith, written comm., Nov. 1982). D.M. Hopkins (written comm., 1983) reports that he found evidence of eolian sand and a paleosol between Karmuk beds and an older marine unit near my section 62.

The presence of erratic cobbles often striated, within the gravel lag facies along many of these unconformities suggests that some transgressive episodes were accompanied by ice-rafting across the Arctic Basin (c.f., Hopkins, 1982b). Although the concentration of these cobbles and boulders along erosional surfaces may be, in part, a product of the erosion of still-older erratic-bearing marine units, their presence on the northern coast of Alaska in marine units older than the Flaxman Formation (Leffingwell, 1919; MacCarthy, 1958) is evidence of late Pliocene and early Pleistocene ice-rafting.

Inner-shelf lithofacies

The most common lithofacies exposed in the bluffs between Skull Cliff and Peard Bay is the inner-shelf facies, a unit composed of fine interbeds of moderately well-sorted medium to fine sand, silty sand, clayey silt and silty clay. This unit commonly overlies the shoreface erosional unconformity and gravel lag facies with a sharp or gradational contact. It is laterally the most continuous sequence in the bluffs. Typical examples of the inner shelf lithofacies are exposed in Sec. 61 and 62 between ca. 6 and 10 m a.s.l. and in Sec. 56 between 7 and 14 m a.s.l.

Beds and laminae range from a few mm to a few cm in thickness. Contacts are abrupt or gradational over a few centimeters. Lines and layers of chert pebbles are common within the sandier layers. Medium to fine sand beds may also display ripple cross-bedding and flazer bedding. The wavey-bedded nature of some interbeds is apparent in many sections, a feature produced perhaps by seabed currents. These units are generally fossiliferous throughout, but mollusk shells tend to be concentrated in the sandier layers.

The inner shelf lithofacies exposed along Skull Cliff are similar to sediments described by Barnes and Reimnitz (1974, p. 451-452) from the inner shelf of the Beaufort Sea where wind-driven waves and longshore currents influence the seabed. This similarity, the interbedded character of the Skull Cliff deposits, and the depth requirement of mollusk fossils suggests that these sediments were deposited in an inner shelf environment at water depths less than 30 m.

The most obvious difference between the inner-shelf lithofacies and modern shelf sediments is that bedding disruptions caused by iceberg scouring are rarely observed in the older sediments. The frequency of scouring is high on the modern shelf (Phillips and others, 1984), hence one might expect to see such evidence in older similar deposits. The apparent lack of ice berg scour may imply perhaps that scouring was less frequent during this period or that such scoured sediments have been eroded. On the other hand, the rapid infilling of such scours with sediments similar to the host sediments may make them difficult to recognize when excavating sections.

The mixed and interbedded character of this lithofacies is a product of the wave and current regime in the inner shelf zone.

Sandier layers represent sediment redistributed and deposited during the open-water season when storms accompanied by a large fetch can generate waves that influence the sea bed. Large waves generate oscillatory bottom currents capable of resuspending sand which then may be moved laterally. In contrast, during periods of loose ice cover or quiet water, finer-grained sediments are deposited and winnowed by the longshore and shelf currents. During the winter, when sediment input from local rivers or eroding coastal bluffs effectively ceases and currents are minimal, only very fine grained sediments are deposited. This process must have resulted in the net accumulation of inner shelf sediments during episodes of higher sea level, quite unlike the modern Chukchi shelf where very little sediment is accumulating today (Phillips and others, 1984).

Inner shelf sediments in Sec. 16, 17, 20, 21, 24, 26, and 65 display discordant folds that are not readily attributable to gouging by grounded ice (Reimnitz and Barnes, 1974) nor to ice-wedge growth (Black, 1974). Notably these folded sediments have been documented as far south as Kilimantavi, 30 km south of Wainwright (Sec. 112, Fig. 2.1b), and as far north as Barrow (Sec. 48, Fig. 2.1a). At Sec. 20 and in Sec. 16 and 65, inner-shelf sediments are isoclinally folded into structures with an original amplitude of over 6m. These features slid by some unknown mechanism over clay-rich sediments and were passively folded, retaining all of their primary sedimentary structures and bedding characteristics. Marine units above and below the folds are notably horizontal and undisturbed. The deformation is quite localized; generally it dampens rapidly and the beds flatten to the horizontal within a distance of less than 400 m. Folds exposed at Sec. 20 appear to be overturned, seaward and toward the north. Folding is observed most commonly in the early Pleistocene Tuapaktushak beds (Aminozone 3) and less commonly in the middle Pleistocene Karmuk beds (Aminozone 2) (beds defined in Chapter III and VII).

The origin of this discordant folding is unknown. One possible explanation is that the high amplitude folds were caused by low-angle sediment slides that occurred while the sediments remained submerged. Even on very low angle slopes, elevated pore-water pressures in one particular layer may be enough to create unstable conditons locally within the sediment pile. Such slides have been interpreted from seismic profiles in the Canadian segment of the Beaufort Sea, but it is unclear how they are actually deformed in cross-section (S. Blasco, GSC, pers. comm., 1982). The possibility that these features were deformed by ice-push can not be completely dismissed, although the nature of the folding appears more compressional than that usually associated with the gouging of ice keels (Reimnitz and Barnes, 1974). The origin of these features of the inner-shelf lithofacies remains an enigma and warrants further study.

Beach and barrier bar lithofacies

Cross-bedded coarse, cobbly gravel and coarse to medium sand occupies portions of the bluffs facing the Chukchi Sea at Barrow, Walakpa Bay, and in the Wainwright area. Similar deposits are found exposed where they intersect river bluffs inland from Icy Cape and Point Lay. Based upon their textural character, primary sedimentary features, and stratigraphic setting, these deposits are interpreted as beach and barrier bar lithofacies. The recognition of these sequences is of critical importance (1) for determining the position of sea level at different times in the past and (2) for locating sources of gravel and sand for construction.

At Barrow, deposits of beach gravel and sand crop out in coastal bluffs extensive for about 2 km southwestward from the village (Sec. 49, Plate II) toward Walakpa Bay (south of Sec. 46, Plate II) where the transgressive sequence is inset into an older inner shelf lithofacies belonging to Aminozone 2 of middle Pleistocene age (bed defined Chapter III) (Fig. 2.6). The sequence consists of low-angle, planar crossbedded to horizontally bedded medium sand, coal-rich sand, and coarse sandy gravel that coarsens upward into the higher-angle planar-crossbedded, coarse sandy to cobbly gravel and pea gravel. This facies is interpreted, by analogy with the modern beach system, as representing some portion of the shoreface or lower foreshore beach complex. Wave-action has concentrated mollusk valves and fragments, mostly Hiatella arctica, Mya sp., and Macoma, throughout, leaving the sequence highly fossiliferous.

A similar sequence of planar-crossbedded beach gravel and sand containing thin laminae of detrital coal crops out up to 8 m a.s.l.for about 2.5 km along the bluffs just south of Walakpa Bay. The deposits at this site, like those at Barrow, are inset into older interbedded inner shelf deposits of Aminozone 3 (defined in Chapter III). Frayed and frozen logs of spruce (Picea sp.), initially delivered to the ancient beach as driftwood, protrude from the perennially frozen gravel at numerous points along the exposure. Abundant fragments and whole valves of Hiatella, Mya, Astarte, and Macoma are also found embedded in the gravel sequence.

Wainwright village is situated on a low ridge partially blocking Wainwright Inlet from the Chuckchi Sea. The ridge is interpreted to be an ancient spit or baymouth bar. Crossbedded coarse sand and gravel, characteristic of the beach and barrier bar lithofacies crops out in the bluffs lining the inner edge of this ridge on the northwestern shore of Wainwright Inlet (Sec. 79, 81, 82, Fig. 4.19). Sections in this bar are rich in fossil mollusks, plant macrofossils and large fragments of frozen driftwood (mostly spruce, Picea). Large foreset beds of sand and gravel dip 23° to 25° to the northeast at the base of the bluff near Sec. 79 and are overlain by planar to horizontally bedded coal-rich sand. Similar foreset sequences are exposed elsewhere in the bluffs. The character of these foresets suggests that a portion of beach

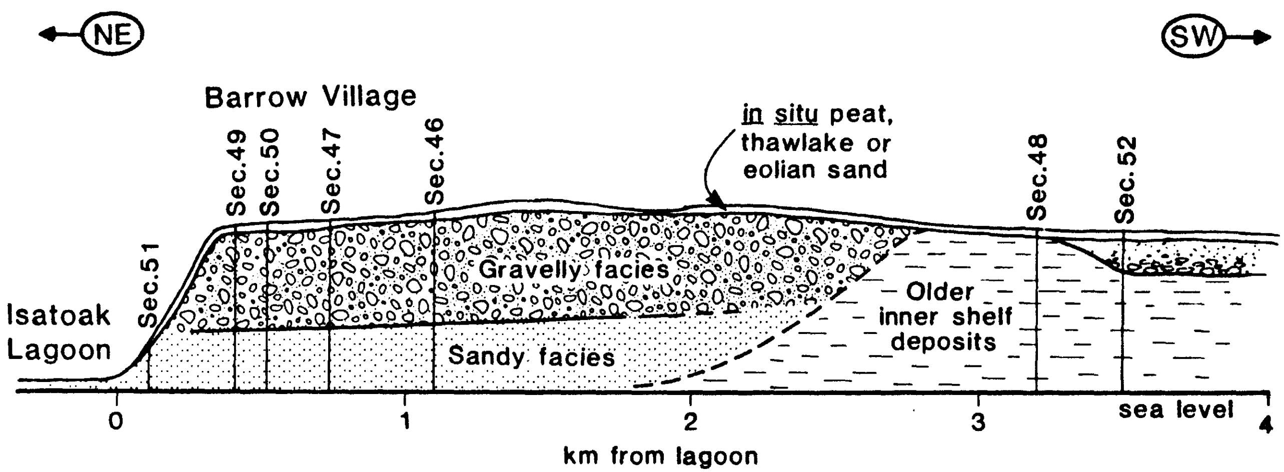


Figure 2.6. Diagrammatic cross-section of the onlap of the beach lithofacies over older fine-grained sediments at Barrow. Section locations are shown on Plate II. Sedimentary structures and bedding not shown.

sequence beneath Wainwright developed as a northeastward prograding baymouth bar or barrier bar complex. During a period of higher sea level this bar complex incompletely partitioned the then swollen Kuk River estuary from the Chukchi Sea.

Lagoonal lithofacies

Lagoon-backed barrier bar systems occupy nearly 50% of the northern coast of Alaska today (Short, 1979). Fine-grained sediments, similar to those in modern lagoons, are preserved within portions of the Gubik Formation and indicate that lagoonal environments lined portions of the coast during former episodes of higher sea level.

Horizontal to small-ripple-bedded deposits of silty sand and silt occupy portions of the bluff between Barrow and Walakpa Bay (Sec. 39, 77, 78) and are also exposed at a few inland sites along the Kokolik (Sec. 105 and 98) and Kukpowruk rivers (Sec. 90 and 94). These sediments are commonly dark brown in color and rich in disseminated peats, plant fragments, and twigs. Moreover, specimens of the brackish water mollusk Macoma balthica were found in several of these deposits. Their presence adds strength to the interpretation that these fine-grained sediments represent former lagoonal environments.

Stratigraphic and amino acid data indicate that the lagoonal lithofacies exposed south of Barrow were deposited contemporaneously with the beach and barrier bar lithofacies exposed near Walakpa Bay and Barrow. The base of the lagoonal deposits is marked by an erosional unconformity (e.g., Sec. 77 at 6.5 m asl) which truncates older, and contorted inner-shelf lithofacies. The lagoonal sediments appear to grade laterally into the beach sequences, although I have not actually been able to physically trace them into the Barrow gravels due to the lack of exposures near Nunavak Bay. The stratigraphic setting resembles transgressive barrier bar systems studied along the United States coast of the Gulf of Mexico (Fig. 2.7). The Barrow complex differs, however, in that the Barrow region was, and still is, a major point of convergence between the northeastward flowing Alaskan Coastal Current and the westward flowing Beaufort Gyre. This convergence results in the development of the modern cuspate spit complex known as Point Barrow. A similar feature developed during an earlier higher sea level event. The shoreline at that time may have resembled the modern spit complex between Barrow village and Point Barrow to the north, but the entire coast was evidently displaced southwestward (Fig 4.20). The lack of older Pleistocene sediments exposed in many sections between Walakpa Bay and Barrow suggests at least partial removal of such deposits by shoreface erosion during this transgression, although remnants probably exist in the subsurface. Progradation was initiated during the culmination of the sea level rise to about 10 m a.s.l. at which time the extensive gravel complex at Walakpa Bay was deposited, eventually forming an eastward trending cuspate spit complex in the Barrow area (see discussion in Chapter IV).

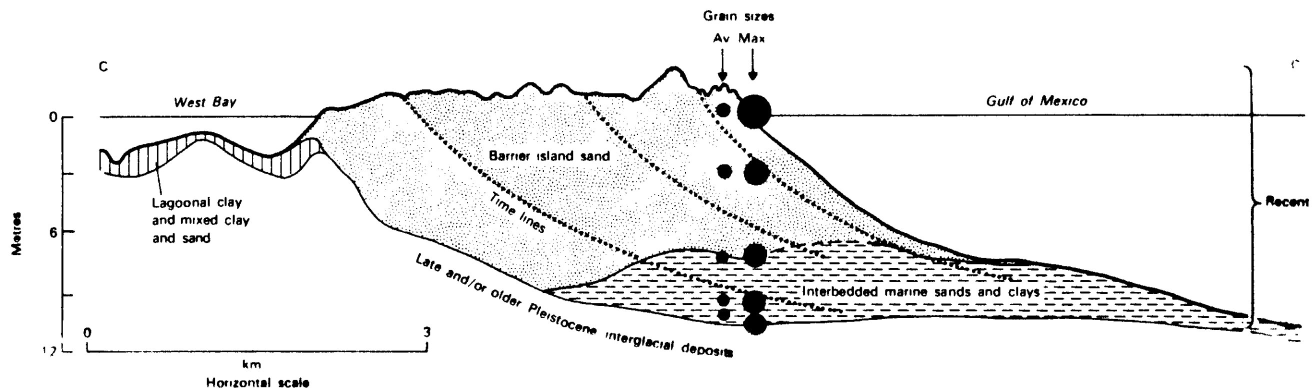


Figure 2.7. Stratigraphic section through coast of Galveston Island Texas (Bernard and others, 1962. The stratigraphic setting of lagoonal and beach deposits at Barrow is similar to the setting depicted above.

Tidal Delta (?) lithofacies

Planar-crossbedded to trough-crossbedded fine pebbles, coarse to medium sand, and detrital coal characterize the sediment exposed in the upper half of the bluffs from section 6 to section 10 half way between Kunarak and Papigak creeks. Viewed in section from the modern beach, these sediments appear as very low angle to sub-horizontal prograding foreset beds dipping about 5.5° toward the north; individual bedding planes vary from 1° to 7°N. In cross-section, normal to the modern beach where sediments are exposed along erosional gullies, the bedding is more complex in character indicating bidirectional flow nearly perpendicular to the modern coast (Fig. 2.8). In general the sequence is more coal-rich at the base becoming sandier upwards. Much of the sandier beds are highly oxidized. Mollusks, including Natica and other gastropods, are common throughout the sequence and paired valves of Hiatella arctica can be observed in growth position. Where the mollusks are interbedded with a great deal of coal, the fossils are soft and punky as if partially decomposed by the more acidic environment generated by the coal.

The origin of this deposit is not entirely clear and similar sediments have not been found exposed elsewhere on the coastal plain. The rather large scale bidirectional cross-bedding and high energy character of the sediments has suggested to earlier workers (D.M. Hopkins, pers. comm., 1981) that this may represent an ancient tidal-delta formed at the outlet between barrier islands fronting a lagoon when sea level was higher than at present. Amino acid data on Hiatella arctica, Macoma, and Siliqua patula from these sediments fall into aminozone II (Chapter III) suggesting that deposition took place during the middle Pleistocene (AAL-2833, 2839, and 2840) when sea level was approximately 23 m a.s.l. The abundance of detrital coal in these sediments may have been derived from an ancient fluvial system draining into the lagoon much like the nearby Kogru River where the modern beaches are black with detrital Cretaceous coal.

Eolian Sand lithofacies

Medium to fine-grained sand of eolian origin blankets large portions of the Arctic Coastal Plain west of the Colville River,in places forming extensive dune fields. These deposits were initially described by Black (1951) and O'Sullivan (1961) and recently have been the focus of studies by Carter and Robinson (1978) Carter and Galloway 1981, Carter (1981a, 1981b), Galloway (1982), and Carter (1982, 1983a, 1983b). The longitudinal dunes of the sand sea between the Colville and the Meade rivers began forming either between 36 ka and 50 ka B.P. or beyond the range of radiocarbon and remained active until about 12 ka B.P. (Carter 1983a, 1983b). An extensive blanket of loess (also known as the "Upland Silt" or the "Foothills silt") was also deposited during this period as a distal eolian facies along the margin of the Arctic Foothils southwest of the dunes (Carter, 1983b). A thin

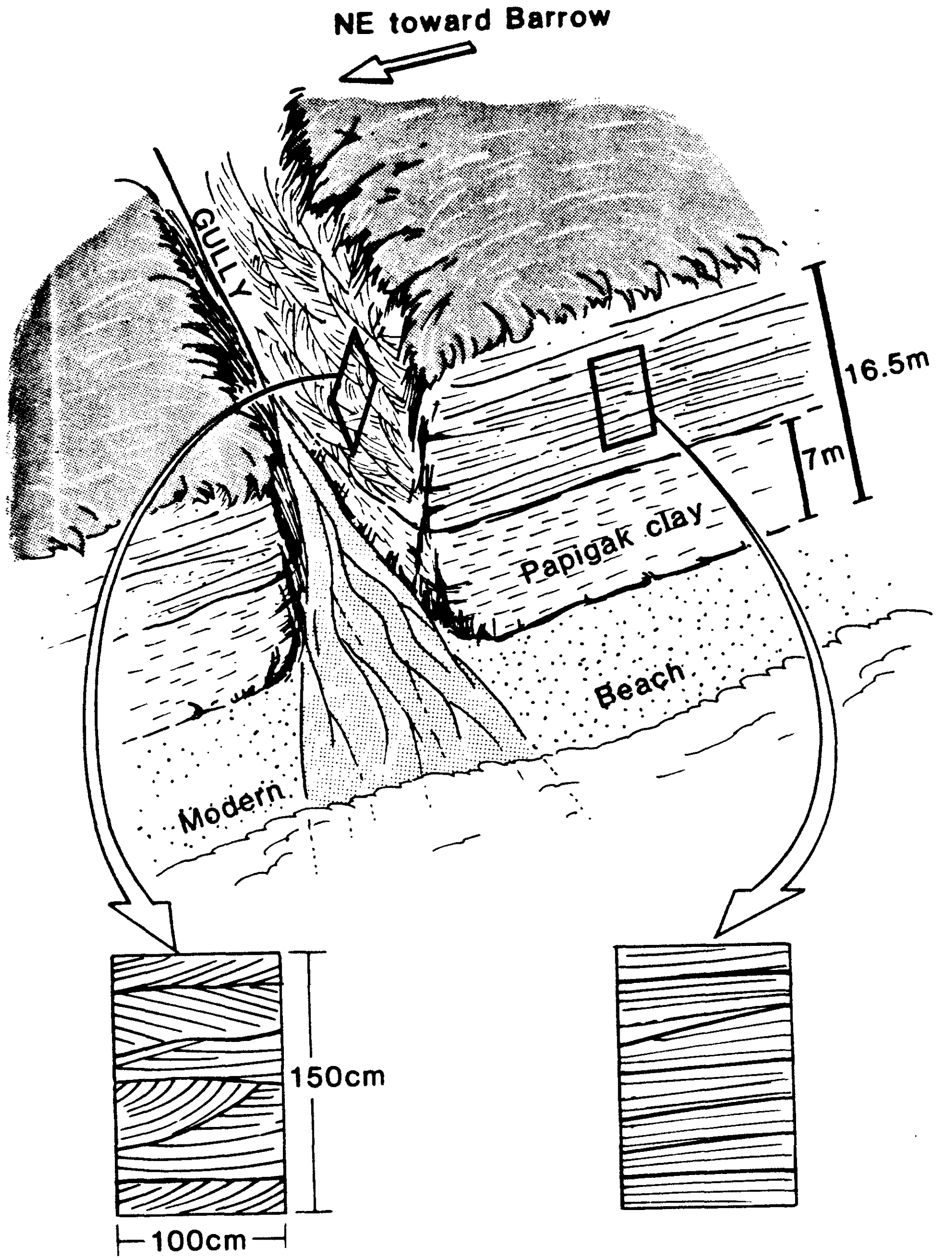


Figure 2.8. Schematic diagram of the tidal-delta (?) sediments exposed at Sec. 10. Stratified bidirectional cross-bedding in the gully exposure suggests paleocurrents moving nearly perpendicular to the modern coast. Most crossbed sets dip landward toward the E-SE probably caused by onshore summer storm waves. Sea level at the time of deposition may have been about 23 masl indicating a paleo-waterdepth of less than 6.5 m.

cover of younger eolian sand subsequently obscured the older Pleistocene dunes between 8 and 11 ka B.P., and smaller parabolic and longitudinal dunes, now stabilized, formed at sometime after 3500 years B.P. (L.D. Carter, quoted in Williams, 1983b).

A thin discontinuous mantle of eolian sand covers portions of the western coastal plain (Williams, 1983a, 1983b). It is distingushed from other facies by being composed of well-sorted medium sand with a massive to weakly stratified character. It lacks any marine macrofossils but on occassion contains reworked, windblown, ostracode valves. Good examples of eolian sand are exposed in the bluffs at Sec. 12, 16, 25, 26, and 38 along Skull Cliff (Plates I and II) and in Sec. 79, 80, 81 and 82 near Wainwright (Fig. 4.17 and 4.19). Although extensive, well-developed dune complexes are not found, localized dune fields are present along major river courses and on the upland southwest of Karmuk Point. Radiocarbon dates on buried peats in dunes along the Meade River indicate eolian activity in that area more recently than 11,500 years B.P. (Williams, 1983a).

Eolian deposits exposed in bluff sections between Barrow and Peard Bay form a thin discontinuous mantle blanketing the older marine units. A portion of this sand cover, mapped by Black (1964) as part of the Barrow unit, is actually composed of thaw-lake sediments developed after about 11 ka B.P. As the lakes grew, they trapped eolian sand. Complicating the recognition of such deposits is the fact that many thaw lakes developed in ice-rich marine sand--sand that also acts as a local source for eolian transport. Hence, lakes formed in marine sand later acted as catchments for eolian sand, the latter of which is then mixed with the host material as the lake expands.

The materials of eolian origin vary in grain size from medium (2 to 1ϕ) to fine (3 to 2ϕ) sand and are typically frosted and subangular to subrounded in shape. Where eolian processes have ready access to marine sands, however, the eolian deposits often include a greater portion of polished, more rounded grains. The sand is primarily composed of clear quartz. Black chert, feldspar, and heavy minerals are minor constituents. Stratigraphically, the eolian sands, when present occupy only the highest portion of the bluff or form stabilized dunes in many stream valleys facing the Chukchi Sea. Rarely are eolian sediments found interbedded between marine units.

<u>Thaw-lake Facies</u>

Thaw-lake sediments, like the eolian facies above, form a discontinuous cover overlying the older marine units in many sections along Skull Cliff. In aerial photographs of the coast the outline of old, drained thaw lakes can be seen where such sediments are exposed in the bluffs. Due to the loss of ice in the surface sediments during thermokarst aggradation, the tundra surface over the drained lake is commonly a few meters lower than the surrounding terrain (cf. Sellman and others, 1975). All radiocarbon-dated thaw-lake sequences between Barrow and Peard Bay are less than 12 ka old.

The thaw lake sequences are generally similar though they vary in detail. The base is commonly undulatory with a relief of less than 0.5 m. A layer of disseminated organics or peat forms the base. As the lake expands, host sediment and vegetation topple from the banks into the lake, and eolian sand maybe added through time. The basal layer becomes covered with material carried first by traction and then in suspension. The actual lake bottom gradually comes to consist of mud or fine sand with plant debris continuing to accumulate at the lake margins. In most cases, this process results in the gradual accumulation of medium-fine sand and sandy silt interbedded with stringers or clots of detrital peat. In modern thaw lakes with sandy bottoms small symmetrical wave ripples with continuous and some bifurcating crestlines can be observed. Such features can sometimes be seen in section. Fresh water ostracods are ubiquitous in these sediments and distributed to new lakes on the legs of birds or in their feces. If the thaw-lake is developed in marine sediments, both fresh water and reworked marine forms will be preserved. Ostracodes provide a useful means of distinguishing among fresh-water, marine, and eolian lithofacies. Because the ostracod valve is somewhat aerodynamic ("frisbee-like"), fossil valves can be reworked into eolian sediments; but reworked valves are noticeably worn and abraded.

Superposed thaw-lake facies of different ages have been observed at Sec. 64 and 66 (Plate I). Where observed, the lower, older lake deposit is intruded by ice-wedge pseudomorphs, indicating that ice wedges formed shortly after the first lake had drained. Subsequently this surface was somehow flooded and a second thaw lake later formed on the site of the older lake. The younger thaw lake sediments in Sec. 64 cited are dated at 10,430 ± 150 yrs B.P. (I-12923) and those in Sec. 66 are 8950 ± 150 yrs B.P. old (I-12951).

To date, thaw lake sediments have not been found beneath any marine beds along Skull Cliff. In shallow boreholes in the Beaufort Sea, however, P.A. Smith (USGS, pers. comm., 1983) has reportedly found two thaw lake sequences buried by marine sediments.

Fluvial lithofacies

Fluvial gravel, sand, and fine-grained overbank sediments comprise a significant portion of the Gubik Formation, especially east of the Colville River. Fluvial processes have reworked and removed portions of the marine record across the western edge of the coastal plain particularly between the Utukok and Kukpowruk Rivers leaving only broad meandering stream valleys entrenched into the Cretaceous bedrock. Marine deposits are, however, patchily preserved in the subsurface along interfluves. The complex fluvial history of these rivers south of Wainwright has yet to be studied in adequate detail. Deposits exposed beneath terraces in many of the meander scars contain reworked valves of marine mollusks indicating the presence of older marine units upriver.

Gravel and sand deposits of fluvial origin are distinguished from the beach and barrier bar lithofacies by the nature of the bedding and the stratigraphic setting. Well-rounded gravel and cobbles can occur in both marine and fluvial environments as river materials are reworked during transgressions and marine units are fluvially eroded, hence particle shape is not always a useful criteria.

Observations Concerning Shallow-Marine and Nearshore Sedimentation Across the Western Coastal Plain

The stratigraphy of the various lithofacies and associated discontinuities observed in the Gubik Formation conform closely with depositional models described for both Holocene and ancient asymmetric transgressive and regressive sequences (Reading 1978; Ryer 1977, and references therein). Particularly in the bluffs between Barrow and Peard Bay, the marine sediments associated with each high sea stand are bounded below by a disconformity at the base of a coarse layer and above by the disconformity associated with the next younger transgression. Because of the effects of shoreface erosion, little depositional record of the beachface remains (Bruum, 1962; Swift, 1968). Progradation occurs only when the rate of sea level rise dramatically decreases and a stable shoreline profile can be maintained. "Transgressions are recognized by the presence of erosional disconformities that separate progradational sequences" (Ryer, 1977, pg. 186). Similarly, the stratigraphic record of marine transgression along Skull Cliff consists of inner-shelf marine lithofacies disconformably overlying older inner- or central-shelf lithofacies. Stratigraphic examples of transgressive shoreface erosion (Fig. 2.9), particularly along barrier island coasts adjacent to gently sloping coastal plains, are most commonly cited from the eastern coast of the United States (Fischer, 1961; Dillon, 1970; Kraft, 1971; Swift, 1975) and from the coastal areas of the Gulf of Mexico (Bernard, LeBlanc, and Major, 1962). Based upon my observations of the Gubik stratigraphy, and discussions of similar sequences in the literature, particularly Ryer (1977), a model for transgressive marine sedimentation on the coastal plain can be summarized as follows:

Coincident with hemispheric deglaciation, a relatively rapid eustatic rise of sea level results in the southward enchroachment of the sea across the Chukchi Shelf. Because of the flat topography of the shelf, the horizontal migration of the shoreline during this period is probably dramatic. Rivers draining the coastal plain begin alluviating in their lower reaches as they attempt to adjust to the rapidly rising base level. Most of this material, along with migrating beachface deposits, is subsequently eroded and removed as the sea transgresses across this bedrock plain. Similarly unconsolidated marine or eolian sediments

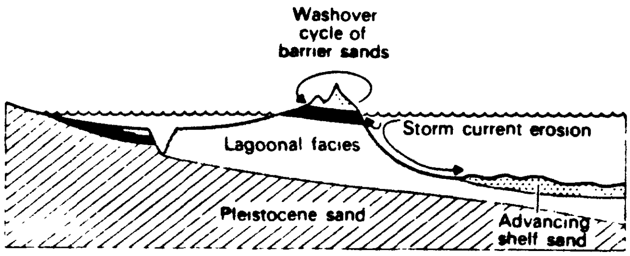


Figure 2.9. Mechanism of landward barrier migration by shoreface erosion (after Swift 1975, from Elliott, 1978).

stranded during and after the previous regression and subaerial exposure are partially eroded and removed to offshore environments. This shoreface erosion results in the formation of a transgressive disconformity, upon which lag deposits of sand or coarse gravel or portions of the transgressive beach are deposited. The presence of striated, erratic cobbles and boulders upon some of these surfaces or incorporated in shelf sediments suggests that certain transgressive events were times of glacial ice-rafting in the Arctic Basin. The preservation of older marine units beneath a disconformity seems to be fortuitous, dependent upon the configuration of the coast and the rate of sea level rise.

During the waning stages of the transgression as the rate of sea level rise diminished, progradation begins. Sediments are deposited on the disconformity as soon as water deepens enough so that the bottom is below wavebase. Sediment is continually supplied to the offshore areas by river effluent and by the erosion of unconsolidated materials from the coastal bluffs. Subsequently, beach and barrier bar environments prograde and mature, probably resembling their modern counterparts.

Coincident with the following glacial episode, or as a consequence of gentle uplift, regression commences. Regressive sediments have not been clearly recognized in the Gubik Formation at Skull Cliff, therefore, it is difficult to reconstruct how this process proceeded. Presumably as regression continued and the shoreline retreated seaward, a thin offlap sequence of lithofacies was deposited across the Chukchi Shelf. Laterally migrating river systems may have assisted with the removal of some of these sediments particularly south of the Utukok River. Gradually, a polar desert environment dominated by extensive eolian activity and sparse vegetative cover evolved across the arctic coastal plain and exposed continental shelves (cf., Carter, 1982). During the subsequent marine transgression, however, a portion of the depositional record of this period, along with much of the regressional record, were eroded. The preservation of paleosols, aeolian sediments, and some older ice-wedge pseudomorphs, however, suggests that the paucity of Pleistocene sediment is also a function of limited primary deposition.

The repeated inundation of the western coastal plain throughout the late Cenozoic has left a thin, but extensive sediment record from which one may reconstruct the sea level history. Unlike the sloping Beaufort Sea shelf where continual subsidence(?) has preserved a more continuous marine record (Dinter, 1983), the sea level record of the western coastal plain is limited to only high sea level events. No doubt, the sea level history recorded on the western Arctic Coastal Plain is fragmentary and incomplete, particularly for those events that never rose above present sea level.

Conclusion

Evaluation of the environmental significance of different sediment types observed in bluffs between Barrow and Peard Bay resulted in the recognition of several lithofacies that can be explained in terms of modern depositional regimes. These lithofacies include; (1) the gravel/cobble lag facies deposited following transgressive shoreface erosion; (2) the inner-shelf facies deposited in water depths less than 30 m; (3) the beach and barrier bar lithofacies deposited in the nearshore environment; (4) the lagoonal facies deposited behind barrier island chains; (5) tidal-delta sequences deposited between barrier islands; (6) the the eolian sand facies deposited as a thin discontinuous blanket across the coastal plain; (7) the thaw-lake facies deposited beneath short-lived thermokarst lakes; and (8) the fluvial facies deposited by major rivers draining the western coastal plain. The distribution and preservation of these lithofacies allows the reconstruction of nearshore and shallow marine sedimentation on the North Slope. The marine stratigraphy of the Gubik Formation is not a symmetrical transgressive/regressive stratigraphy. Rather, its stratigraphy is highly asymmetric, recording primarily the shoreface erosion and progradational phases of the ideal transgressive/regressive cycle (symmetric model of Kauffman, 1969).

CHAPTER III
AMINOSTRATIGRAPHY

Introduction

Since the advent of amino acid geochronology in the late 60's and early 70's, the technique has been successfully applied to a broad variety of Quaternary stratigraphic problems (see Hare and others, 1980; Wehmiller, 1982a, for reviews). Although knowledge concerning the kinetics of amino acid diagenesis and protein hydrolysis in fossil materials is incomplete; the method is of fundamental importance to Quaternary researchers as it provides an objective dating tool useful beyond the range of the radiocarbon method with the potential for differentiating materials more than two million years old in some areas and perhaps over 5 million years old in the Arctic.

The extent of amino acid diagenesis in the indigenous protein of fossil mollusks provided the most useful means of differentiating and correlating the marine strata of the Gubik Formation across the western Arctic Coastal Plain. The aminostratigraphic and geochronologic achievements of the method in this study are intended to supplement existing knowledge (Miller and others, 1977; Nelson, 1982; Brigham 1983a; Miller, 1982) concerning the applications and limitations of amino acid geochemistry as a geochronological tool in Arctic areas.

Fundamental Background

Amino acids are the structural components of protein in all living cells. Although hundreds of amino acids exist, only twenty are common in the proteins of living organisms (Kvenvolden, 1975). In mollusk shells, protein exists as a structural matrix secreted into the extrapallial fluid between the mantle and the shell upon which calcium carbonate is deposited during shell growth (Hare, 1967; Kennedy and others, 1969).

Amino acids may occur in one of two optically-active forms, or enantiomers which are mirror images of each other (Fig. 3.1). Amino acids which rotate plane-polarized light to the right (dextrorotary) are D-amino acids, whereas those which rotate polarized light to the left (levorotatory) are L-amino acids. Essentially all protein amino acids in living organisms are of the L-stereoislomer configuration and it is this fundamental observation that allows amino-acid diagenesis to be used as a geochronological tool (Hare and Abelson, 1968). Following death of the organism, the L-amino acids slowly and reversibly invert to the D-stereoisomer until a thermodynamically-stable equilibrium mixture is attained. For an amino acid with a single chiral carbon, such as aspartic acid or leucine, this process is called racemization (Fig. 3.1). Racemization proceeds until the proportion of the two isomers, expressed as a ratio of D/L, increases to an equilibrium value of 1.00. For amino acids with two chiral carbons, such as

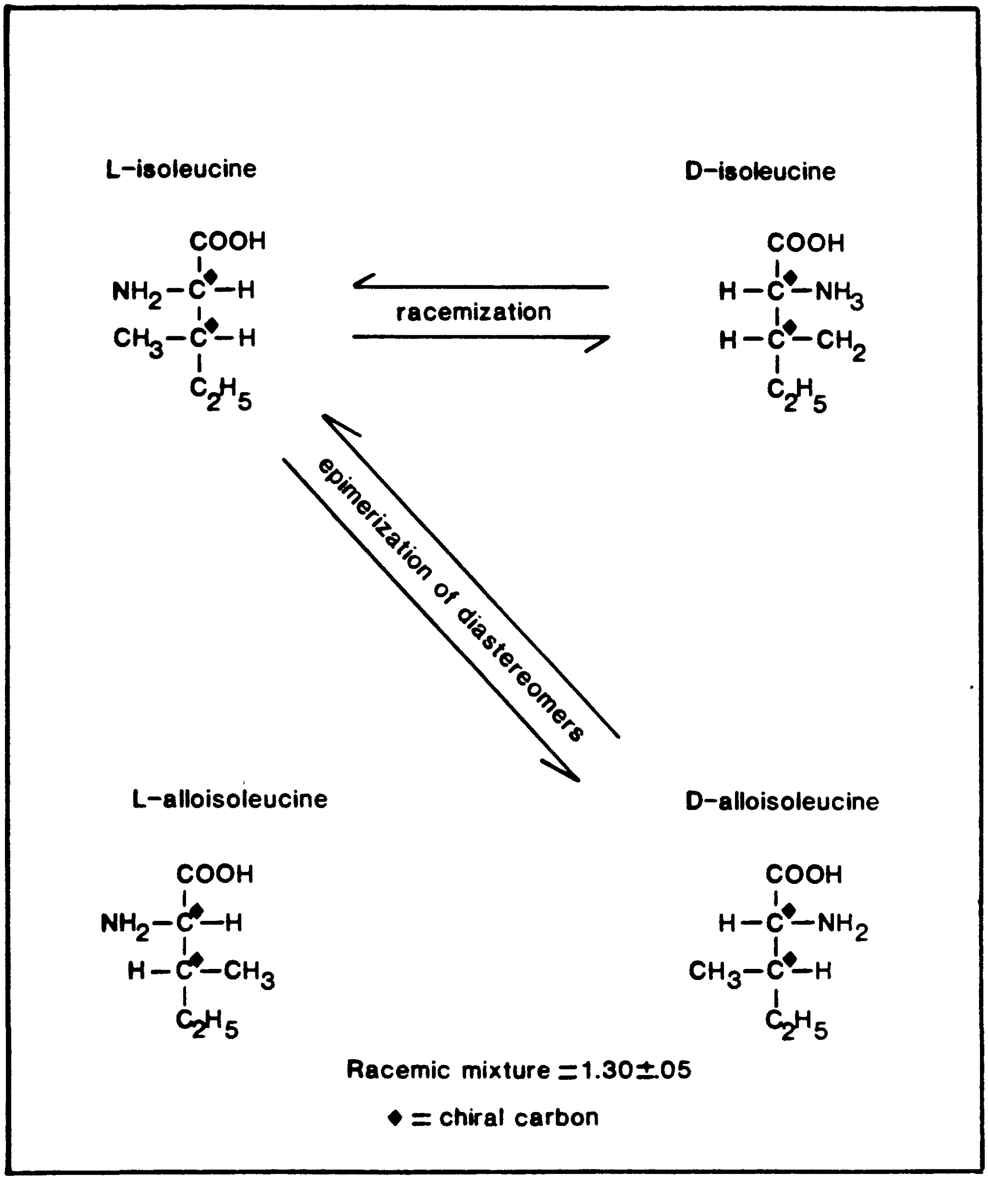


Fig. 3.1. Stereoisomers of isoleucine illustrating the epimerization of L-isoleucine into D-alloisoleucine.

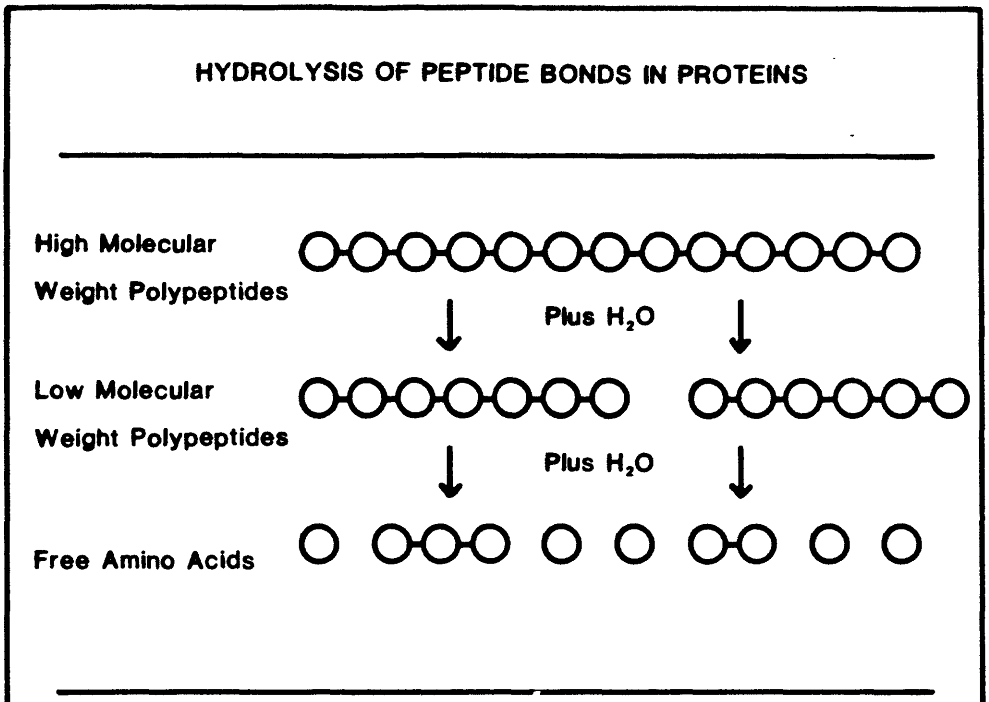
isoleucine and theronine, the interconversion is from the L-configuration to the D-configuration of its diastereomer. Consequently, L-isoleucine (hereafter, Ile) reversibly converts to D-alloisoleucine (hereafter, aIle) until the proportion of D/L increases to 1.30 ± 0.05 (Miller and others, 1979). This process of interconversion about the α-carbon is called epimerization (Fig. 3.1).

The extent of racemization (or epimerization) is primarily dependent on temperature, time, and taxon. Because of the natural integrity and buffering character of the carbonate matrix, leaching and pH have little direct influence on shell diagenesis (Hare and Mitterer, 1969; Miller and others, 1979; Miller and Hopkins, 1980). Temperature, however, is the most critical rate-controlling factor, approximately doubling the rate of epimerization for every 4°C rise in temperature.

Each amino acid racemizes at a slightly different rate and is also influenced by its state in the decaying protein matrix. Kriausakul and Mitterer (1978, 1983) have illustrated that racemization proceeds relatively slowly in amino acids internally bound within a polypeptide chain and in structurally free amino acids. In contrast, terminally-bound amino acids (i.e. those at the ends of peptide chains) invert at a much faster rate. Moreover, those bound at the NH_2-terminus racemize faster than those bound at the COOH-terminus. This rate, especially in NH_2-terminal amino acids, is apparently influenced by the formation of diketopiperazines (Mitterer, 1982; Bada and others, 1982).

The state of each amino acid, as above, is also dependent upon the continued protein deconvolution and hydrolysis of peptide bonds (i.e., the addition of water) between adjacent amino acids. Moreover, bond strength varies between different amino acids. Hence, the rate of hydrolysis influences the apparent or overall rate of racemization. Generally, the extent of hydrolysis, determined by the proportion of free to bound amino acids, increases with increasing fossil age (Miller and Hare, 1980; Brigham, 1983) (Fig. 3.2). Exactly how the rate of hydrolysis changes with time, and precisely how it effects rates of racemization is not known. Hare (1980) has shown that hydrolysis in bone collagen proceeds through two phases such that epimerization begins only after denaturation has reduced the complex protein molecules to lower molecular weight forms. The activation energy for at least the initial denaturation step, may be twice as high as for racemization. Whether or not this initial slow diagenesis proceeds in a similar manner in the less complex matrix protein of shells at low temperatures (i.e., <-5°C) is not known. It is possible that at cold sites insufficient energy may be available to complete denaturation of bone collagen in 10^5 to 10^6 years (P.E. Hare, unpubl.), a fact that would, in turn, severely retard the rate of epimerization.

Because the rates of racemization (epimerization) and hydrolysis are most sensitive to temperature, the accuracy of



$$\left(1-\left[\frac{[\text{Total}]-[\text{Free}]}{[\text{Total}]}\right]\right)\times 100=\%\text{Free}$$

Figure 3.2. Hydrolysis of peptide bonds between amino acids produces an increasing proportion of free amino acids. Because amino acids epimerize at different rates (i.e., free, internally peptide-bound, or terminally peptide-bound), the rate of hydrolysis influences the rate of epimerization.

absolute age determinations from amino acid analyses is a function of the accuracy of the thermal history estimated for a particular sample (see Chapter VI). Such an estimate is difficult to determine, especially for the Quaternary when air and oceanic temperatures fluctuated rapidly. It is possible to minimize the effect of the thermal parameter by analyzing samples collected from a limited geographical area across which one can reasonably assume that the paleotemperature fluctuations have uniformly influenced all deposits of similar age. When evaluated in this context, amino acid ratios provide a valuable index of relative age, particularly where stratigraphic relationships are ambiguous (Miller and others, 1977).

In this study, amino acid ratios have been used to determine the relative age of superposed strata and to correlate such strata across the northwestern Arctic Coastal Plain. This application, termed aminostratigraphy (Nelson, 1978; Miller and Hare, 1980), provided the most useful means for differentiating stratigraphic units in individual exposures of the Gubik Formation and for determining their relative ages. Moreover, it provided a chemical means for evaluating the correlation of these units with other marine sequences in adjacent arctic areas. Age-calibration of the ratios is based upon radiocarbon dates, paleontological age estimates, paleomagnetic inclination data, U-series dates, a single U-trend date, and kinetic model age estimates (Chapter VII).

Analytical Methods

Many geologists and geochemists using amino acid stereochemistry for relative or absolute age dating rely upon the extent of isoleucine epimerization in the total acid hydrolysate (free plus peptide-bound amino acid assemblage: hereafter, Total). Because this reaction proceeds slowly in arctic areas due to the low temperatures and limited hydrolysis, early diagenesis is more reliably measured by the extent of isoleucine epimerization in the free, or naturally hyrolyzed, (hereafter, Free) amino acid fraction (Miller and others, 1977). In this study, the extent of isoleucine epimerization, expressed as a ratio of D-alloisoleucine/L-isoleucine (hereafter, aIle/Ile) in both the Free and Total fractions, was used to distinguish marine units of different age and to identify disconformities between conformable units of similar lithology. Additional parameters, including the percentage of free amino acids and the concentration of free leucine were examined to assess the increasing extent of natural protein hydrolysis in the samples. Leucine, in particular, is of interest because it is a stable amino acid retained by the shell matrix over long periods of geologic time. Amino acid determinations were made using a cation-exchange high performance liquid chromatographic (HPLC) analyzer automated by Dr. Gifford Miller (University of Colorado, unpublished) and similar to that described by Hare (1975) utilizing O-Pthalaldehyde fluoresence detection (Benson and Hare, 1975).

Molluscan Taxa Analyzed

Amino acid analyses undertaken for the purpose of establishing a regional stratigraphic framework were restricted to specimens of Hiatella arctica Linné, Mya sp. (including M. truncata Linné and M. arenaria Linné, Macoma sp. (including M. balthica, M. brota, M. lama and M. obliqua), and Astarte sp. (primarily A. borealis). Representatives of at least one and often two of these genera were present in nearly all fossiliferous marine deposits and produced consistent, comparable results. By analyzing several genera from selected units, other strata enclosing only one of the above types could be correlated. Analyzing several genera also provided a useful means of cross-checking the correlations and detecting reworked populations.

X-ray diffraction analyses were carried out on powders of selected shells from several stratigraphic units (including aminozones 1, 2, and 3) to determine whether recrystallization of aragonite to calcite had occurred. Such recrystallization is thought to alter the preservation of shell protein. Modern samples of Hiatella arctica were also analyzed for comparison. Although several samples of Hiatella arctica and Mya truncata were suspiciously dark gray in color, all of the samples were found to be unaltered and composed entirely of aragonite (Steve Forman, pers. comm., February, 1983).

Because the relative proportion of each amino acid variesin different genera, the rate of epimerization is different for each genera. Consequently, the absolute value of the aIle/Ile ratios derived from one genus are not directly comparible to ratios derived from another genus of the same age. Isoleucine in Hiatella and Mya epimerizes at a similar rate. In general aIle/Ile ratios in Mya sp. are <5 percent lower than in Hiatella sp. from the same unit. Astarte epimerize slightly faster than Hiatella and produce ratios only slightly higher in similar strata. Macoma, by comparison, racemize much faster than Hiatella and produce ratios in the Total fraction nearly twice those observed in Hiatella. In rare cases when sample identification is difficult, due to either small sample size or abrasion of the shell fragment, these four genera can be distinguished from one another by diagnostic amino acid patterns (Andrews and others, 1982; submitted).

Wehmiller (1980) hypothesized that the differences observed in the relative rate of racemization (epimerization) between genera is caused by the greater stability of aspartic acid peptide bonds in carbonate matrices. That is, genera with a greater proportion of aspartic acid (expressed as a ratio of aspartic acid/leucine) containing a greater proportion of "hydrolysis-resistent" peptide bonds are expected to racemize more slowly than genera with less aspartic acid (i.e., the rate of racemization is inversely proportional to the amount of aspartic acid in the sample). Wehmiller found that faster racemizing genera have an aspartic acid/leucine ratio of less than 3.0 whereas slower racemizing genera have a ratio between 5.0 and 7.0. This relationship generally holds for the four genera studied here, however, the relationship is not straight-forward (Table 3.1). Among those genera analyzed, Wehmiller's hypothesis would predict that the relative rates of epimerization would be as follows:

Fastest ——————————————————>Slowest
Macoma > Astarte ≥ Mya > Hiatella
—— increasing amount of aspartic acid ->

Macoma is by far the fastest epimerizer and contains the least amount of aspartic acid. In theory, Mya should epimerize faster than Hiatella, however, in practice this is certainly not true. In most cases, Mya ratios are lower than Hiatella and are rarely higher. The relationship between rates of epimerization and the concentration of aspartic acid is evidently more complex than suggested by Wehmiller (1980).

Analytical and geological variation

Ideally, all valves of a single genus of the same age, with the same thermal history should yield the same ratio of alle/Ile. Alle/Ile ratios in monospecific specimens of the same age do differ however, and the difference is derived from both geological and analytical sources.

Geological sources of variation include subtle differences in the ground temperature history; differences in burial history; and differences inherent in the shell. Brigham (1983b) has shown, for example, that alle/Ile ratios and the absolute concentration of amino acids in Hiatella can vary slightly from one anatomical region of a valve to another. Most importantly this paper demonstrates that the scatter of alle/Ile ratios in a monospecific sample from a single stratum can be reduced by consistently analyzing a single region of the shell, preferably the hinge. In addition, Brigham (1983b) showed that an extreme scatter of alle/Ile values can be obtained from a single deposit, particularly if samples for analysis are taken from the anterior or posterior growth edge. Layered mollusks, such as Protothaca, Saxidomus, and Glycymeris are sampled from a single, structural layer to achieve consistent results (Wehmiller and others, 1978). To minimize the amount of variation in the alle/Ile ratio between stratigraphic units of different age, mollusk valves for amino acid analysis were sampled almost exclusively from the hinge region. In rare cases, when only shell fragments could be found, growth edges were used.

I investigated the alle/Ile ratios from different layers along different portions of Cardita crebricostata and found a great deal of inherent variation (Table 3.2). After analyzing over 100 valves of C. crebicostata sampling through the entire shell near the hinge and noting their inconsistency, I discovered that alle/Ile ratios within the middle layer alone varied over 50% from the beak to the outer growth edge. This small experiment needs to be confirmed by additional tests, but, it is my opinion that

Table 3.1. Relative abundance of Aspartic Acid (expressed as a ratio of aspartic acid/leucine, ASP/LEU) in fossil Mollusks from the Alaskan Arctic Coastal Plain

Genus	ASP/LEU in Total Mean	Range	No. of Samples
"Fast Racemizers"			
Macoma	2.95 ± 0.60	2.2 - 3.9	10
Astarte	5.87 ± 0.70	4.55 - 6.8	8
"Slow Racemizers"			
Mya	7.50 ± 1.36	5.22 - 9.4	11
Hiatella	10.27 ± 1.00	9.2 - 12.4	9
Selected LC[1] data for modern samples from Wehmiller (1980, pg. 348)			
Macoma	2.22		
Protothaca	6.0		
Saxidomus	5.2		
Tivela	7.3		

[1]LC = liquid-chromatography

Table 3 2. AIle/Ile ratios from different layers of Cardita (Cyclocardia) crebricostata and portions of Neptunea sp.

Part	aIle/Ile Free	Total
	CARDITA (AAL-2094)	
A (whole thickness of shell at hinge)	0.677 ± .012	0.088 ± .013
B (hingeward inner layer) 3	0.548 ± .039	0.046 ± 0.00
C (edgeward inner layer)	0.701 ± .026	0.104 ± .005
	NEPTUNEA (AAL-2808)	
D (top of columella)	0.238	0.044
E (bottom of columella)	0.294	0.026
F (top of columella)	0.120	0.046
G (bottom of columella)	0.216	0.036

Cardita are not useful for aminostratigraphy. Results derived from analyzing different portions of the columella of a single Neptunea suggest this may also be true of that genus.

Analytical variation is largely caused by changes in columnn performance (usually less than 10%) and by changes in the proportion of effluent to O-Phthalaldehyde (OPA). Because each amino acid couples with the B-mercaptoethanal and OPA at different rates, changes in the properties of column flow to OPA can alter the fluorescent yield, i.e., the apparent concentration of amino acids detected (Cronin and others, 1979). Throughout the course of this study, a ratio of OPA flow to column flow of 1.1 was maintained and commercial amino acid standards were analyzed every other day on both of the analyzers to monitor the reproducibility of results. The mixture analyzed consists of Hamilton Co. Type H amino acid calibration standard dilluted to a concentration of 250 p moles/10 ul to which is added D-alloisoleucine (Sigma Chemical Co. No. I-0380) and L-Norleucine (Sigma Co. No. N-6877). Over a 5 1/2 month period in 1982, aIle/Ile standards from both analyzers varied by only 4 to 5% (Andrews and others, 1982). In addition, powdered samples of Saxidomus distributed by J. F. Wehmiller were analyzed to establish an interlaboratory data set for calibration purposes. G.H. Miller's unpublished results of analyses of these samples are listed in Table 3.3 for reference by other laboratories. In general, the INSTAAR laboratory averaged 1% to 5% higher than the mean results of all laboratories using ion-exchange HPLC separation for both the liquid and powdered interlaboratory test samples (except that for sample ILC-A powder only, the INSTAAR lab averaged 13% lower) recently compiled by Wehmiller (1982b). The more recent results (Table 3.3) for ILC-A are close to the mean.

Sample preparation

The method of preparing mollusk samples for analysis was reevaluated and changed during the period of September, 1980, through September, 1983, when samples for this study were analyzed. Prior to May, 1982, samples were prepared as outlined in Appendix A.2 under what is referred to as "1981 prep." Through a series of experiments outlined by Miller and others (1982) it was determined that the post-dissolution, pre-hydrolysis transfer of aliquots to separate vials for the Free and Total hydrolysate caused fractionation of the amino acid population. During this process, the high molecular weight polypeptides containing the least racemized Ile fraction remained adhering to the glasswalls of the original test tube, and were lost from the sample. As a result, the 81 prep produced relatively high aIle/Ile values.

After May, 1982, when the fractionation problem was discovered, a new preparation method was adopted, referred to as the "76b prep." This preparation is called the 76b prep because it is similar to the preparation initially used by G.H. Miller during a portion of his first year of operation in the laboratory. This preparation (Appendix A.3) eliminated the fractionation problem by

Table 3.3. AIle/Ile ratios from Interlaboratory samples supplied by J.F. Wehmiller (1982b). Analyses completed using HPLC by G.H. Miller (Oct. 1983).

Sample ID	INSTAAR ID	Total aIle/Ile	Mean
81-ILC-A Powder	AAL-3513	A 0.154 B 0.164 C 0.156	$\overline{X}$ = 0.158 ± .005
81-ILC-B[1] Powder	AAL-3514	A 0.55 B 0.47 C 0.59	$\overline{X}$ = 0.54 ± .05
81-ILC-C Powder	AAL-3515	A 1.14 B 1.16 C 1.17	$\overline{X}$ = 1.16 ± .04

1982 Results on Same Samples as reported in Wehmiller, in press.

		Liquid	
81-ILC-A	AAL-2691	0.19 ± .01	
81-ILC-B	AAL-2692	0.585 ± .02	No Frees Analyzed
81-ILC-C	AAL-2693	1.16	

		Powders	aIle/Ile Free
81-ILC-A	AAL-2694D	0.151	0.53 ± .03
81-ILC-B	AAL-2695D	0.54	≈1.08
81-ILC-C	AAL-2696D	1.08	1.40

[1]Wehmiller (pers. comm. to G.H. Miller, Nov., 1983) remarked that other laboratories also obtained more variation for ILC-B than for other samples.

preparing separate shell chips for the Free and Total fractions and by dissolving, hydrolyzing, and drying the Total fraction exclusively in one vial. The aIle/Ile ratios produced after this preparation are generally 40% ± 5% lower than those produced by the 81 prep (see Miller, in press, for actual conversion factors). The aIle/Ile ratios in the Free fraction of both preps are statistically the same.

Following the 1980 field season, over 220 valves, most of which were Cardita, were analyzed using the 1981 preparation method. Following the 1981 field season, however, few samples were prepared until after the use of the 76b preparation was initiated (total of 216 valves). Both sets of data produced similar stratigraphic groups however, I had not differentiated the two oldest groups in the 1981 prep results that are now apparent in the 76b results. This is partly due to the fact that I had not yet analyzed a sufficient number of samples from the oldest units. Because the 76b preparation provides the truest measure of the extent of epimerization and hydrolysis in the fossils, results from the 76 b prep were used exclusively to define the aminozones described in this study. Repreparations of numerous samples from the first field season provide a means of calibrating 1981 prep results so that data could at least be used for stratigraphic correlations (Table 3.4).

Data Reduction

The HPLC amino acid analyzer systematically resolves aspartic acid, theronine, serine, glutamic acid, glycine, alanine, valine, methionine, isoleucine, and leucine. The diastereomers of theronine are resolved as one peak. The system resolves D-alloisoleucine from its diastereomer, L-isoleucine, and also resolves L-norleucine, a non-protein amino acid added as an internal standard. During each analysis, a Hewlett-Packer Computing Integrator (Model 3390A) automatically computes the area under each amino acid peak on the chromatogram. This data from both the Free and Total fractions was then entered onto a random access disk file on an Apple II computer. Upon demand, a data reduction program (written by Margaret Eccles, Institute of Arctic and Alpine Research) was used to then calculate the absolute concentration of each amino acid based upon the norleucine calibration peak added to each sample during preparation. The program also calculates the relative abundances of each amino acid, the aIle/Ile ratio and the ratio of several unstable/stable amino acids in both fractions. Both the Free and Total data were also used to compute the percentage of free amino acids in each sample.

Recognition of Aminostratigraphic Groups

Data From Mollusks

AIle/Ile ratios were determined on over 400 valves of Hiatella, Mya, Macoma, and Astarte collected from the Gubik

Table 3.4. Isoleucine Epimerization in Mollusks from the Gubik Formation Using the Now-abandoned 81 prep method.

Horizon	Species	AIle/Ile Free	Total	
Pelukian	Ha	ND	0.024 ± .004	(n = 8)
	Mt	ND	0.020 ± .002	(n = 5)
Middle Pleistocene	Ha	0.36 ± .037	0.061 ± .002	(n = 15)
	Mt	0.35	.064	(n = 8)
Early Pleistocene	Ha	0.61 ± .07	0.158 ± .05	(n = 7)
	Mt	0.51 ± .05	0.158 ± .02	(n = 12)
Late Pliocene	Ha	0.68 ± .03	0.333 ± .03	(n = 4)
	Mt	0.97 ± .06	0.213 ± .05	(n = 4)

Formation. Based upon the samples prepared under the 76b procedure, 5 groups or aminozones were recognized, each of which represents a period of high sea level (Table 3.5). The term aminozone is used in this study to refer to a specific range or grouping of aIle/Ile ratios that characterize a single stratum and represent a unique depositional time period. An aminozone is recognized by analyzing numerous monospecific shells from areas where the stratigraphic relationships are clear.

A plot of the Free verses the Total aIle/Ile ratios for each genus demonstrates the consistent increase in both ratios with increasing stratigraphic age (Figs. 3.3 - 3.6). Scatter in the older samples typically increases, although the data for aminozones 4 and 5 are sparse. Astarte ratios contain the greatest amount of scatter relative to the other genera although the reason for this is not known.

With increasing age, the total percentage of free amino acids also increases as peptide bonds are continually broken (Table 3.6). A plot of the percentage of free leucine, a stable amino acid, illustrates, in particular, the strong relationship between the extent of epimerization in the Total fraction and the extent of hydrolysis irregardless of genus (Fig. 3.7). Although the Free aIle/Ile ratio increases faster than the Total aIle/Ile ratio, epimerization of hydrolyzed amino acids is actually slower in the free fraction. The rate at which this reaction approaches equilibrium decreases as the proportion of D-alloisoleucine increases and its inversion becomes increasingly significant. Thus, the age difference between a Free ratio of 0.7 and 0.8 is greater than the age difference between Free ratios of 0.3 and 0.4. This apparent change in the epimerization rate is reflected for each genus in Figs. 3.3 through 3.6. For Hiatella, Mya, and Macoma this change occurs between a Free aIle/ratio of 0.4 to 0.5, but for Astarte the change is closer to 0.6. The fact that Macoma consistently yields higher aIle/Ile ratios than the other genera in the same strata probably reflects a faster rate of hydrolysis in Macoma (Table 3.6; c.f., Wehmiller, 1980).

Data From Wood

Samples of fossil wood were collected along with mollusks for amino acid analysis from several stratigraphic sections. Although aIle/Ile ratios in wood have been used by a few workers for differentiating and dating Wisconsinian age material in the Yukon (Rutter and others, 1980) and the Puget Lowland (Easterbrook and Rutter, 1982), the potential of this material for amino acid analysis in older strata has not been thoroughly demonstrated. There is apparently some data to suggest that older samples may be inherently contaminated with bacteria, and that an alternative preparation method may be necessary for samples older than 100 ka (N.W. Rutter, pers. comm., August, 1983).

As a cooperative project, N.W. Rutter, University of Alberta, processed several samples of wood and shell from the Gubik

Table 3.5. Extent of Isoleucine Epimerization in Marine Units of the Gubik Formation, western Arctic Coastal Plain, Alaska[1].

HIATELLA

Aminozone	F	(CV%)[2]	T	(CV%)
Modern	ND[3]	(n = 8)	0.0115 ± .0015	(13%)
1	ND	(n = 20)	0.014 ± .003	(21%)
2	0.40 ± .052	(13%) (n = 28)	0.038 ± .007	(18.4%)
3	0.52 ± .045	(8.6%) (n = 11)	0.090 ± .018	(20%)
4	0.58 ± .08	(14%) (n = 8)	0.150 ± .025	(16%)
5	0.75 ± .07	(9.3) (n = 4)	0.235 ± .017	(7.2%)

MYA

Aminozone	F	(CV%)	T	(CV%)
Modern	----		----	
1	ND	(n = 5)	0.018 ± .003	(20%)
2	0.36 ± .097	(27%) (n = 27)	0.040 ± .009	(22.5%)
3	0.54 ± .08	(15%) (n = 27)	0.083 ± .014	(16.8%)
4	0.64 ± .12	(18.7%) (n = 4)	0.158 ± .030	(19%)
5	----		----	

MACOMA

Aminozone	F	(CV%)	T	(CV%)
Modern	ND		0.013 ± .002	(15.4%)
1	ND	(n = 7)	0.024 ± .003	(12.5%)
2	0.47 ± .08	(17%) (n = 26)	0.085 ± .014	(16.5%)
3	0.56 ± .04	(7.1%) (n = 4)	0.136 ± .010	(7.3%)
4	0.61 ± .05	(7.5%) (n = 6)	0.189 ± .009	(5%)
5	----		----	

ASTARTE

Aminozone	F	(CV%)	T	(CV%)
Modern	----		----	
1	ND	(n = 11)	0.021 ± .005	(24%)
2	0.45 ± .14	(31%) (n = 10)	0.058 ± .015	(26%)
3	0.58 ± .10	(16.3%) (n = 12)	0.071 ± .019	(26.8%)
4	0.76 ± .01	(1.3%) (n = 3)	0.118 ± .005	(4.2%)
5	0.89 ± .10	(11.2%) (n = 4)	0.185 ± .025	(13.5%)

1 Values represent peak height measurements, mean ± $^{1}\alpha$; F = Free, T = Total amino acid fraction
2 CV % refers to the coefficient of variation, where CV% = (standard deviation/lmean) X 100
3 ND = not detectable

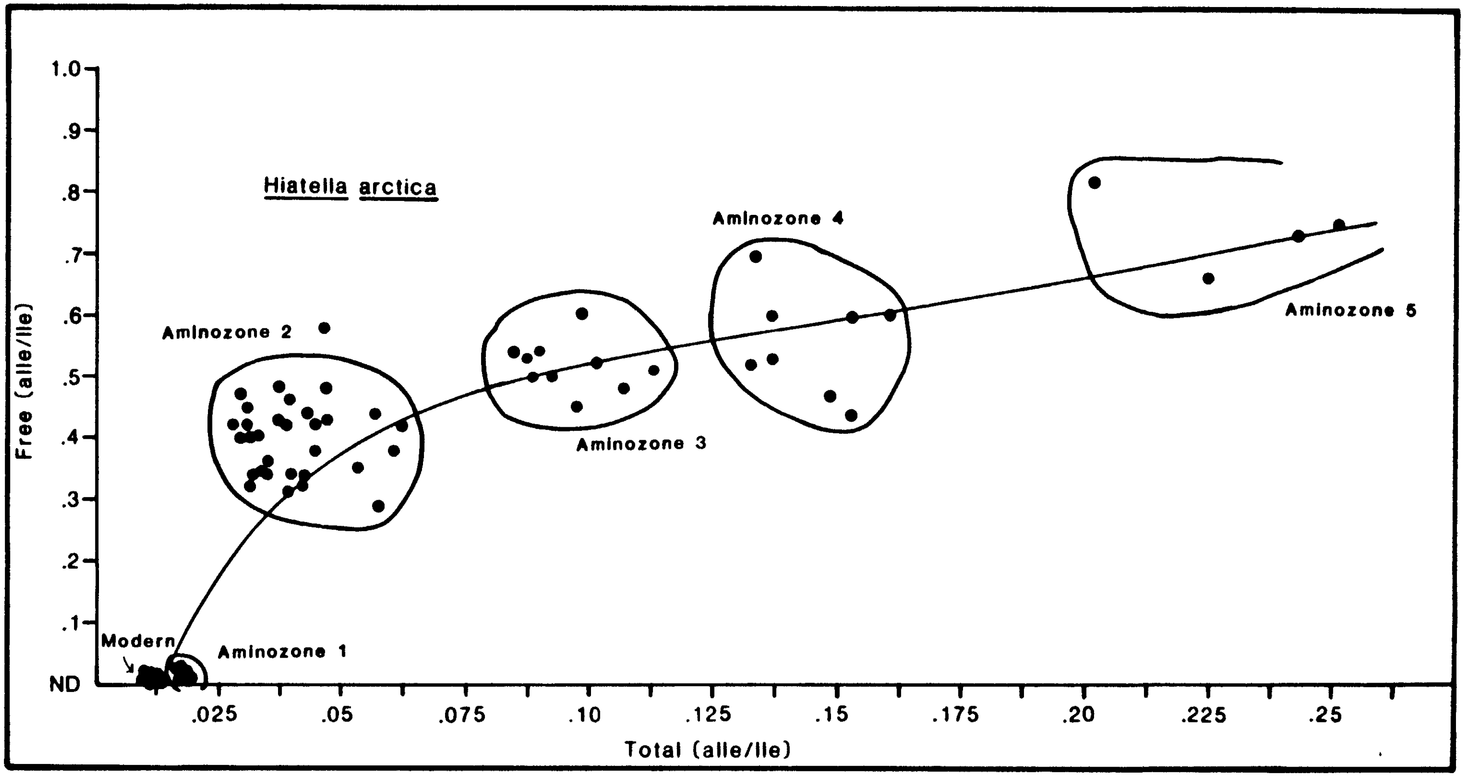


Fig. 3.3. Plot of the Free versus the Total aIle/Ile ratio in all samples of <u>Hiatella arctica</u> from the Gubik Formation (76b prep). Each point represents <u>an individual</u> valve. This species is best for differentiating deposits of different age.

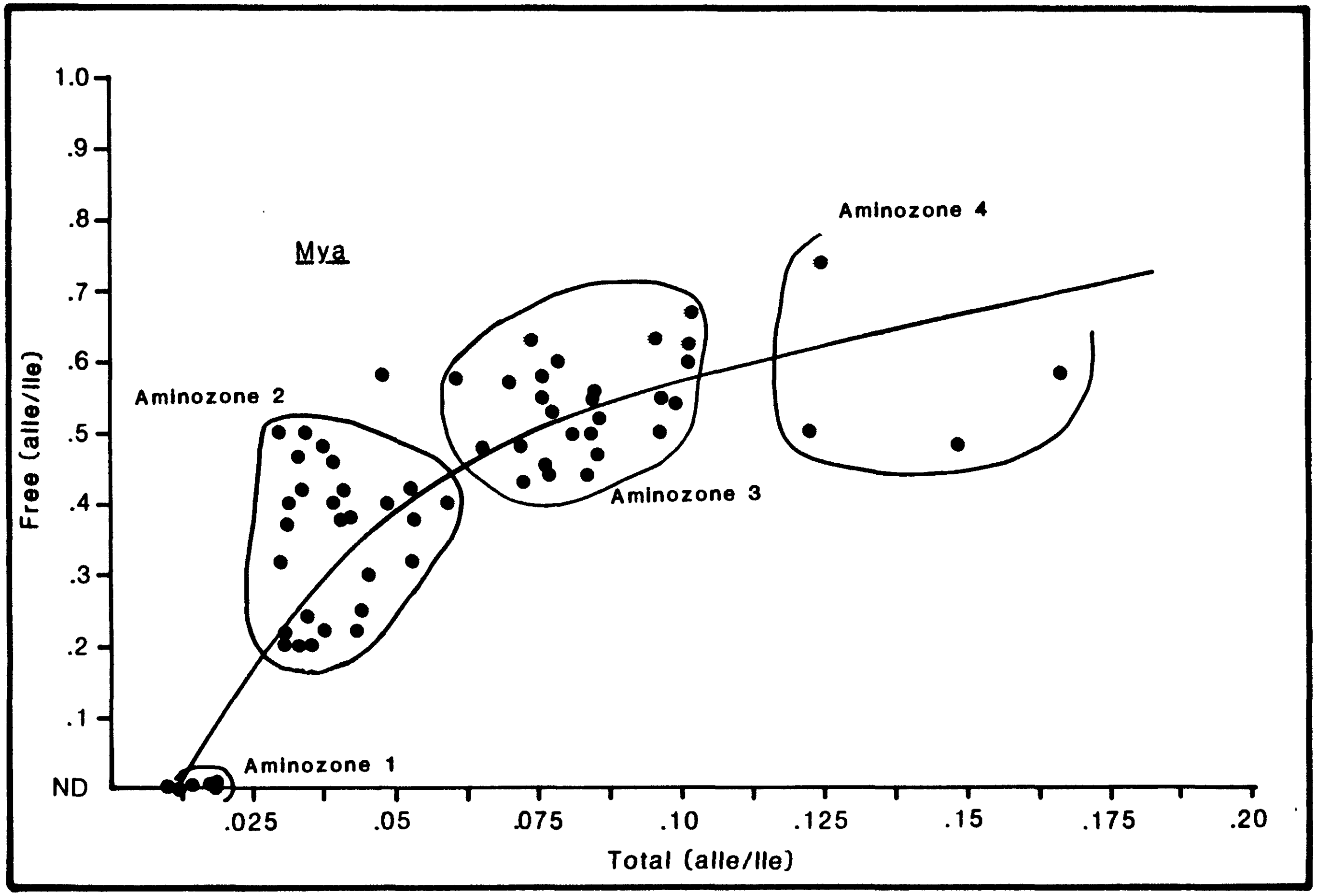


Fig. 3.4. Plot of Free versus the Total aIle/Ile ratio in all Mya from the Gubik Formation (76b preparations).

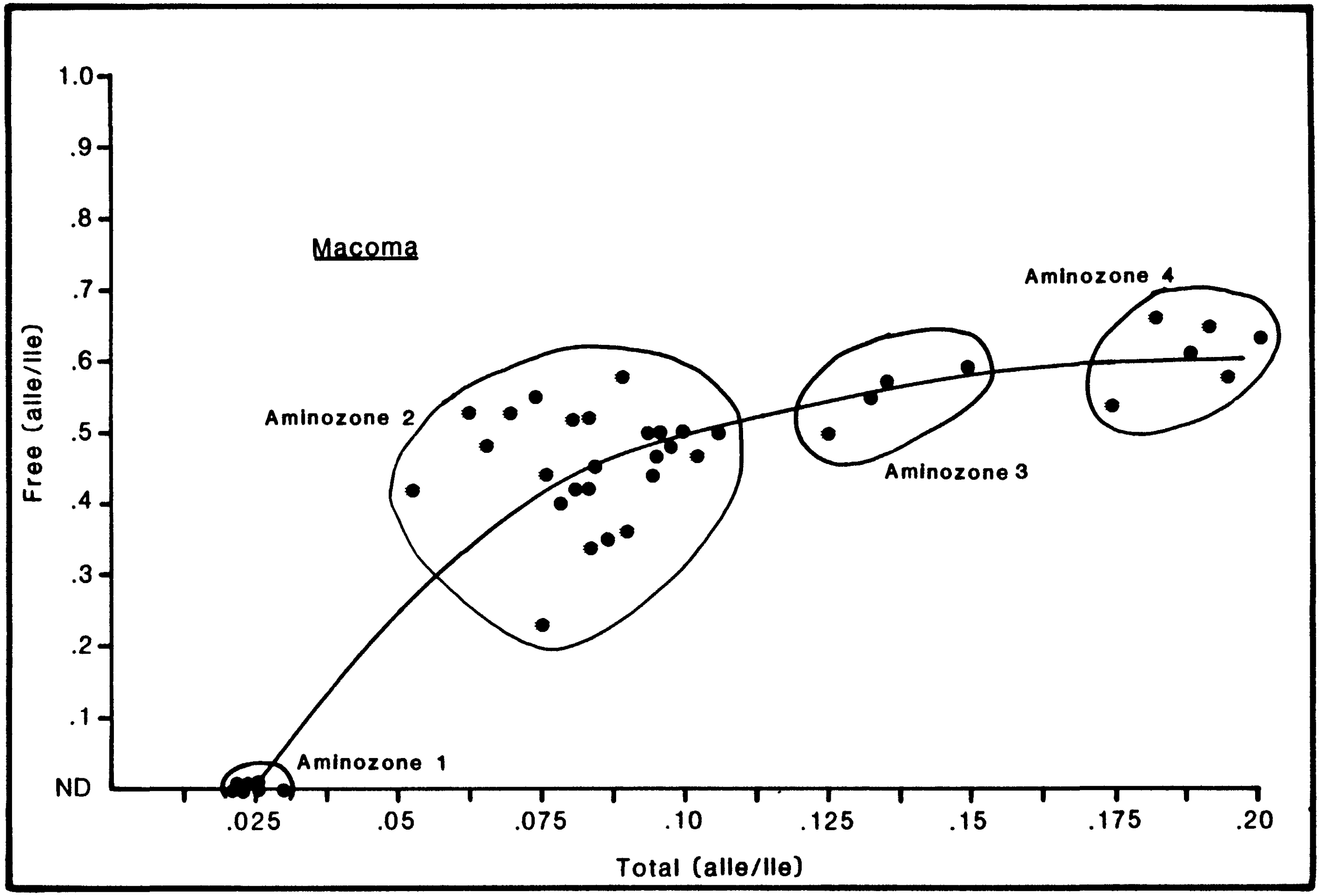


Fig. 3.5. Plot of the Free versus the Total alle/Ile ratio in all *Macoma* from the Gubik Formation (76b preparations).

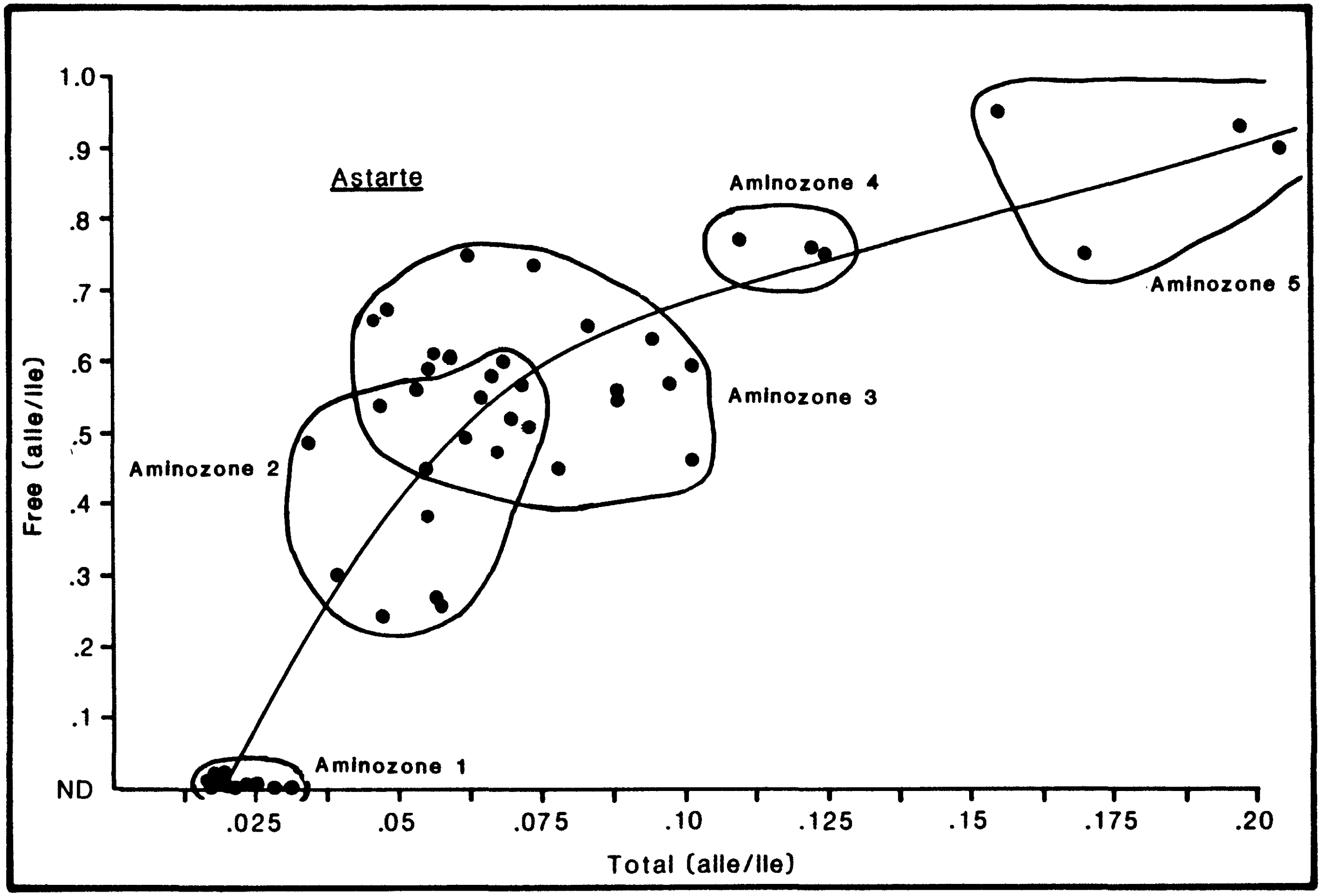


Fig. 3.6. Plot of the Free versus the Total aIle/Ile ratio in all *Astarte* from the Gubik Formation (76b preparation). *Astarte* display more scatter than the other genera.

Table 3.6. Percentage of Free Amino Acids including Aspartic acid through Leucine from Stratigraphic Units of the Gubik Formation.[1]

Aminozone	Hiatella	Mya	Macoma	Astarte
Modern	0 - .2 (6)	----	----	----
1	1.6 ± .6 (11)	1.6 ± .9 (6)	3.3 ± 1.0 (6)	2.6 ± .4 (3)
2	7.7 ± .3 (25)	7.5 ± .26 (22)	13.7 ± 2.9 (18)	12.3 ± 4.0 (5)
3	12.5 ± 3.0 (14)	15.2 ± 5.2 (11)	25.0 ± 7.5 (8)	10.2 ± 1.6 (5)
4	21.4 ± 8.16 (2)	17.0 ± 6.3 (4)	35.2 (1)	17.0 ± 5.3 (3)
5	38.3 ± 16.7 (4)	----	----	30.6 ± 12.9 (4)

[1]Number in parentheses indicates number of valves included in the mean.

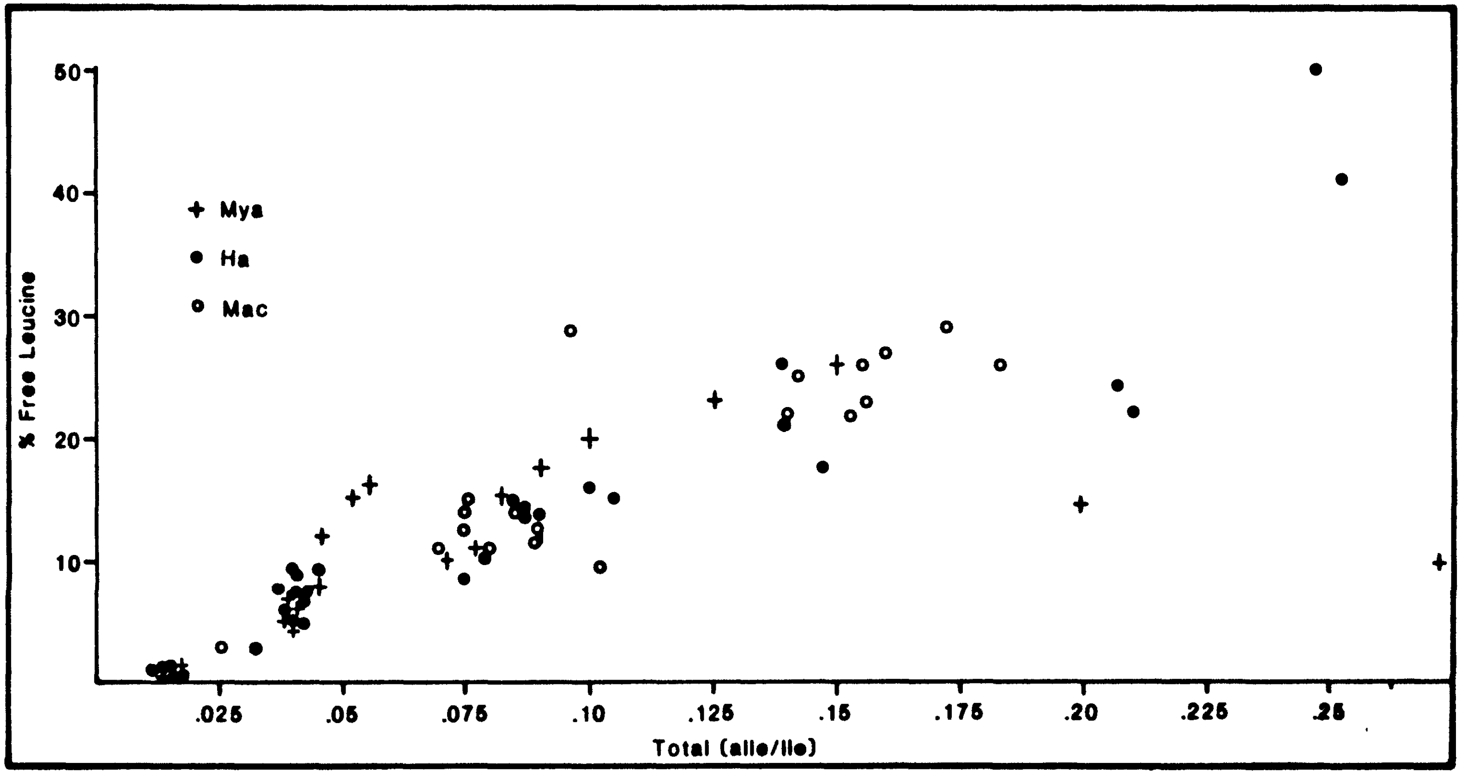


Fig. 3.7. Plot of the percentage of free Leucine versus the Total aIle/Ile in <u>Mya</u>, <u>Hiatella</u> (Ha), and <u>Macoma</u> (Mac).

Formation, using gas chromatography to examine the extent of racemization of L-aspartic acid to D-aspartic acid in the samples. Although glutamic, leucine and proline show a great deal of scatter, the results from these analyses (Table 3.7) yield aspartic acid D/L ratios that are consistent with their stratigraphic position. The only exceptions are samples 81 Akb 545 and 81 Akb 580 which gave ratios similar to the aminozone 1 wood samples. The reason for this is unknown. Results from wood are not nearly as species-dependent as mollusks (N. Ruter, pers. comm., 1982), therefore, the fact that both of these samples are willow should not be critical.

Like the HPLC data on mollusks from the same aminozone, the enantiomeric ratios in the wood samples are low. For example, wood samples with aspartic ratios of 0.14 from the Old Crow Basin, northern Yukon, are considered to be mid-wisconsin in age (Hughes and others, 1981) whereas infinitely dated, last interglacial samples from northern Alaska yield a ratio of about 0.15.

Summary

The extent of isoleucine epimerization measured in fossil *Hiatella*, *Mya*, *Macoma*, and *Astarte* subdivides the stratigraphic units of the Gubik Formation into five groups, or aminozones. The extent of hydrolysis in the matrix protein of these fossils, indicated by the percentage of free amino acids, concurs with this subdivision. In concert with traditional stratigraphic methods, these aminozones provide a means of correlating disjunct sections, especially those containing different yet contemporaneous marine lithofacies such as those outlined in Chapter II. The absolute age of these aminozones is outlined in Chapter VII following a full evaluation of the stratigraphic and paleontologic evidence described in Chapters, II, IV and V.

Table 3.7. Extent of Racemization in Wood and Shell from the Gubik Formation.[2]

Aminozone	Akb Field No (Sec.)	Lab No.	Species	D/L Ratio[1] ASP	GLU	LEU	PRO
			WOOD				
Holocene	653	UA-1281	willow	0.051	0.096	0.024	---
1	210 (36)	UA-872	spruce	0.131	0.066	0.024	0.020
	231 (38)	UA-871	spruce	0.158	0.074	0.025	0.023
	248 (46)	UA-873	spruce	0.143	0.053	-----	-----
	523 (78)	UA-1276	spruce	0.175	0.079	0.018	-----
	556 (81)	UA-1279	spruce	0.125	0.057	0.017	-----
2	193 (32)	UA-874	spruce	0.288	0.095	0.118	0.021
	580 (90)	UA-1280	willow	0.127*	0.064	0.036	-----
	545 (80)	UA-1278	willow	0.191*	0.082	0.017	-----
	658 (112)	UA-1282	alder	0.238	0.0521	0.016	-----
			SHELLS				
1	249 (47)	UA-877	_Hiatella_	0.084	0.027	0.015	0.019
	214 (38)	UA-876	_Hiatella_	0.073	-----	-----	0.018
$\geq$3	211 (36)	UA-876	_Cardita_	0.446	-----	0.191	0.146

[1]ASP, aspartic acid; GLU, glutamic acid; LEU, leucine; PRO, proline.
[2]Samples analyzed by Dr. N. Rutter, University of Alberta.

CHAPTER IV
MORPHOLOGY AND STRATIGRAPHY OF PALEOSHORELINES

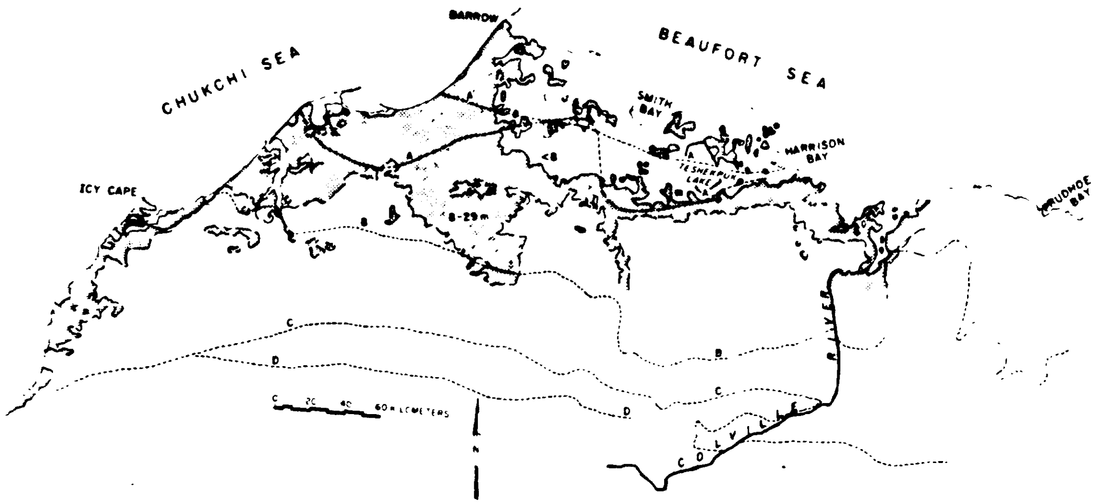
> It is said that after the earth turned over, all of the land was covered with water...It is said that after there was the great flood these...leftover bones of the whales [were] stranded inland.
>
> --Elijah Kakinnaag, during the 1978 Elders Conference, Barrow.

The Arctic Costal Plain of Alaska is a monotonously flat landscape that developed from repeated episodes of marine and subaerial erosion and deposition throughout Tertiary and Quaternary time. The Tertiary history of the western coastal plain is poorly known. However, it is apparent that much of this period was dominated by the erosion and pedimentation of folded Cretaceous bedrock. South of Icy Cape, more strongly folded rocks are unconformably overlain by unconsolidated Pleistocene sediments. In contrast to the eastern portion of the coastal plain where the Sagavanirktok Formation records the predominantly non-marine Tertiary history, the western coastal plain lacks little record of this period. The Miocene Papigak clay discovered overlying Cretaceous rocks along Skull Cliff during the course of this study (and discussed in Chapter II) is equivalent to the upper marine Nuwok member of the Sagavanirktok Formation and effectively requires that much of the coastal plain erosion took place prior to the Miocene. These are the only known pre-Pliocene rocks of Tertiary age on the coastal plain west of Barrow.

During the middle and late Pliocene and Pleistocene, the western coastal plain was repeatedly inundated by the Chukchi Sea in response to changes in world sea level and more local, subtle tectonic movements. Although direct evidence indicating the landward extent of the earliest marine transgressions is lacking, evidence of the younger events is patchily preserved where it has not been obscured by periglacial activity or removed by fluvial processes. From topographic map studies, interpretations of side-looking airborne radar imagery, and subsequent field reconnaisance, portions of marine strandlines and terrace surfaces were identified.

Previous Shoreline Age Estimates

O'Sullivan (1961) made the first notable attempt to recognize and map the distribution of Pleistocene marine shorelines (Fig. 4.1). During the course of three field seasons he examined exposures of the Gubik Formation across the entire coastal plain and identified a series of surfaces that he considered to correspond to major transgressive events. His study was hindered by the lack of adequate topographic maps in 1957 and by the fact that he was unable to accurately date the surfaces and geomorphic features he was attempting to correlate. Discrepancies of 200-300



A.ʹ Late Sangamon or mid-Wisconsin (O'Sullivan 1966)
A. Wisconsin (Black 1964)
B. Sangamon (O'Sullivan 1966)
C. Mid-Pleistocene (O'Sullivan 1966)
D. Late Pliocene (O'Sullivan 1966)
Surfaces less than 8 m and 8 to 29 m in elevation are also shown.

Figure 4.1. North Slope showing limits of major transgressions as proposed by O'Sullivan (1961) and Black (1964). Diagram from Sellman and others (1975).

ft (60-90 m) in the 500 ft (150 m) contour line were apparent between older reconnaissance base maps and newer topographic maps available at the time of his study (O'Sullivan, 1961, p. 138). Despite these inconsistencies, O'Sullivan concluded that four Pleistocene shorelines could be recognized, the three oldest of which had been differentially uplifted as much as 170 m across the eastern coastal plain. The fourth and youngest shoreline was inferred to stretch from the mouth of the Kogru River eastward past Teshekpuk Lake and intersect the Chukchi Sea coast near Skull Cliff.

Black (1964) later summarized knowledge of the Gubik Formation incorporating results of field work that he had conducted between 1945 and 1950. Like O'Sullivan, he attempted to recognize and correlate paleoshorelines on the coastal plain based upon stratigraphic studies and geomorphic arguments (Fig. 4.1). Black subdivided the Gubik Formation into 3 lithologic units each of which in part, represented marine transgressions and he assigned to them ages of Illinoian, Sangamon, and Wisconsin.

His observation that,

> ... the boundary between the foothills and the coastal plain reveals, in many places, a wave-cut cliff that is warped from a point near sea level on the west near Cape Beaufort to 600 ft [183 m] elevation south of Barrow and down to a few tens of feet near Umiat....

concurred with O'Sullivan's observations that some surfaces appear to be warped. They differed, however, in their interpretations of <u>how</u> the surfaces were warped (hence how different features correlate) and concerning the absolute age of presumed strandline features (Figure 4.1).

McCulloch (1967) subsequently produced a more detailed map of Pleistocene shorelines across the western coastal plain between Peard Bay and Cape Beaufort based upon stratigraphic work during boat traverses along the major rivers (Fig. 4.2). He identified two middle Pleistocene transgressions (labeled "middle Pleistocene" and pre (?)-Illinoian") at 60-70 m a.s.l. and about 33 m respectively, and outlined evidence for a continuous Sangamon shoreline along the coast up to 13 m a.s.l. Age estimates were based upon molluskan biostratigraphy and geomorphic expression. His interpretations differ from those of O'Sullivan (1961) and Black (1964) and provide more informative detail.

A more recent study of Pleistocene shorelines west of Barrow is that of Williams (1983a, 1983b) in the Meade River and Wainwright Quadrangles (also see Williams and others, 1977). Relying primarily on geomorphic evidence and the spatial analysis of the surficial materials in the area, Williams recognizes at least 3 marine transgressions that include only a portion of the strandlines recognized by McCulloch (1967) (Fig. 4.3). His youngest strandline lies well below 20 m a.s.l. and is believed to delineate the Pelukian or last interglacial (125 ka B.P.) high sea

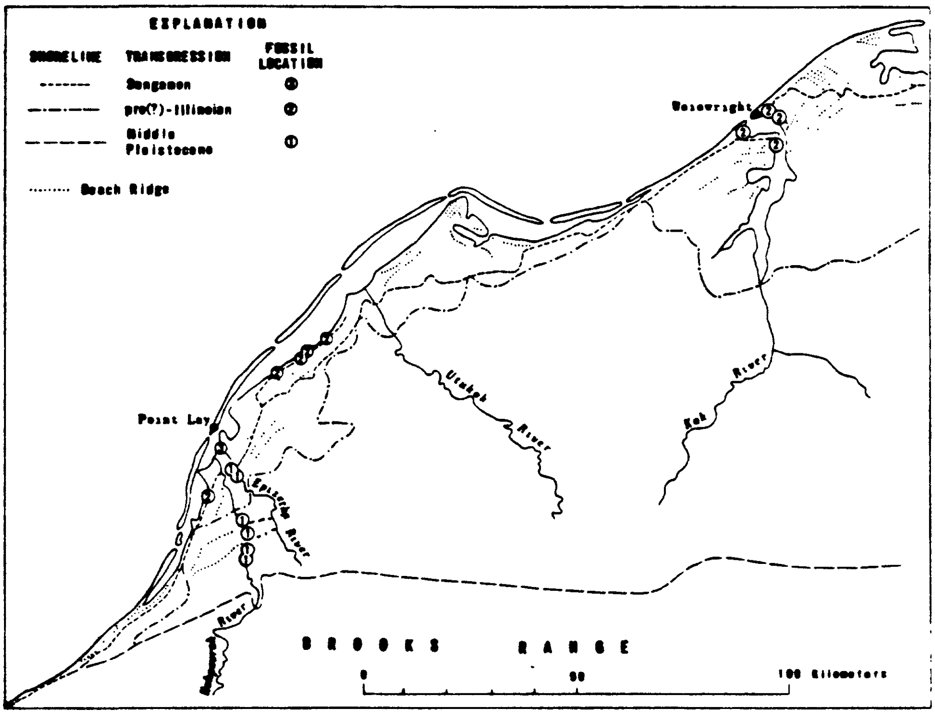


Figure 4.2. Map of the far western coast plain showing the shorelines mapped by McCulloch (1967).

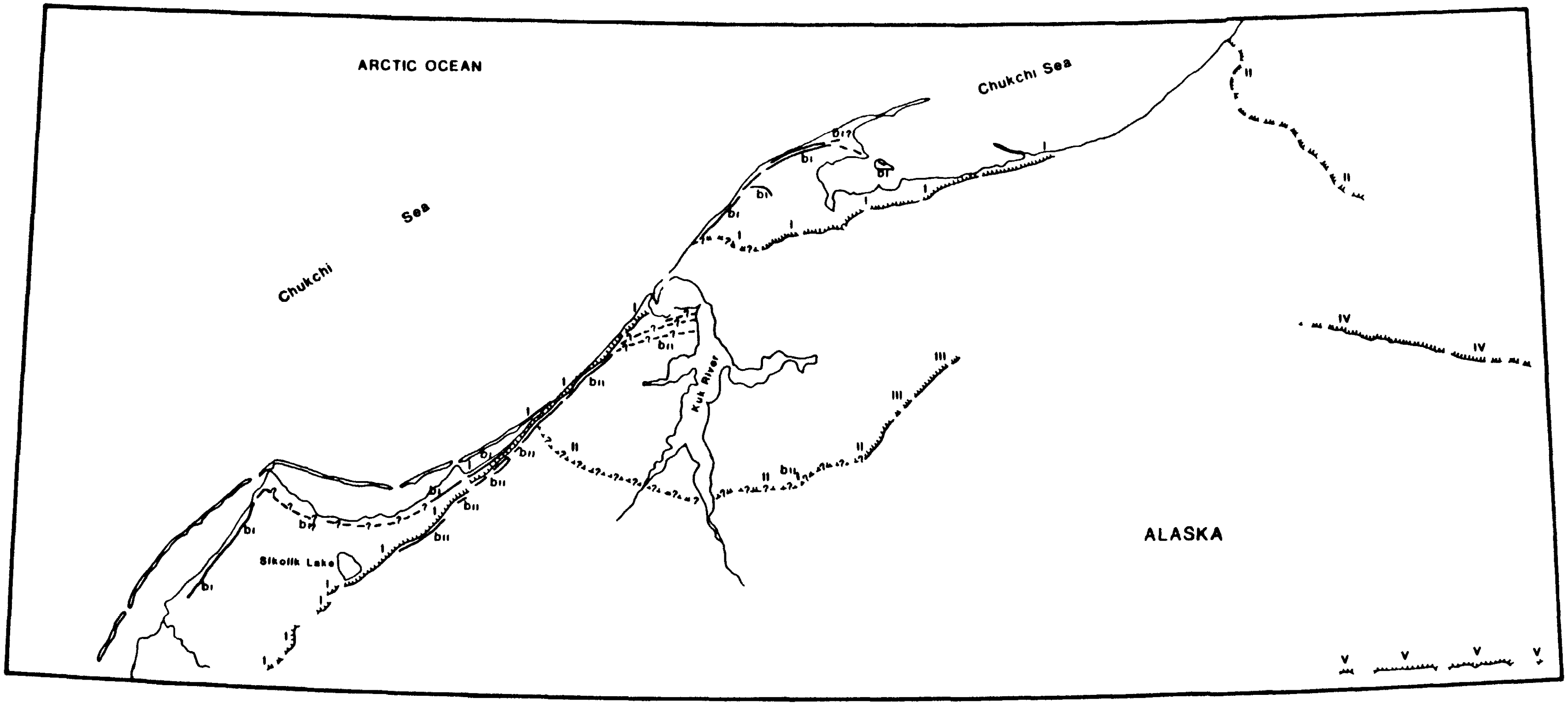


Figure 4.3 Marine shorelines between Skull Cliff and Icy Cape as mapped by Williams (1983a, 1983b). I, marine wave-cut cliff indicting sea level about 20 m above present; II and III, indistinct shoreline modified by thaw-lake activity. Sea level ≈ 20 m asl - 30 m asl, correlated with pre-Illinoian (McCulloch, 1967) or Kotzebuan transgression (Hopkins, 1967a); IV, Escarpment bounding northern limit of dune field, probably not of marine origin; V, excarpment between 38 and 68 m asl separating dune fields on North from upland silt, probably cut by streams.

stand. This strandline, in part, abuts an ancient sea cliff composed of Cretaceous bedrock (Nanushuk Group) overlain by older unconsolidated marine deposits. Sediments higher than the last interglacial strand but below a second, higher strand at roughly 30 m a.s.l. are interpreted by Williams as "pre-Illinoian or Kotzebuan" in age representing a major mid-Pleistocene transgression (McCulloch, 1967; Williams, 1983a, 1983b). Marine deposits above this strandline are inferred only as pre-Kotzebuan in age, as interpreted earlier by McCulloch (1967) and Hopkins (1967a).

Stratigraphic investigations have been carried out in the Barrow, Meade River, and Wainwright quadrangles by D.M. Hopkins and collaborators, U.S.G.S. (1976-1979), but their work is unpublished. Field notes and paleontologic reports from these investigations were made available to J.R. Williams and myself during our somewhat parallel but independent studies.

Reconnaissance observations - Marine Gubik/Cretaceous Boundary

Reconnaissance studies of the western Arctic Coastal Plain facilitated mapping of the boundary between unconsolidated marine deposits and Cretaceous bedrock from Wainwright southward toward Cape Beaufort (Fig. 4.4). The contact was identified by examination of outcrops along the Utukok, Kokolik, Epizetka, and Kukpowruk rivers, along exposures around thaw lakes, and from surface sediments exposed through the tundra. In particular, this mapping delineated for the first time that the marine cover between the Utukok and Kukpowruk rivers outlines, in effect, a large embayment separated from an embayment along the Kuk River by a peneplained promontory of Cretaceous bedrock, inland from Icy Cape (Fig. 4.4). These areas are geographically addressed as the Point Lay embayment, the Icy Cape promontory, and the Kuk River embayment.

The inner edge of the marine Gubik/Cretaceous boundary is delimited by patchy fossiliferous gravel and sand overlying bedrock between 30 m and 45 m a.s.l. as observed at Sec. 87 and 88 on the Kukpowruk River, and Sec. 109 on the Utukok River. Elsewhere, the boundary is defined by patches of gravel and sand that thin or terminate at the base of a gentle escarpment above 33-40 m. For example, the eastern shore of a large unnamed lake between the Utukok and Kokolik Rivers ("Pinnedin Lake") is lined with Cretaceous bedrock. The western shore of this lake, however, is confined by unconsolidated sediments. Similarly, the southeastern shores of Tunusiktok and Pugutak lakes are lined with bedrock, but both lakes are bisected by a ridge of fossiliferous, well-rounded gravel and sand that marks the approximate position of a paleoshoreline in that area (Fig. 4.5).

The marine Gubik/Cretaceous boundary north of the Utukok River occurs between 33 and 40 m a.s.l. and is locally delineated by patches of fossiliferous gravel north of Sikolik Lake (Williams and others, 1977) (Fig. 4.4). This boundary outlines a promontory

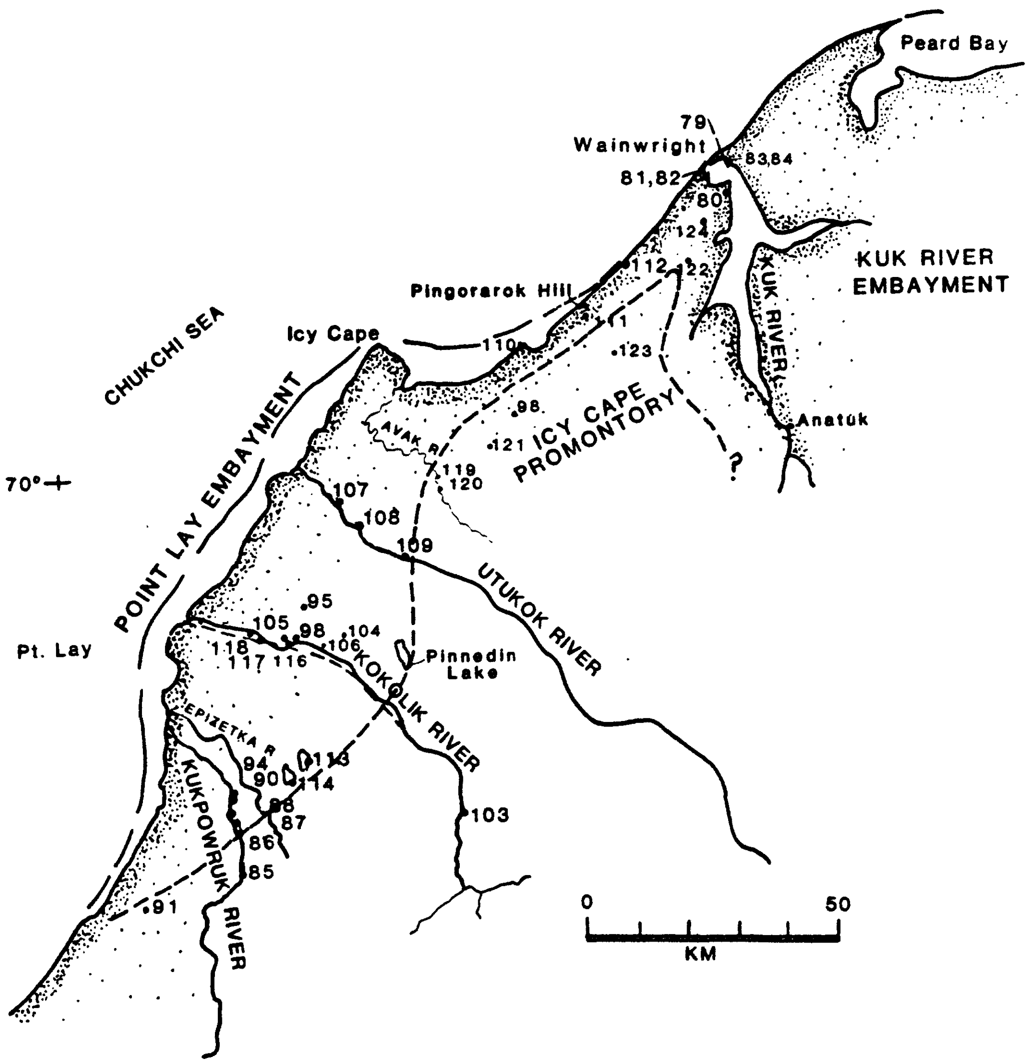


Figure 4.4. Map showing the approximate boundary (dashed) between marine beds of the Gubik Formation and Cretaceous bedrock.

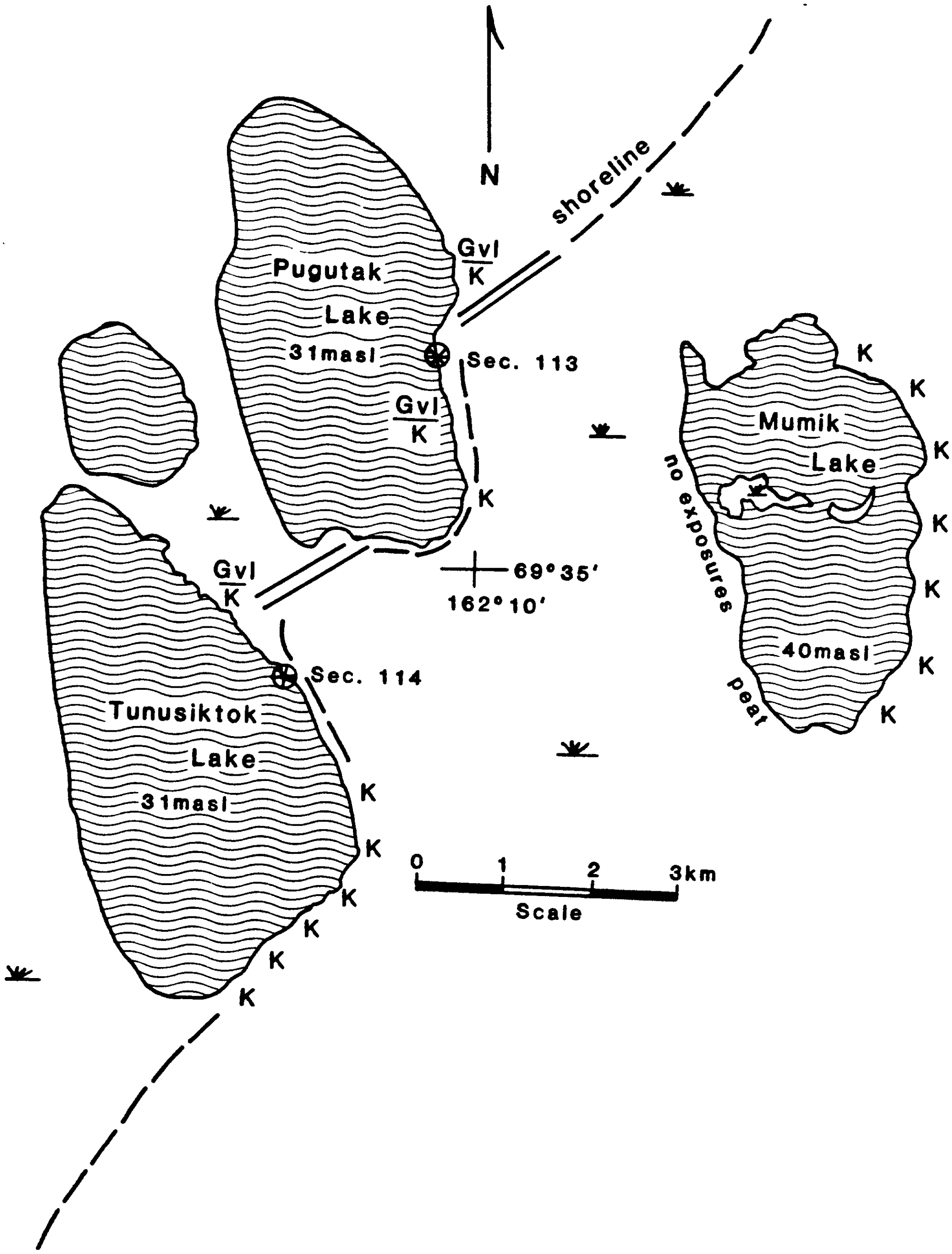


Figure 4.5. Map of gravel marine deposits (Gvl) where they overlie Cretaceous bedrock (K) north of the Epizetka River.

of Cretaceous bedrock that lacks any marine cover, although it was perhaps inundated by the sea at sometime prior to the early Pleistocene. At present, only a thin, discontinuous mantle of eolian sand occurs among frost boils over Cretaceous clay across this area.

Near Pingorarok Hill, half way between Icy Cape and Wainwright, the Gubik/Cretaceous contact swings southwestward, then eastward crossing the Kuk River a few km north of Anaktuk (Fig. 4.4). This boundary is drawn on the basis of geomorphic criteria that delineate the terrain of numerous, oriented thaw lakes underlain by significant accumulations of unconsolidated sediments from the terrain of fewer, irregularly shaped lakes where unconsolidated sediments are thin or absent and the lakes are underlain by Cretaceous bedrock. The Gubik/Cretaceous boundary east of the Kuk River has not been adequately studied.

Initial Mapping using SLAR Imagery and Longitudinal Profiles

Side-looking aircraft radar (SLAR) imagery of the Point Lay, Utukok River, Lookout Ridge, and Ikpikpuk River quadrangles (Fig. 4.6) were made available to me by John W. Cady, U.S.G.S., for the purpose of (1) mapping geomorphic features such as Pleistocene beach ridges, (2) ground-truthing the interpreted features in the field, and (3) determining the usefulness of this expensive imagery (Cady, 1984). The most useful aspect of SLAR imagery is that it is taken by aircraft at low angles (between 10° and 30°) that enhance the radiance of even subtle changes in slope (Jensen and others, 1967). On the North Slope, the SLAR imagery was particularly useful for identifying dissected portions of paleoshoreline features, as well as for identifying extensive river-bluff exposures for stratigraphic study. Such features could not be readily identified using conventional aerial photographs in this flat terrain, and the available 1:63,360 topographic maps with their contour intervals of 25 to 50 ft (8 to 15 m) often lack adequate resolution and detail. An example of the SLAR imagery from the Lookout Quadrangle is shown in Fig. 4.7. Prior to the 1981 field season, stereo strips of north- and south looking SLAR imagery were examined and paleoshorelines were mapped (Fig. 4.8). Many of the linear features observed in the Point Lay quandrangle coincided with paleoshores mapped by McCulloch (1967). Other ridges are evidently old dunes of unknown age. These features were notablydifficult to distinguish even from the air by helicopter.

Terrace surfaces were also distinguished by constructing longitudinal profiles along major river courses in the western coastal plain, where exposures of marine sediment could be examined, and along interfluves where surfaces presumably have been least dissected (Fig. 4.9).

Profiles (Figs. 4.10 - 4.13) were drawn by estimating surface elevations on 1:63,360 scale maps at 2 km intervals and connecting these points. The profiles represent only the gross characteristics of the true topographic profiles because of the

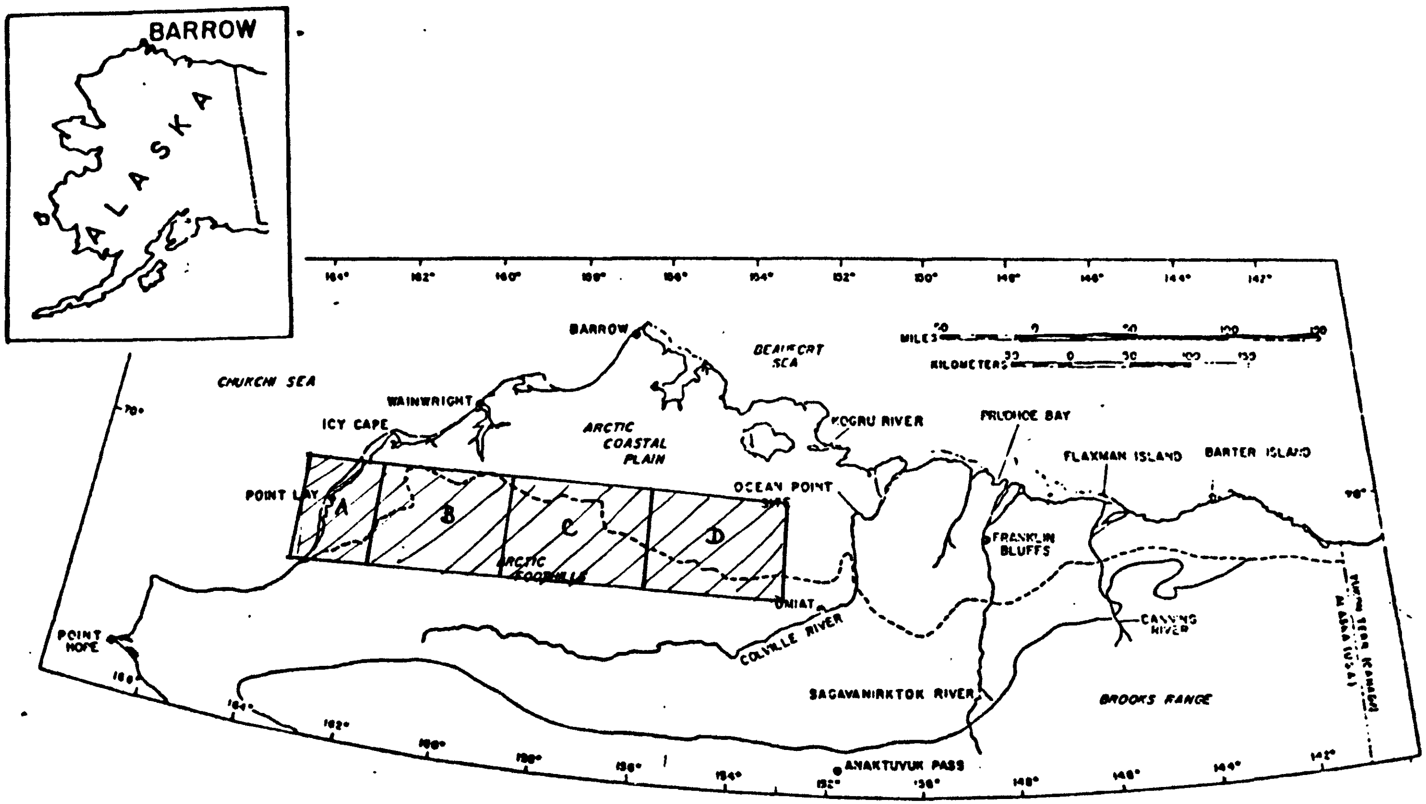


Figure 4.6. Map showing the quadrangles of the North Slope covered by a side-looking airborne radar imagery. (A) Pt. Lay, (B) Utukok River, (C) Lookout Ridge, (D) Ikpikpuk River.

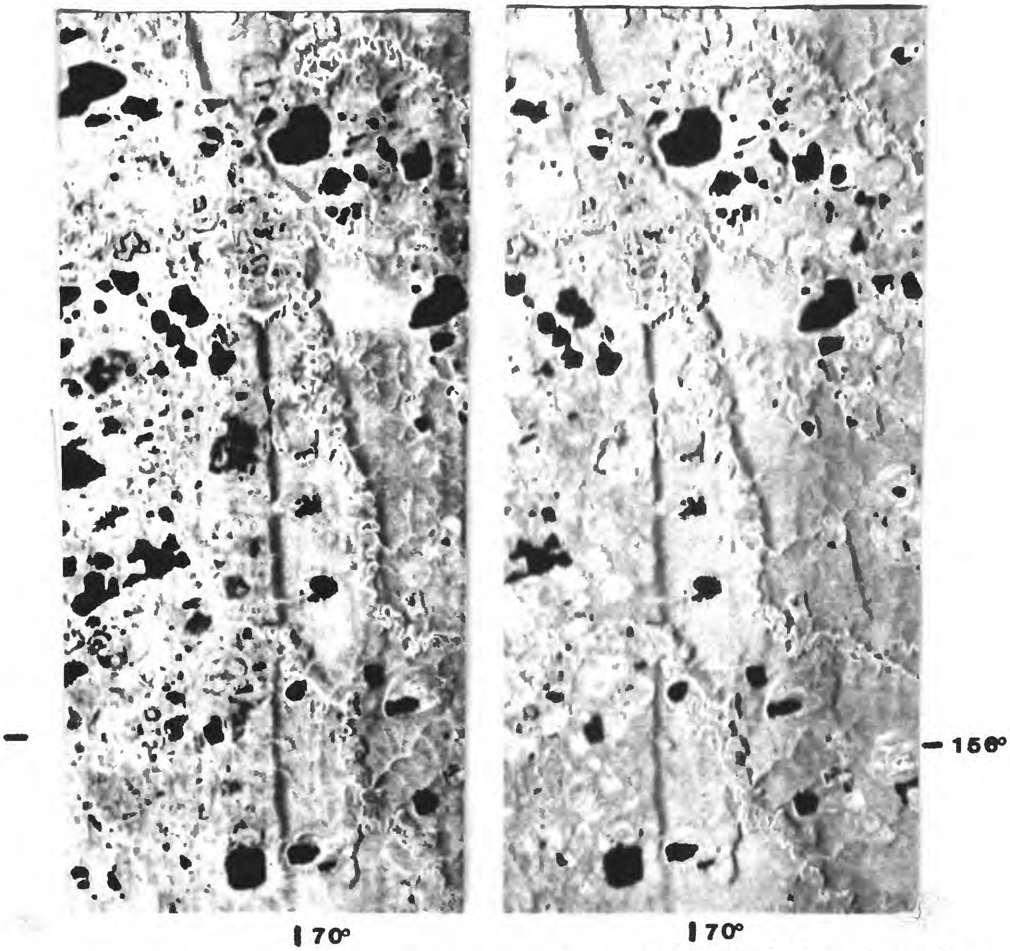


Figure 4.7. Side-looking aircraft radar (SLAR) imagery of a major break in slope about 20-30 m high at the north edge of the Lookout Ridge Quadrangle (from Cady, submitted manuscript). The stereo pair above were taken by radar looking toward the north, hence surfaces facing the radar appear lighter while slopes facing away from the radar (toward the north) appear darker. The Topagoruk River appears at the top of each stereo strip. The feature shown was first interpreted by Hopkins (1959) as a wave-cut scarp cut by a marine transgression of late Pliocene or early Pleistocene age. More recently, L.D. Carter (personal comm. to Cady, 1983) has interpreted the scarp as a river-cut bluff face, however both Cady and myself favor Hopkins interpretation.

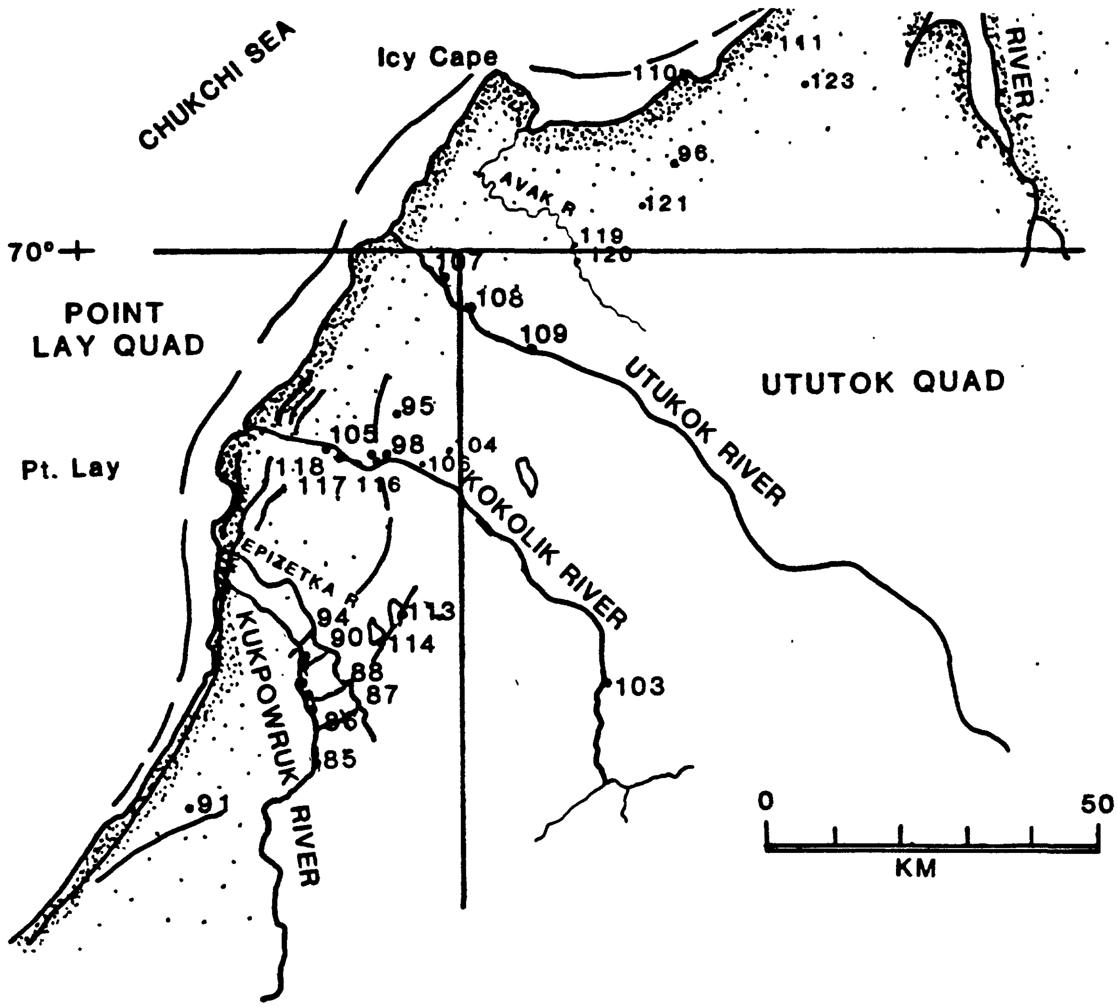


Figure 4.8. Map showing paleoshoreline features identified from SLAR imagery of the Point Lay Quadrangle.

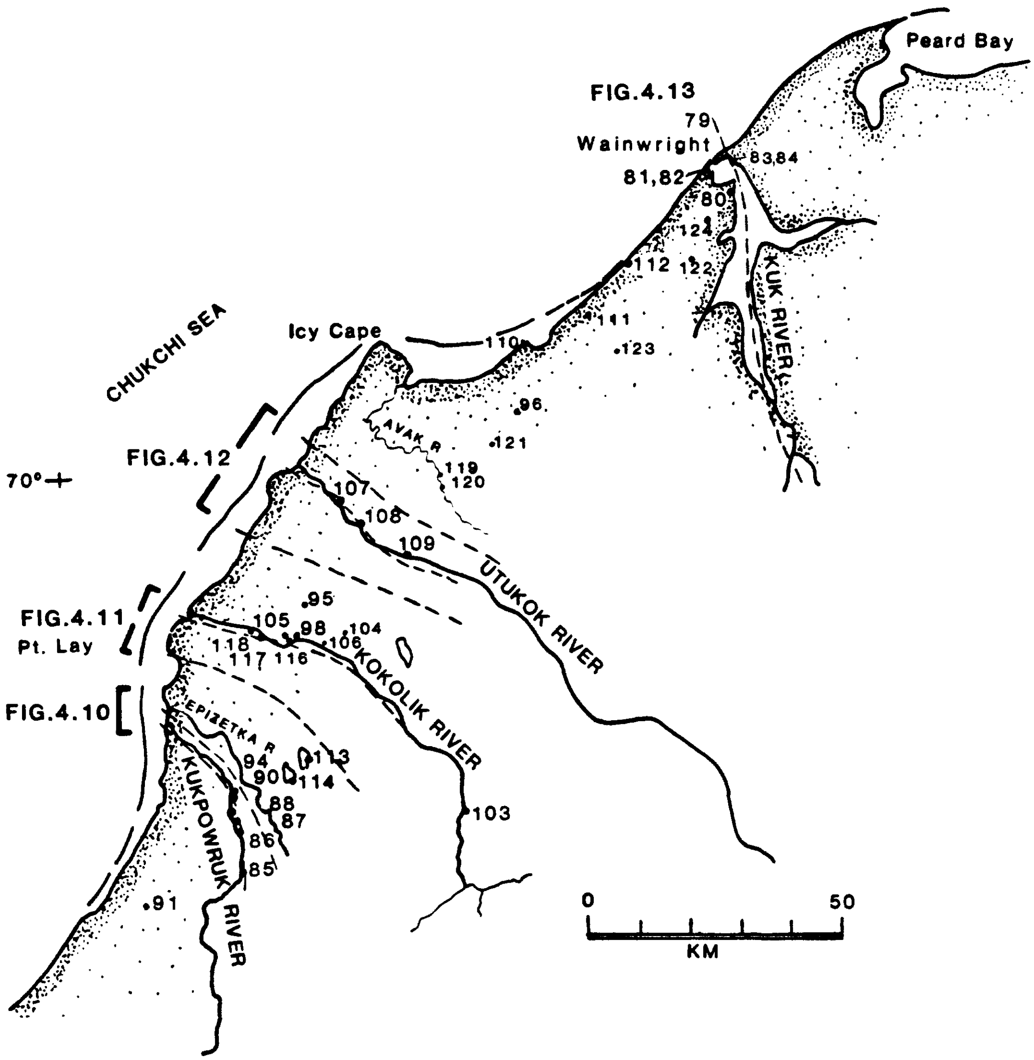


Figure 4.9. Map showing the location of longitudinal profiles and stratigraphic sections at the western end of the coastal plain.

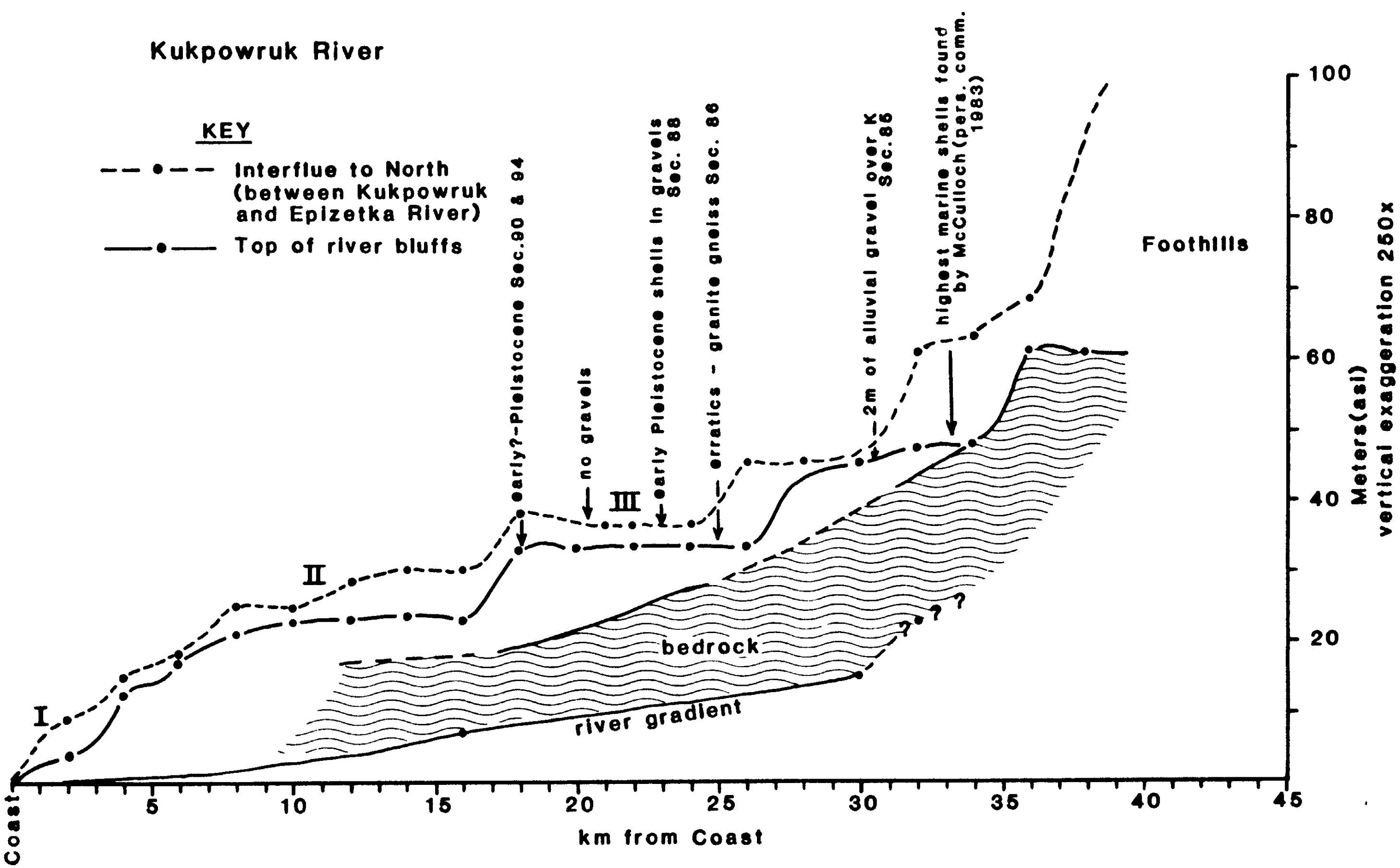


Figure 4.10. Longitudinal Profile of the Kupowruk River and adjacent interfluves.

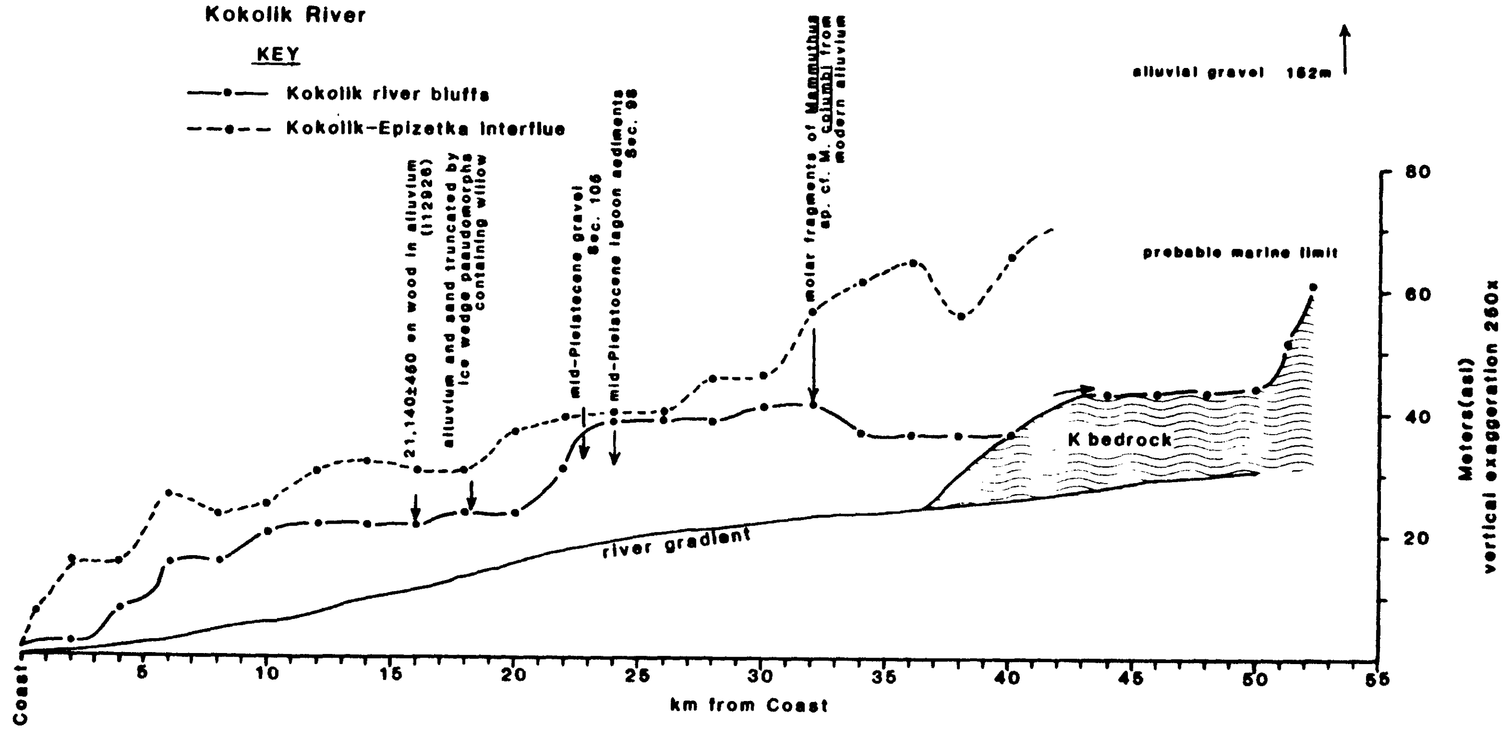


Figure 4.11. Longitudinal profile of the Kovolik River and adjacent interfluves.

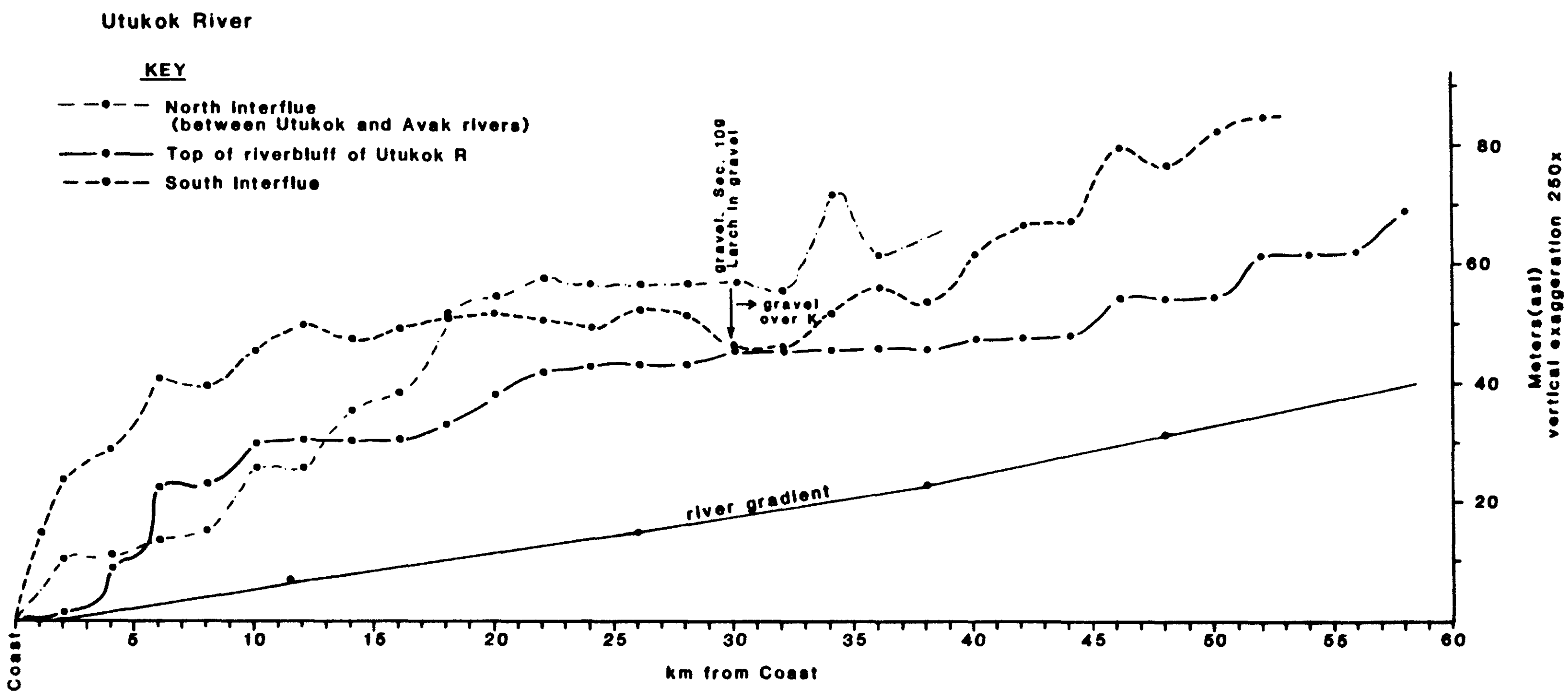


Figure 4.12. Longitudinal profile of the Utukok River and adjacent interfluves.

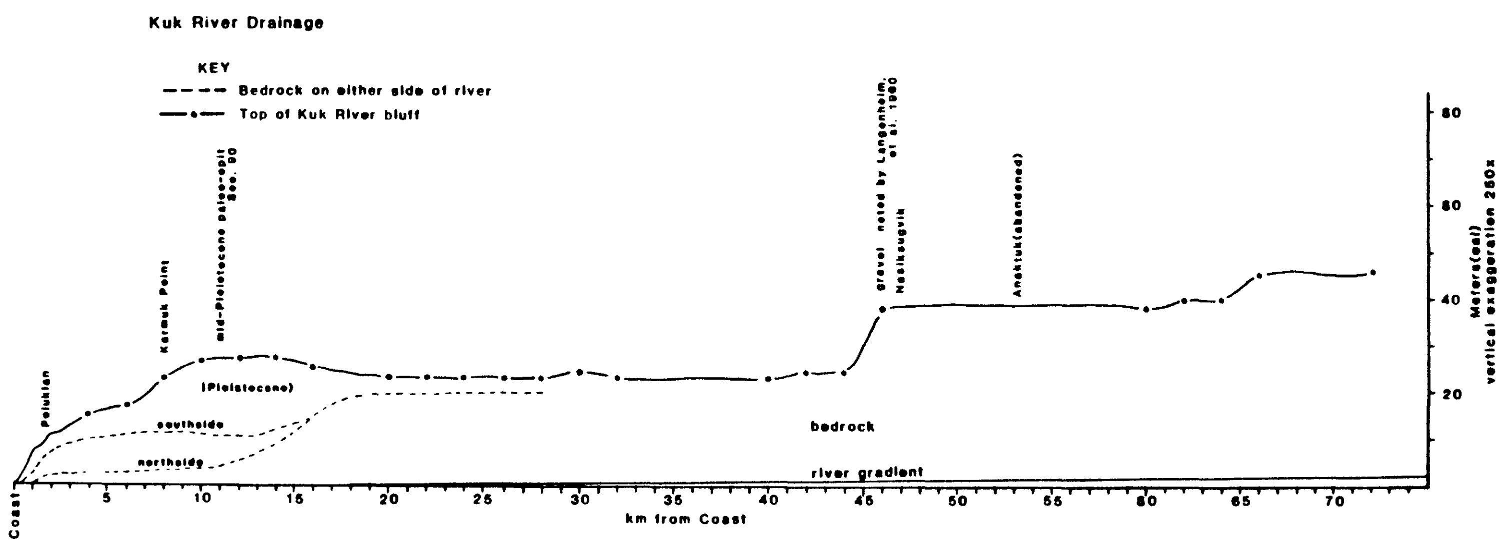


Figure 4.13. Longitudinal profile of the landscape in the lower Kuk River drainage.

broad contour interval on available maps (contour interval of 25 to 50 feet, equivalent to 8 to 15 m). Points between widely-spaced contours were extrapolated by averaging local spot elevations. It is appropriate to mention that all of the elevations quoted in this study, with the exception of measured stratigraphic sections, are map elevations from 1955 series U.S.G.S. topographic maps. These maps are photogrametric maps not published until after 1965; they were never field checked and are not entirely accurate. The same maps were used by Williams (1983a, 1983b) but were not available to McCulloch (1967) at the time of his study. In addition to stream dissection, solifluction and the local accumulation of eolian sand, variations in the volume of excess ice in the upper 6 to 8 m of permafrost (up to 70 - 80%) adds topographic variability to the ancient marine surfaces (Sellman et al., 1975). For example, the thawing of ground ice during the formation of thaw lakes commonly lowers the land surface by as much as 3 or 4 m, and thaw-lake basins as deep as 21 m have been reported in areas of ice-rich fine-grained silty-sediment (Williams and Yeend, 1979; Williams, 1983a).

Discussion of Profiles

The profiles outlined in Figures 4.10 - 4.13 generally suggest an irregular stair-step series of terraces from the highest features at roughly 60 - 70 m a.s.l. to the lowest surfaces between 10 and 15 m a.s.l.

Surfaces above 40 m a.s.l.

Marine sediments or even alluvium containing reworked marine shell fragments are rarely found higher than 40 m a.s.l. across the western coastal plain. D.S. McCulloch (pers. comm., 1983) reported that he found sparse fragments of marine shells in sediments below a surface as high as 48 m a.s.l. on the Kukpowruk River at latitude 69°23', T.1S.R.44W, Sec. 6, but the specimens were poorly preserved and not collected. Likewise, Warren Yeend (U.S.G.S., pers. comm. to J.R. Williams, Nov. 1982) reported that he observed valves of *Hiatella arctica* on river bars of the Topagoruk River (Fig. 1.1) well south of the northern escarpment of the upland silt that rises above 60 m a.s.l. Both occurrences suggest that older marine deposits may be present at higher elevations but are now obscured beneath the eolian upland silt.

Distinct terraces of marine origin are not apparent from topographic maps or the SLAR imagery above 40 m a.s.l., but, a gently sloping surface at about 45 m a.s.l. can be recognized in the longitudinal profiles drawn along the Kukpowruk, Kokolik, and Utukok Rivers. This surface abuts a gentle escarpment that rises to about 60 m a.s.l., particularly in the region of the Kukpowruk and Kokolik Rivers (Figs. 4.10 and 4.11). A similar escarpment marks the northern edge of the upland silt just east of the Meade River near the boundary between the Meade River and Lookout Ridge quadrangles (Fig. 4.7). As yet, there are no clear grounds to assume that either escarpment was carved by an early Pleistocene high sea stand, however this interpretation is favored at present.

W.A. Yeend (in Williams and others, 1977) identified what he thought were primary beach deposits (he mapped Qb) on upland surfaces in the foothills along the Kokolik and Utukok Rivers at an elevation of about 150 m. Gravel deposits on either side of the Kokolik River were examined in 1981 and shown to be 11 m thick where they rest directly on bedrock at one site (Sec. 103; Fig. 4.8). These sediments consist of coarse, rounded, cobbly gravel composed primarily of chert and sandstone. Rare clasts up to 20 cm in diameter are present. Contrary to Yeend's interpretation, the deposit represents very old alluvial gravel rather than beach gravel. This interpretation is based upon the coarseness of the deposit, the degree of rounding observed, and the distribution of the deposit. No marine fossils were found.

Surfaces between 20 and 40 m a.s.l.

The most convincing evidence for high marine terraces on the western coastal plain lies in the stratigraphy of patches of marine sediments exposed along the banks of the rivers draining the region. Corraborative evidence can be derived from the surface morphology of the Wainwright area southward to the edge of the coast near Noakok Pass, although intensively modified by eolian and periglacial activity.

Along the narrow interfluve between the Epizetka and Kukpowruk Rivers, the most southernly profile has the most strikingly terraced morphology, characterized by pronounced surfaces along the interfluve at 61, 46, 36, 30, 22 and 12 m a.s.l. (Fig. 4.10). At a few localities along the Kukpowruk River, bluffs on the right bank expose a thin cover of unconsolidated Pleistocene sediments that overlie gently folded Cretaceous sandstone and shale. At a point approximately 17 km from the coast (stratigraphic sections 90 and 94, Fig. 4.8) the 33 m surface is underlain by 18 m of unconsolidated sediment (Fig. 4.14). Overlying Cretaceous bedrock that rises about 11 m above the Kukpowruk River, the base of the unconsolidated section consists of 4-5 meters of interbedded medium to fine sand and peaty silt containing bits of detrital coal. In Sec. 90, (approximately 50 m upstream from Sec. 94) evenly disseminated clots of peat and several well-preserved fragments of willow were found at the base of this unit. At this same level marine shells including one complete specimen of Neptunea, were recovered in Sec. 94. These sediments are overlain by well-sorted and ripple-crossbedded, fossiliferous marine sand that ranges in thickness from 60 cm to 2 m. Contacts above and below the unit may be sharp or gradational. The top of the bluff section is composed of 3.5 to 4.5 m of bedded, cobbly, well-rounded gravel and sandy gravel containing only shell hash. The sequence is then capped by 30 cm of silt and peat. The upper half meter of gravel in Sec. 94 is infiltrated with silt and many of the cobbles and pebbles are coated with pedogenic carbonate. The shaft of a juvenile, proboscidian humerus (mammoth?) (81 Akb 596) was found buried in silt at the top of the

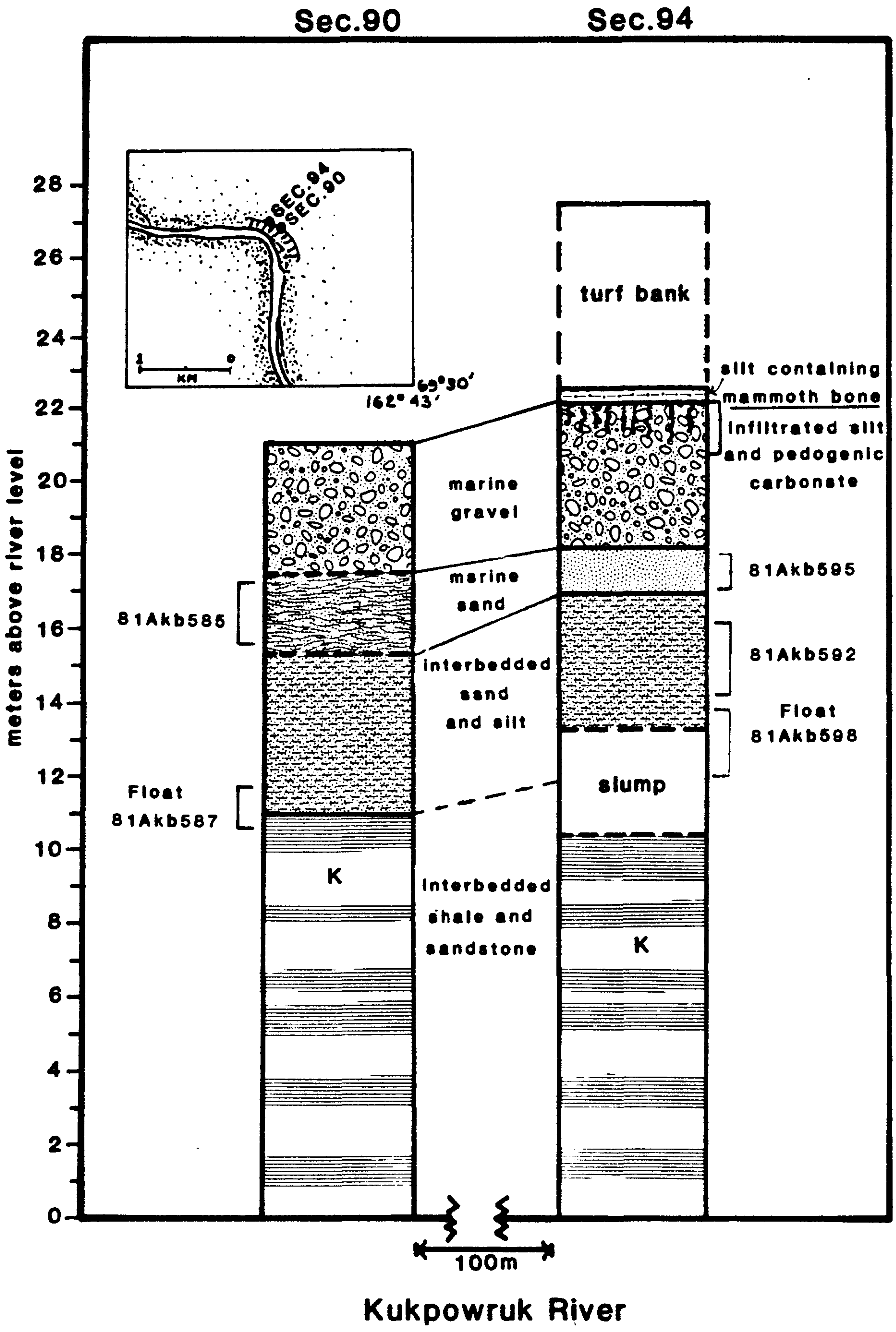


Figure 4.14. Stratigraphy of middle Pleistocene fossiliferous marine sediments on Kupowruk River. Mollusk samples are indicated by field numbers. Amino acid results are listed on Table 4.1.

gravel unit, several cervical vertebrae of Odobenus sp. (walrus) and rib and mandibular cetacean fragments (whale?) (81 Akb 588A) were recovered from the marine sand. Additional Odobenus sp. (walrus) vertebrae, presumably derived from the sand, were found littering the base of the section (81 AKB 588B).

Specimens of Hiatella arctica, Astarte sp., and Macoma sp. from the marine sand and basal interbedded peaty silt and sand were analyzed to determine the extent of isoleucine epimerization (Fig. 4.14 and Table 4.1). The quality of the results from all three genera were poor, leaving some uncertainty about the age of units. Ratios on Hiatella and Macoma from the fossiliferous sand in the middle of the section indicate that the deposit includes valves perhaps both of middle and early Pleistocene age (AAL-3439A 81 Akb 595 aminozone 3; AAL-3441A 81 Akb 585 aminozone 2). Two chalky fragments of Hiatella from the basal interbedded sand and silt gave ratios indicating a late Pliocene age (AAL-3440, 81 Akb 592, aminozone 4). Moreover, Hiatella and Astarte collected as float at the base of the section confirm that valves of all three ages are present (AAL-3437, 81 Akb 598, AAL-3438, 81 Akb 587). One possible interpretation is that the basal interbedded unit is of late Pliocene age (aminozone 4) and the fossiliferous sand and gravel units above are of middle Pleistocene age (aminozone 2). Approximately seven km upstream from this section on the same surface, however, (Sec. 88, Fig. 4.9) marine shells were also found in a poorly exposed section of cherty cobbly gravel lying directly on Cretaceous bedrock. The aIle/Ile ratios in Hiatella arctica from these beds indicate that the gravels at this site are of early Pleistocene age (aminozone 3). It is difficult to ascertain whether the deposit at sec. 88 is of primary marine origin or whether it is alluvium, and hence, whether the marine fossils are reworked. Because the site is positioned seaward of but near an escarpment leading to the next higher (46 m) surface, beach sediments contemporaneous with the 33-36 m surface might be expected here. An erratic angular boulder of granite gniess, 50 cm in length was found in slumped gravel, possibly alluvium, at section 86 which is still nearer to the escarpment.

Geomorphically, it seems most reasonable that the scarp just seaward of sections 90 and 94 (see Fig. 4.10) is a wave-cut scarp marking the inner limit of the middle Pleistocene (aminozone 2) transgression. At sections 90 and 94, accordingly, the marine sand and upper gravel might represent sediments deposited by an early Pleistocene transgression (aminozone 3) that overlie older marine deposits (aminozone 4). The scarp near section 86 most likely represents the inner limit of the early Pleistocene transgression. At present, this interpretation is favored until additional samples can be collected and analyzed to verify that middle Pleistocene (aminozone 2) shells are present in sec.s 90 and 94.

North of the Epizetka River, the escarpment inland from Sec. 88, above, can be traced along the southern and eastern shores of Tunusiktok and Pugutak lakes (Fig. 4.5 and Fig. 4.8 Sec. 113 and

Table 4.1 D-alloisoleucine/L-isoleucine Ratios on Mollusks from marine units exposed along the Kukpowruk River (Fig. 4.14).

Section (loc., Fig. 4.8)	Stratigraphic Position	Field No.	Lab. No.	Genus[2]	alle/Ile Free		Total	
88	Float from gravel	81Akb577	AA1-2831A	Ast	0.51	(2)	0.065	(2)
			B	Ha	0.50		0.088	
			C	Ha	0.50	(2)	0.095	(2)
			D	Ha	0.58		0.045	(2)
90	Float, base of section	81Akb587	AAL-3438A	Ast	0.56	(3)	0.098	(3)
			B	Ast	0.46	(2)	0.104	(3)
			C	Ha	0.48		0.093	(2)
			D	Ha	0.38		0.061	(2)
			E	Ha	0.51	(2)	0.109	(2)
90	Marine sand 15.5-17.5 m a.s.l.	81Akb585	AAL-3441A	Mac	0.42	(2)	0.093	(3)
			B	Ha	0.43	(2)	0.068	(2)
94	Float, base of section	81Akb598	AAL-3437D	Ha	0.45	(2)	0.079	(2)
			E	Ast	0.50	(2)	0.061	(2)
			F	Ha	0.46		0.054	(2)
94	Marine sand 17-18 m a.s.l.	81Akb595	AAL-3439A	Ha	0.53		0.108	
			B	?	----		0.059	(2)
			C	?	0.53		0.054	(2)
94	basal interbedded silt and sand	81Akb592	AAL-3440A	Ha	0.60		0.137	(2)
			B	Ha	0.52		0.131	

[1]Numbers in parentheses indicate that the alle/Ile ratio is a mean value of more than one analysis of the same sample.

[2]Ast, <u>Astarte</u> sp.; Ha, <u>Hiatella arctica</u>; Mac, <u>Macoma</u> sp.

114). At the eastern edge of both lakes well-rounded beach gravel and sand is exposed in slumped deposits that overlie Cretaceous bedrock cropping out at lake level. Numerous fragments of Astarte sp. and Neptunea were found as float on the slumped deposit. Three 30 cm diameter boulders of chert and the plate of a mammoth tooth were found on the lake beach. The alle/Ile ratios determined on the six Astarte from Sec. 113 and 114 at the edge of the lakes (Fig. 4.5 and 4.8) indicate that the shells belong to aminozone 3 and are largely of early Pleistocene age (Table 4.2). The free ratio of a few valves is somewhat lower than expected for early Pleistocene and may indicate that the collection includes some samples of middle Pleistocene age, however, this is speculative. It is my interpretation that the deposits at the base of the escarpment mark the highest strand of the early Pleistocene transgression.

Some of the sites along the Kukpowruk River were also visited by McCulloch (1967), who evidently found better exposures here and made several collections of marine fossils 18 years ago. His collections of marine mollusks along the Kukpowruk River above 30 m a.s.l. were assumed to be of one age, hence specimens are mentioned in his text only as an aggregate fauna of 48 species that include Neptunea heros mesleri and Astarte leffingwelli, two forms that are considered extinct.

Profiles along the Kokolik River and along the interfluve betweeen the Kokolik and Epizetka rivers (Fig. 4.11) are less clearly terraced than those along the Kukpowruk River and exposures in meander scars and river bluffs are largely composed of alluvium. Despite this contrast, sediments of middle Pleistocene age (aminozone 2) were found underlying a surface at a map elevation of approximately 38 m a.s.l. At river level at one site (roughly 18-20 m a.s.l.), a thin bed of middle Pleistocene lagoonal sediments was excavated overlying coarse cobbly alluvial gravel and buried by a sequence of coarse pebbly beach gravel (Sec. 105, Fig. 4.15). The top of the exposed section is about 12.5 m above river level, or approximately 28 m a.s.l. Sediments above this level are not exposed.

Approximately 1.7 km upriver of this site in the low bluff of a meander scar (Sec. 98, Fig. 4.15), several ice-wedge pseudomorphs are exposed in alluvial gravel, trucated by lagoonal sediments of middle Pleistocene age (aminozone 2, AAL-2988). (These are the oldest ice-wedge pseudomorphs documented thus far on the Arctic coastal plain). Paired valves of Macoma balthica Linnaeus with periostracum are abundant in the lagoonal sediments at both this site and sec. 105 along with small fish bones and abundant plant debris. The presence of Macoma balthica is notable because today it rarely occurs north of Kasegeluk Lagoon and generally inhabits brackish water near river mouths and lagoons.

Tentatively, I have interpreted the gravel immediately above the lagoonal sediments in Sec. 105 as of primary marine origin, possibly marking the inland extent of an offshore gravel

Table 4.2. D-alloisoleucine/L-isoleucine Ratios (Aminozone 3) in *Astartes* from marine gravel at Tunusiktok and Pugutak lakes (Fig. 4.5).

Section (loc., Fig. 4.4)	Field No.	Lab No.	aIle/Ile[1] Free	Total
113	81Akb642	AAL-2812	A 0.59	0.101
			B 0.45	0.076 (2)
			C 0.59	0.055 (2)
			D 0.55	0.086 (2)
114	81Akb643	AAL-3444	A 0.65	0.084 (2)
			B 0.62	0.092 (2)

[1]Number in parentheses indicates that the ratio represents the mean of more than one analysis of the same sample.

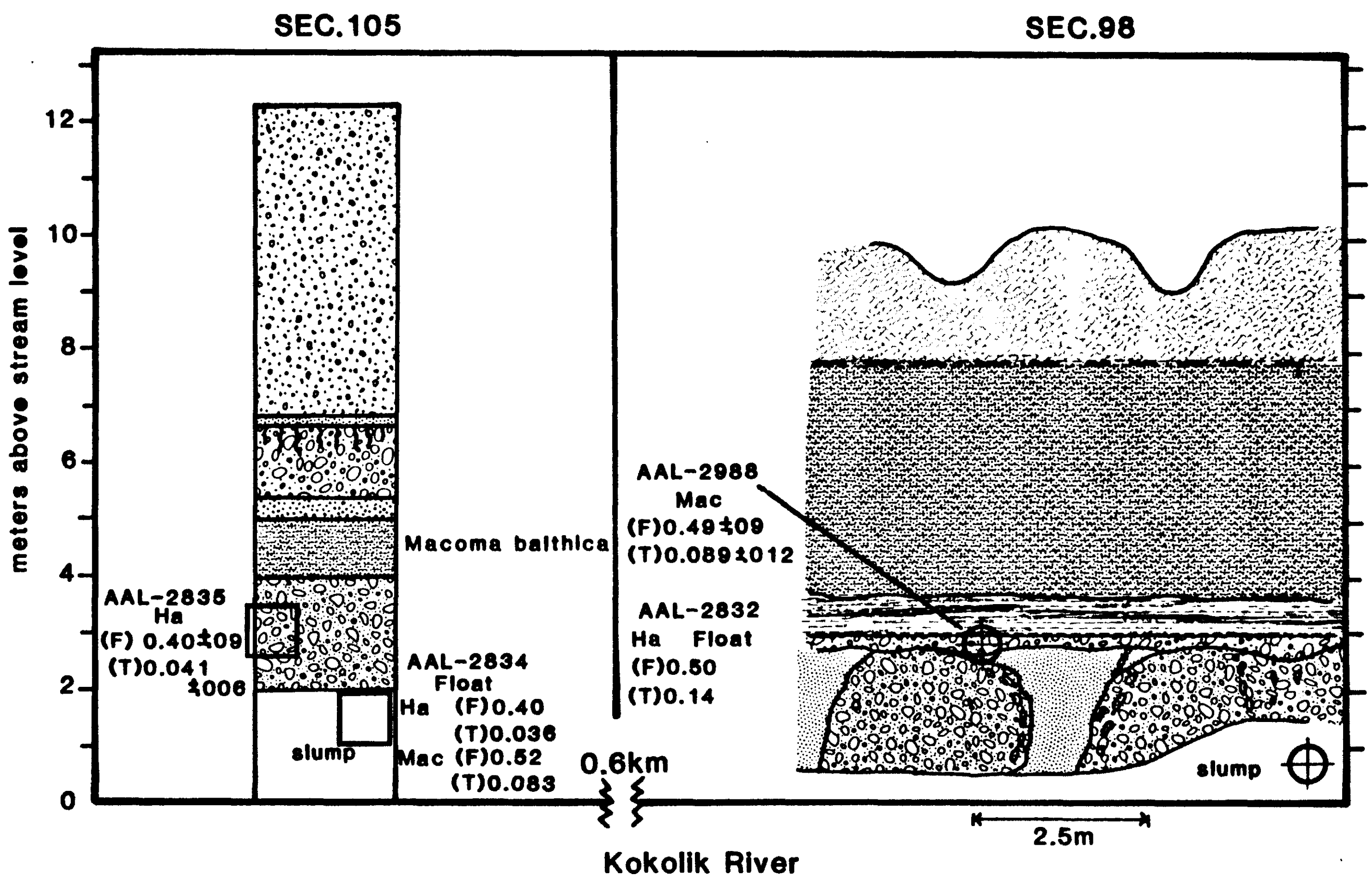


Figure 4.15. Stratigraphy of marine units on the Kokolik River. At Sec. 98, the ice wedge pseudomorphs are truncated by mid-Pleistocene lagoonal sediment containing Macoma balthica.

bar. A short-lived unconformity is suggested by the infiltration of organic material, including small plant fragments, into the upper 10 cm of the coarse gravel (at 22 m a.s.l., Sec. 105). This migrating gravel bar transgressed over the lagoonal deposits in Sec. 105 but did not reach Sec. 98. No other exposures of marine sediments could be found along the Kokolik River, however Sec. 98 must be very close to the position of the mid-Pleistocene shoreline. Fossiliferous gravel found lining a large thaw lake directly north of Sec. 105 is probably correlative with those exposed at the river and represent an extension of the middle Pleistocene gravel bar. No evidence remains to confidently trace this gravel to the Utukok River.

From the Utukok River to Pingorarok Hill at the northern end of Kasegaluk Lagoon, the edge of the coastal plain is characterized by a nearly continuous escarpment, the top of which averages about 43 m a.s.l. (141 ft.) from Sikolik Lake northward (see Fig. 4.3). At the lip of the escarpment a discontinuous ridge of beach gravel can be traced at a map elevation of about 33 - 40 m a.s.l. (100 to 130 ft). This gravel beach was initially mapped by McCulloch (1967) and later in detail by Williams (1983b) and I subsequently examined it during the field work (1981) for this study. The gravel deposits extend northward beyond Pingorarok Hill to where they terminate in the high bluffs just 2.5 km south of Karmuk Point on Kuk Inlet.

Two key stratigraphic sections bear evidence concerning the age of the beach gravel along the escarpment. At the north end of Kasegaluk Lagoon and 3.7 km north of Mitliktavik, a small gully has eroded through the escarpment exposing 2 ages of marine sediment (Sec. 112, Fig. 4.16). The lower portion of the exposure consists of contorted beds of interbedded gray silty sand and dark brown silt that grades upward into a mottled, medium, yellow sand. A single valve of <u>Cardita</u> was retrieved from these sediments. The top of this unit is capped by 8 cm of clayey silt. Overlying these beds are 6 m of fossilferous, horizontally bedded coarse sand and gravel that grades upward into well-rounded, open-matrix pea gravel. The top of the section is capped by eolian sand. Partially broken and whole valves of <u>Neptunea</u> are abundant in the coarse gravels along with sparse shells of <u>Astarte</u> sp. The largest collections of mollusks were made from the gully floor.

Alle/Ile ratios from 4 valves of <u>Astarte</u> sp. and one hinge fragment of <u>Mya</u> sp. collected as float at the base of the section suggest that mollusks of both early (aminozone 3) and middle Pleistocene age (aminozone 2) are present in the section (81 Akb 640, AAL-2811 and AAL-3412). One worn but whole valve of <u>Astarte</u> sp. was recovered in section from the lower portion of the gravel and the ratios from it are interpreted as middle Pleistocene in age (aminozone 2) (81 Akb 656, U.S.G.S. loc. M 8156, AAL-3413). Curiously, however, the mollusk fauna in the lower part of the gravel includes specimens of <u>N.</u> (<u>N.</u>) <u>lyrata leffingwelli</u> (Dall), an extinct gastropod usually restricted to early Pleistocene units along Skull Cliff. Likewise, <u>Natica</u> (<u>Tectonatica</u>) <u>janthostoma</u>

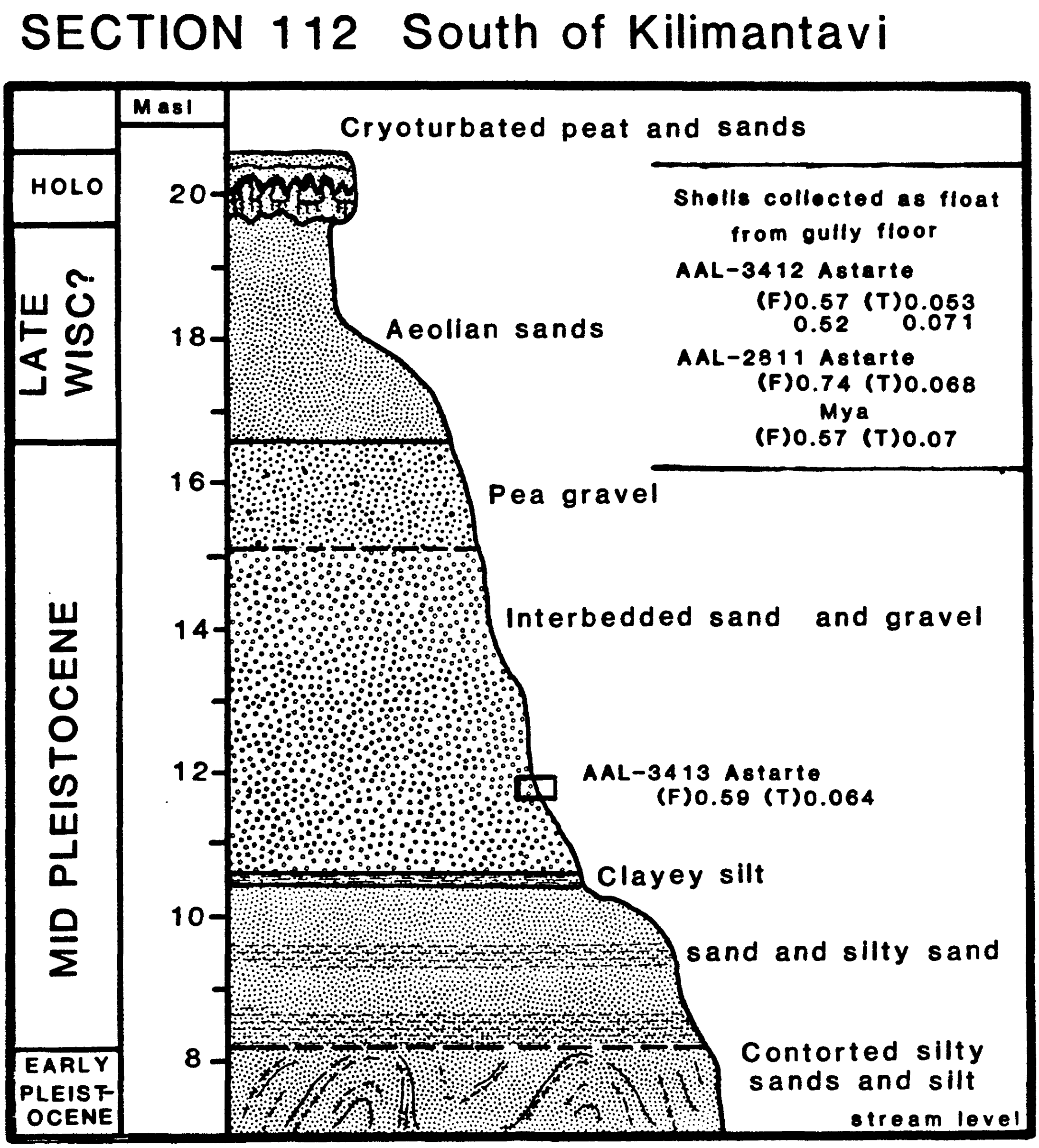


Figure 4.16. Stratigraphy in escarpment facing the Chukchi Sea. Two marine units of different age are exposed here.

Deshayes (U.S.G.S. loc. 7308), an extralimial species from the northwest Pacific and usually found in older units, was recovered by Williams (1983b) from gravel along this portion of the escarpment.

My interpretation of the section is that the contorted beds at the base are early Pleistocene in age, and the gravel sequence beneath the eolian sand is middle Pleistocene in age. Mollusks from the older unit have been reworked into the younger beach gravels, as indicated by the mixed faunal assemblage.

The other critical section is exposed in the high bluffs facing Kuk Inlet 2.5 km south of Karmuk Point (Fig. 4.17). Here, interbedded and crossbedded coaly yellow sand and gray clayey silt are overlain by coarse, well-bedded, open-matrix gravel. While both units contain an abundance of marine mollusks, the sediments are barren of microfossils. Pieces of willow (Salix) were recovered from the lower unit (wood 81 Akb 545, identified by U.S.F.S. Wood Products Laboratory).

AIle/Ile ratios on 8 valves of Hiatella arctica from the lower unit (81 Akb 545, AAL-2552 on 81 prep; 81 Akb 559, AAl-2810 on 76b prep) and 8 valves of H. arctica from the upper gravel (81 Akb 546, AAL-2555 on 81 prep.; 81 Akb 554, AAL-2809 on 76b prep) suggest that the entire section above the Cretaceous bedrock (or Miocene clay?) is middle Pleistocene (aminozone 2) in age (Table 4.3). The section is interpreted to represent a shallow-water marine facies truncated by migration of an emergent spit. The highest strand of the transgression in this area probably lies along the base of a prominent escarpment located 35 km southeast of Karmuk Point near Nasiksugvik on the Kuk River (see Fig. 4.13). Along the top of this escarpment, Langenheim and others (1960) observed "a mesalike cap of gravel" that probably represents much older fluvial or marine gravel. These deposits were not visited during this study.

The morphology and stratigraphy of the marine sequence between the Utukok and Kuk Rivers suggests that the highest strand of the middle Pleistocene transgression abutted the escarpment bordering the coastal plain from Sikolik Lake to Pingoruark Hill (Fig. 4.18). This same strand then appears to curve southeastward and then northeastward continuing around to cross the Kuk River near Nasikugvik, thus forming a large bay over the lower reaches of the Kuk River. The western edge of the bay was partitioned from the Chukchi Sea at that time by a spit and discontinuous barrier chain that terminated near present day Karmuk Point. Along the top of this ancient spit at Sec. 124 (Fig. 4.4) a cherty conglomeratic boulder 30 cm in diameter was found and a number of granitic erratics were observed along the shores of the Kuk Inlet.

It is somewhat difficult to place into stratigraphic context the gravel that can be found perched at the lip of the Icy Cape escarpment between Sokolik Lake and Pingorarok Hill. It appears that at least a portion of this gravel up to ca.

SECTION 80 Karmuk Point

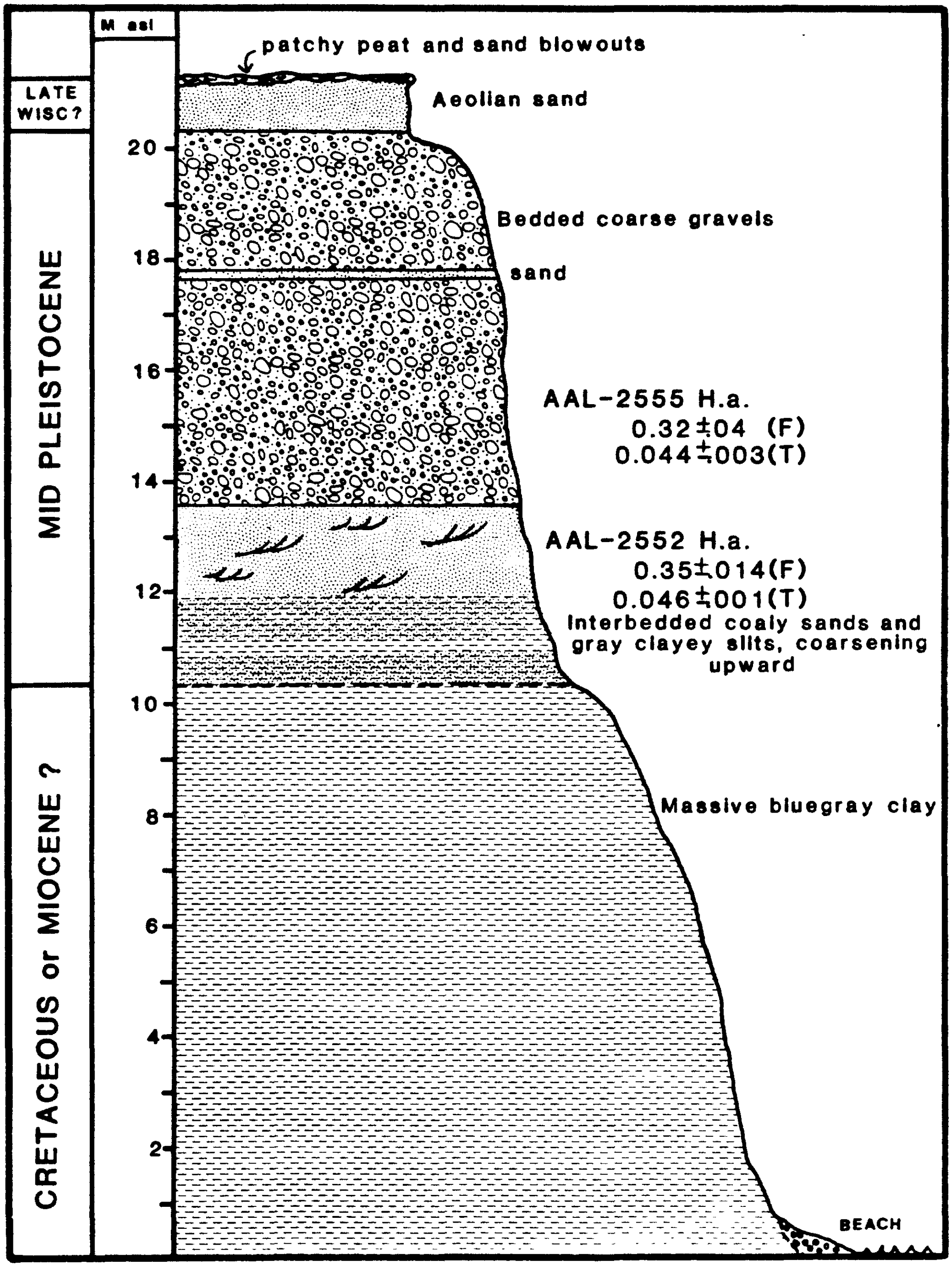


Figure 4.17. Stratigraphy near Karmuk Point. The aIle/Ile ratios indicate that the marine unit is of one age.

Table 4.3. D-alloisoleucine/L-isoleucine ratios (Aminozone 2) from Hiatella in marine sediments at Section 80, near Karmuk Point (Fig. 4.17).

Field No.	M a.s.l.	Lab No.		aIle/Ile Free	Total
			76b PREPARATION		
81Akb554	14-18	AAL-2809	A	0.42	0.027
			B	0.45	0.031
			C	0.46	0.038
			D	0.49	0.037
			E	0.43	0.037
81Akb559	10.5-13	AAL-2810	A	0.40	0.033
			B	0.40	0.031
			C	0.37	0.035
			D	0.42	0.038
			E	0.44	0.057
			81 PREPARATION		
81Akb546	14-18	AAL-2555	A	0.32	0.070
			B	0.37	0.058
			C	0.29	0.043
81Akb545	10.5-13	AAL-2552	A	0.36	(missing)
			B	0.35	0.047
			C	0.33	0.046

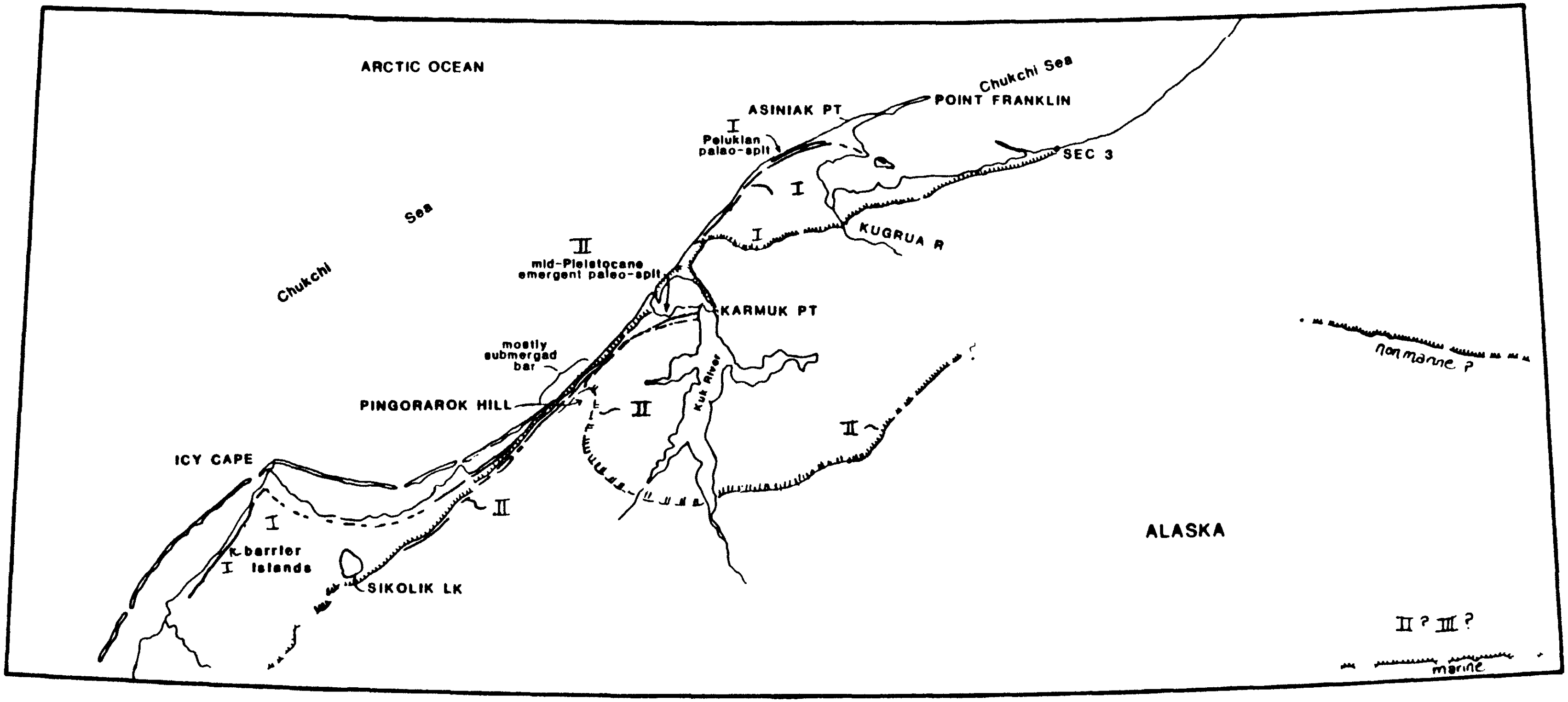


Figure 4.18. Summary map of paleoshorelines between Skull Cliff and Icy Cape. These shorelines are similar to those of Williams (1983a, 1983b) but there ages are better contained. I, includes beach, barrier bars, and escarpments deposited or cut by the marine transgression of Aminozone 1; II, indicates features deposited or cut by the transgression of Aminozone 2. III, escarpment cut during aminozone II or III.

16 m a.s.l. along the slope of the escarpment includes shells of middle Pleistocene age (aminozone 2). The gravel present along much of the top of the escarpment, however, must be of greater age and may, perhaps, represent coarse material (ancient barrier chain?) stranded following the early Pleistocene (aminozone 3) transgression. Additional stratigraphic study along the promontory should clarify this relationship.

The position of the middle Pleistocene strand east of the Kuk River is not known. In all probability it lies just north of the escarpment that continues northeastward from the Kuk River toward the Meade River. Alle/Ile ratios from 6 shells of Hiatella arctica collected in sandy alluvium at Sec. 99 about 32 km south of Atkasook (Fig. 4.20) indicate that most of the reworked valves are of early Pleistocene age but the collection includes one valve of middle Pleistocene age (Table 4.4, 81 Akb 611, AAl-2551 on 81 prep.). Analysis of three valves of H. arctica collected nearby in 1978 by J.R. Williams, U.S.G.S., also yielded mixed age assemblages with one of the three valves having alle/Ile ratios suggestive of a middle Pleistocene age (78 AWL 256, U.S.G.S. loc. 7309, AAL-1376 on "Norway prep"). These collections suggest that during the middle Pleistocene high sea stand, a large bay occupied the valley of Meade River. The head of bay may have abutted the escarpment at the north edge of the Upland Silt (Fig. 4.18 and 4.20).

Surfaces below 20 m a.s.l.

The lowest and youngest shoreline along the western coastal plain is represented by beach, shoreface and lagoonal sediments that reach heights between 10 and 13 m a.s.l. Exposures through these shoreline features were studied only from Wainwright Inlet to Barrow; south of Wainwright shorelines were traced from airphotos and the SLAR imagery. From the amino acid analyses of numerous marine mollusks (aminozone 1), radiocarbon analyses of driftwood enclosed in the beach facies (Chapter VII), and the consistent height of this feature along the entire coast of the Chukchi Sea, the shoreline is interpreted to represent the last interglacial high sea stand, the Pelukian transgression, which took place approximately 125,000 years ago (Hopkins, 1967a; McCulloch, 1967; Williams, 1983a, 1983b).

McCulloch (1967) mapped the Pelukian shoreline south of the Kuk River as beach facies and nearshore marine sediments that lie seaward of a low escarpment thought to represent a wave-cut cliff. The shoreline was placed at the bottom of a break in slope at about 13 m a.s.l. (Fig. 4.2). Subsequently, Williams (1983b) mapped the shoreline in the same approximate position (Fig. 4.3). Low ridges composed of marine sediments of last interglacial age were examined near Nokotlek Point by Hopkins and assistants in 1978; they interpreted the ridges to be barrier bars of Pelukian (Sangamon) age and the low areas separating the ridges from a bedrock escarpment to be an ancient lagoon analagous to present day Kasegaluk Lagoon. I have identified similar ridges between the Utukok and Kukpowruk rivers on the SLAR imagery and have also

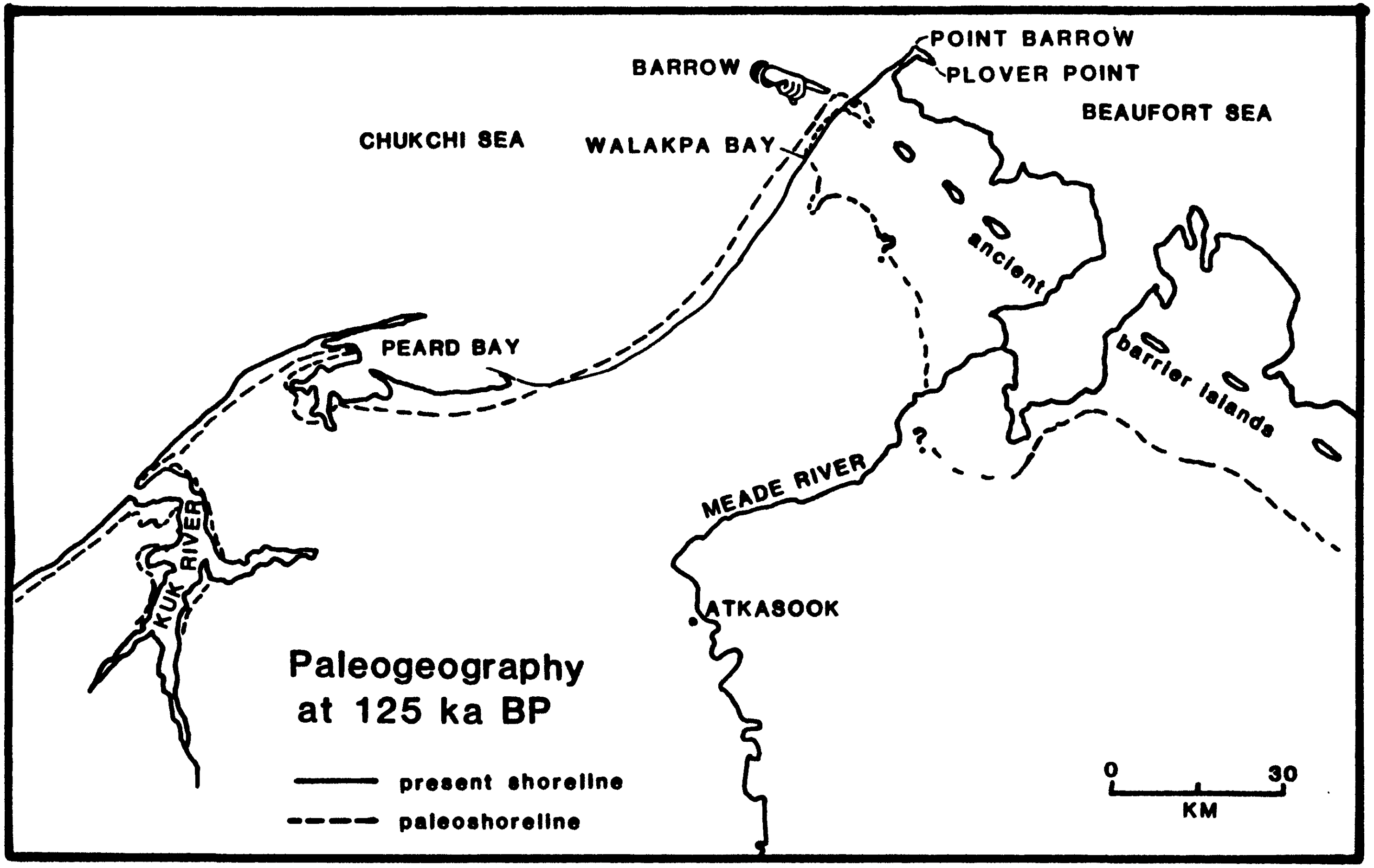


Figure 4.20. Reconstruction of the last interglacial shoreline, based upon the spatial distribution of contemporaneous sediments.

Table 4.4. D-alloisoleucine/L-isoleucine ratios on *Hiatella arctica* collected from exposures of sandy alluvium along the Meade River, 32 km south of Atkasook.

Section	Field No.	Lab No.		aIle/Ile Free	Total
99	81Akb611	AAL-2551 (81 prep)	A	0.56	0.071
			B	0.65	0.183
			C	0.61	0.093
			D	0.63	0.131
			E	0.65	0.150
			F	0.60	0.138
J. Williams USGS	71AWL256 (M7309)	AAL-1376 ("Norway" prep)	A	0.67	0.107
			B	0.44	0.033
			C	0.57	0.072

interpreted them as a system of offshore barrier bars (Fig. 4.2 and Fig. 4.18 and 4.8).

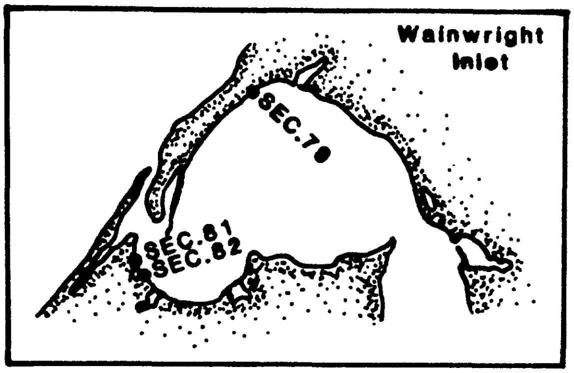
Today, Wainwright Inlet is partioned from the Chukchi Sea by a baymouth bar that was active during the last interglacial. This former beach system is continuous along the coast north and south of Wainwright reaching a height of about 10 m a.s.l. North of Wainwright, the ancient strand can be traced continuously to Peard Bay where it forms an ancient spit extending toward Asiniak Pt., while the low escarpment just landward from Peard Bay formed a wave-cut cliff at the inner edge of the bay (Fig. 4.18). No true beach deposits dating from the last interglacial are exposed in the sea bluffs bordering Peard Bay or just northeast of the Bay. Rather, much of the landscape in that area lies at about 5 - 6 m a.s.l. and consists of nearshore marine sands that appear graded to the last interglacial strand or were deposited during a later high sea stand coincident with the Flaxman formation in the Beaufort Sea. Shells of Astarte sp. from Sec. 3 (Plate I, and Fig. 4.18) on this surface yielded a radiocarbon date of 27,510 ± 500 years B.P. (DIC-2568). In addition, Williams (1983b) reported the occurrence of Macoma balthica, a low salinity mollusk, in marine sand and gravel near the former mouth of the Kogrua River estuary along with logs of driftwood dated as > 40 ka old (I-10,272).

Pelukian sediments are best exposed in bluffs facing Wainwright Inlet adjacent to the village access road and in the southwest corner of the Inlet below triangulation station "Inlet" where the top of the marine deposits is about 10 m a.s.l. At both sites the sediments are richly fossiliferous and consist of thick sequences of well-rounded and cross-bedded coarse gravel and sand (Fig. 4.19; Sec. 79, 81, and 82). Near the access road at Sec. 79 concentrations of twigs, some bird bones, and organic debris occur interbedded with the coarse sediments.

Along Skull Cliff from Peard Bay to a point just south of Walakpa Bay, no remains of the Pelukian shoreline are preserved. This portion of the coastal plain lies well above 10 m a.s.l. and is underlain by older parts of the Gubik Formation.

Pelukian sediments are present in measured sections up to ca. 10 m a.s.l. in the bluffs from Walakpa Bay to Barrow village. As illustrated in Chapter II, most sections are overlain by at least 1 - 4 m of eolian sand, thaw lake sediments, or in situ peat so that the top of beach ridges are expressed at a map elevation between 11 and 15 m a.s.l. (36 to 50 ft. a.sl.).

Approximately 2.6 km south of Walakpa Bay, crossbedded, sequences of fossiliferous coarse gravel and sand of last interglacial age terminate and lap onto older and finer-grained early Pleistocene marine sediments (see Chapter II and III; Plate I). Sedimentologically, these gravel sequences are similar to the extensive sections of gravel and sand exposed near Barrow. In contrast, however, are the bluff exposures along the intervening stretch of coast composed of early and middle Pleistocene sediments



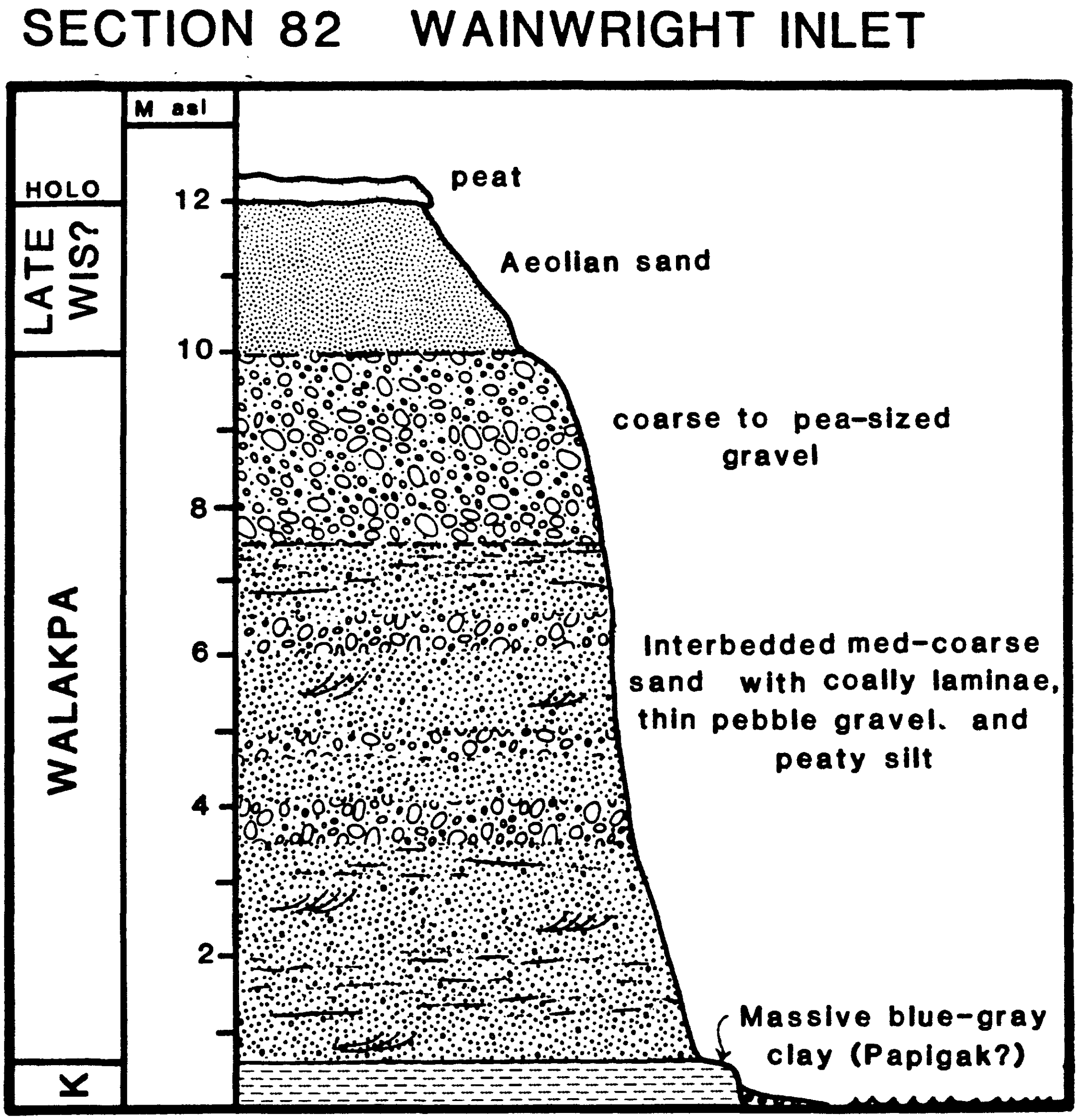


Figure 4.19. Stratigraphy of last interglacial beach deposits in Wainwright Inlet. Section 82 is representative of the stratigraphy at the three sites shown above.

disconformably overlain by a thinner cover of lagoonal and shoreface Pelukian sediments (Chapter II).

The concentrations of gravel at Barrow and at Walakpa Bay may represent beach-barrier bar complexes prograded to the 10 - 12 m a.s.l. high sea stand. The shoreline at that time may have resembled the modern spit complex between Barrow village and Point Barrow to the north, but the entire coast was evidently displaced westward (Fig. 4.20). The lack of middle Pleistocene sediments in many of the sections from Walakpa Bay to Barrow indicates a nearly complete removal by shoreface erosion of such sediments during the Pelukian transgression. Reworked mollusks from older deposits are only rarely found in sediments of the Pelukian age. Progradation of the beach and barrier bar complex was probably intiated during the culimination of the Pelukian sea level rise at which time the extensive gravel complex at Walakpa Bay was deposited. Continued littoral drift of sand and gravel must have then been sufficient to deposit the gravel complex beneath the Barrow airfield and exposed in the gravel pits to the south, eventually forming an eastward trending cuspate spit similar to the modern Point Barrow and Plover Point. The complex forms the westward end of an ancient system of barrier bars that extended eastward from Barrow to the Kogru River (Williams and others, 1977; Carter and Robinson, 1980).

It seems compelling to place the inner strand of the sea at the southern edge of the last interglacial sediments south of Walakpa Bay, although there is no clear morphological evidence to suggest that the sea terminated at that point. Much of the present landscape between Walakpa Bay and Nulavik, 17 km to the south, averages 10 - 13 m a.s.l. -- a figure that includes a significant thickness of Holocene peat and sand. Hence, most of this area probably lay less than 1 m a.s.l. during the Pelukian sea level maximum. The lack of a prominent shoreline here might be expected, considering the form of the modern mainland coast of the Beaufort Sea. Behind the barrier bars of the Beaufort Sea, the coast is intricately embayed and the coastal bluffs are generally less than 2 m high. When sea level falls, the present shoreline will not be a conspicuous feature. The Pelukian mainland shoreline of the Beaufort Sea coast would have been similar and will now have been obscured even more by thaw lake development, wind erosion and deposition, and burial beneath the large alluvial fans of the Meade, Topagaurvk, Chipp, and Ikpikpuk Rivers. The mainland shoreline of the last interglacial transgression is therefore interpreted as lying somewhere near the southern edge of the gravel exposed near Walakpa Bay.

Surfaces below 10 m a.s.l.

Seaward of the 10 m gravel surface at Barrow is a lower surface of marine deposition graded to a sea level $\leq$ 7 m a.s.l. (Fig. 4.21). Sellman and Brown (1973) provide the most thorough discussion of the stratigraphy of this surface based upon a transect of cores and auger holes in the area. They concluded that this pronounced beach ridge system was constructed 31 to 37 ka ago

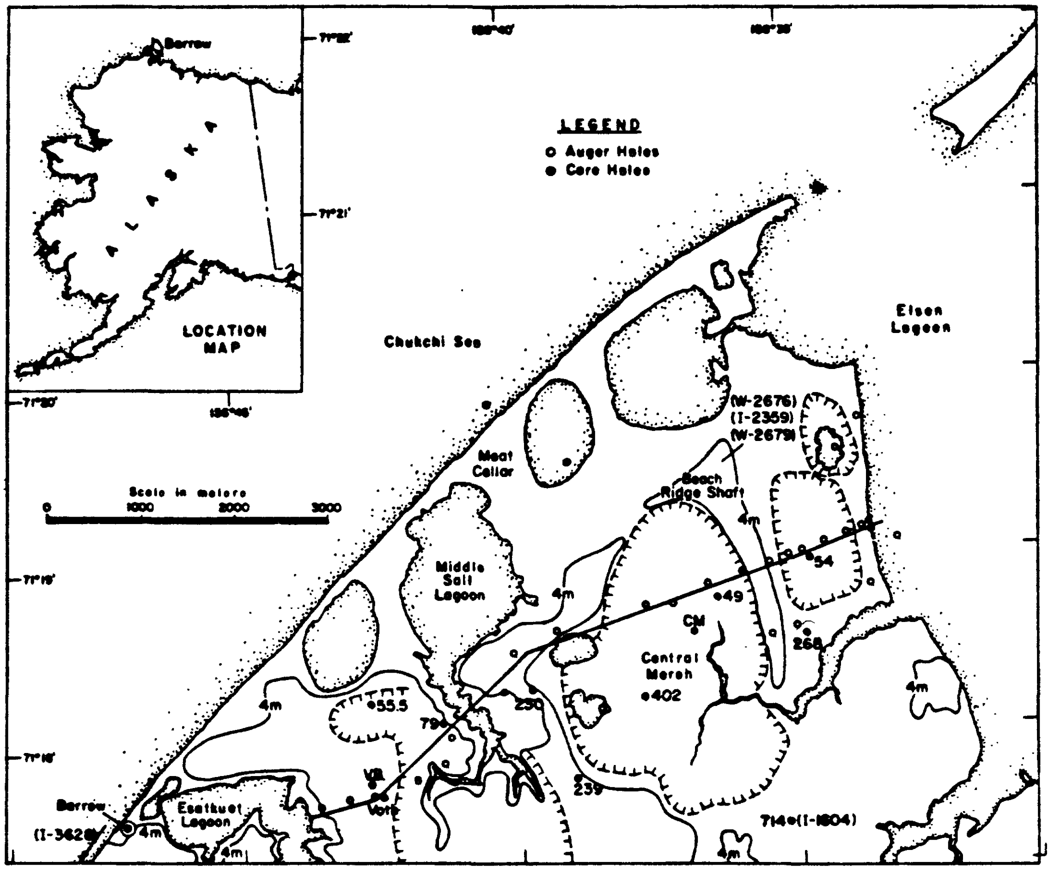


Figure 4.21. Map showing the prominent beach ridge seaward the last interglacial shoreline, as mapped by Sellman and Brown (1973). Although the deposit yielded radiocarbon dates between 31 and 37 ka B.P., the feature may have been deposited as eaarly as 80-100 ka B.P.

by a mid-Wisconsin interstadial high sea stand, based upon 3 radiocarbon assays on buried organics in the deposits. Their conclusion assumes that the mid-Wisconsin high sea stand reached several meters below present and that the Barrow area has subsequently undergone considerable uplift.

The nearly consistent height of the last interglacial Pelukian shoreline at ca. 8 m a.s.l. near Teshekpuk Lake (Carter and Robinson, 1980) to ca. 10 m a.s.l. at Barrow and 10 m a.s.l. at Wainwright indicates that little differential uplift has occurred on the coastal plain during the last 125 ka. Although subtle uplift of less than 1-2 m may have occurred to sediments that overlie the Barrow Arch, it cannot be definitively documented with the available exposures. Moreover, it is difficult to establish which portion of the beach facies is preserved at the top of the marine section and how much of its height is influenced by the presence of excess interstitial ice. Nevertheless, its height close to the mean world sea level value of ca. 7 m a.s.l. for the last interglacial high sea stand must imply that the younger beach ridge system at Barrow was deposited during a post-last-interglacial episode when sea level was again a few meters higher than at present.

Whether these deposits represent a mid-Wisconsin (35,000 ± 10,000 years B.P.) or a late interglacial (≈ 80 to 100 ka B.P.) transgression is not clear, due, in part, to the confusion created by radiocarbon dates on bone, shells, and disseminated organics between 22 and 35 ka B.P. from the deposits (Sellman and Brown, 1973; L.D. Carter, person. comm. of unpubl. dates, Jan. 1983). Although there is a great deal of conflicting evidence concerning the absolute height of the mid-Wisconsin high sea stand (see Thom, 1973, for review) most chronologies around the world agree that sea level was never higher than present. Hopkins (1982) argues that these sediments are correlative with deposits recognized along the Beaufort Sea Coast as the Flaxman Formation and were deposited by a transgression that took place following an abortive, high latitude glacial event during the latter portion of $\delta^{18}O$ stage 5. His interpretation concurs at least with global sea level chronologies that postulate post-last interglacial high sea stands at or above present only during $\delta^{18}O$ stages 5a and 5c (Shackleton and Opdyke, 1973; Bloom et al., 1974; Aharon, 1983). Hopkins' interpretation is favored by this study.

There are no shoreline deposits providing clear evidence of post-Pelukian high sea levels anywhere along the Chukchi sea coast south of Barrow. An erosional bench roughly 6 m a.s.l. observed along the shores of many of the larger bays and inlets (e.g., Walakpa Bay) may have formed during the "Flaxman transgression," but this can only be inferred. Similarly, the low-lying surfaces of marine sand previously described (p. 156) bordering Peard Bay may be contemporaneous. The very limited land area lower than 8 m a.s.l. along the Chukchi Sea coast implies that the shoreline during the Flaxman transgression must have been seaward of the present coast. Any sediments that may have been deposited offshore appear to have been eroded and reworked by the Holocene transgression (L. Phillips, U.S.G.S., written comm., Dec. 1982).

Summary and Discussion of Coastal Plain Neotectonics

The landscape of the western Arctic Coastal Plain contains the stratigraphic and subtle geomorphic evidence to document the approximate position of shorelines representing only four of the six transgressions recognized in the stratigraphic sequence. The recognized shorelines represent an early Pleistocene transgression (aminozone 3), a middle Pleistocene transgression (aminozone 2), the Pelukian or last interglacial transgression (aminozone 1), and a late Pleistocene transgression (Flaxman formation) that probably took place during the latter portion of oxygen isotope stage 5 (5a?). The oldest of these transgressions extended inland to at least 36 m a.s.l. in the vicinity of the Kukpowruk River and slightly higher (> 40 m a.s.l.) along the Icy Cape Promontory and the upper Kuk River embayment. The middle Pleistocene shoreline lies at about 22 m a.s.l. along the Point Lay embayment and Icy Cape Promontory, and greater than 20 m near Karmuk Point and along the Kuk River embayment (Fig. 4.22). Inland from Skull Cliff, the shoreline must lie somewhat greater than 18 m a.s.l. to have deposited shallow marine sediments in that area. Both of these transgressions flooded much of the western coastal plain whereas, the Pelukian transgression reached a height of only 10 m a.s.l. overwhelming only the margins of the region. The youngest transgression is represented by beach deposits up to 7 m a.s.l. in the Barrow area.

The position of strandlines occupied during older, late Pliocene transgressions can not be determined from the available stratigraphic data, or inferred from any clear geomorphic criteria. Based upon ratios from float and redeposited marine shells typical of aminozone 3, however, the sea is known to have transgressed to a position significantly higher than 40 m a.s.l. at some time during the early Pleistocene.

The miscorrelation of presumed marine surfaces of different elevation led O'Sullivan (1961) and Black (1964) to propose dramatic neotectonic warping of the coastal plain. In particular, they postulated that the eastern portion of the coastal plain had been uplifted some 90 to 180 m relative to similar surfaces to the west facing the Chukchi Sea. The mapping of strandlines across the coastal plain west of the Meade River (Fig. 4.8 and 4.18) indicates that the shorelines of middle Pleistocene age and younger are not dramatically warped. Subtle, tectonic movements of less than 10 m are suggested, however, (1) in the area of Skull Cliff and (2) the Point Lay Embayment, south of Icy Cape between the Utukok and Kukpowruk Rivers, and (3) between Icy Cape and Point Franklin.

Gentle uplift of the coast along Skull Cliff is suggested by the proximity of an undulating Cretaceous bedrock surface that is infilled with Miocene marine clays up to ca. 8 m a.s.l. (See

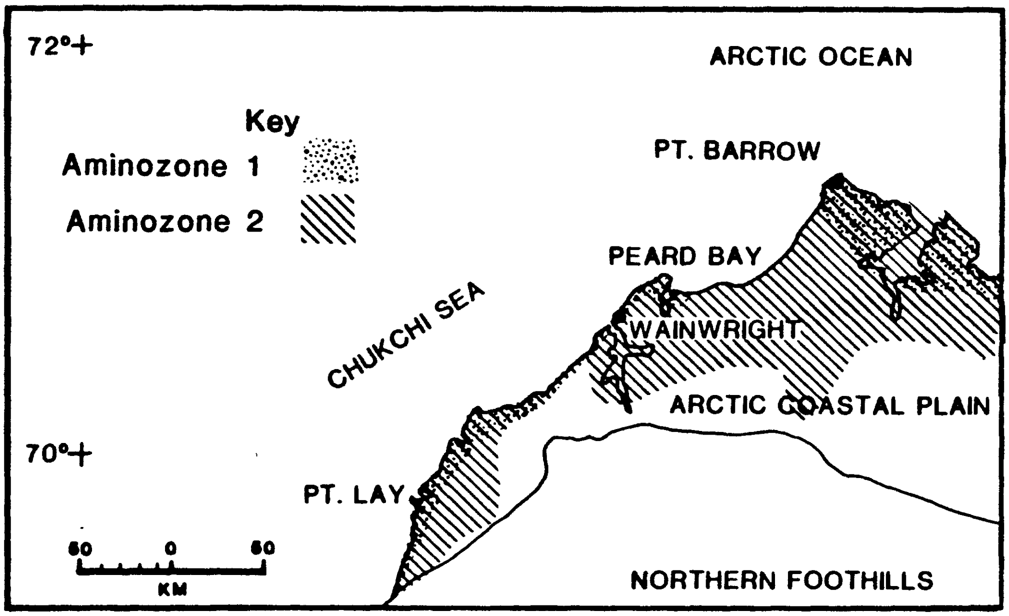


Figure 4.22. Paleogeography of the last interglacial (Aminozone 1) and mid-Pleistocene (Aminozone 2) high sea stands.

Chapter II; Plate I and II). These sediments are disconformably overlain by at least 2, but commonly 3 superimposed marine units of the Gubik Formation. Evidence that uplift has occurred within the time-frame of the Quaternary, however, is not directly indicated by the stratigraphy and spatial distribution of the marine units, themselves. The top of the highest middle Pleistocene age marine units in Skull Cliff lie at ca. 17 m a.s.l., roughly 5 m below the middle Pleistocene strandline found further inland. This is, notably, the only section of coast where bedrock is exposed above sea level.

Corroborative evidence of uplift is provided by the fact that up to within a few kilometers of the coast, all of the surface drainage between Walakpa and Peard Bay is directed northeastward away from the coast and toward the Beaufort Sea. In addition, local structural features, such as those indicated by the removal of Papigak clay between sections 16 and 25 and between 21 and 64 (Plate I), suggest differential uplift. The warping near section 20 directly coincides with the oil seep mapped at Skull Cliff.

The observed local structural features are probably related to slow regional uplift in this area. From Skull Cliff to Point Barrow the area has been uplifted over the Barrow arch almost continuously since the late Cretaceous resulting in the removal of over 1000 m of section (Ken Bird, U.S.G.S., pers. comm. June 1983). Movement of this magnitude is indicated by changes in the degree of compaction of the bedrock from bore holes that outline an anticlinal axis of uplift centered over the Skull Cliff area (Ervin, 1982; Arthur Grantz, U.S.G.S., written comm. to D.M. Hopkins, Dec., 1982) to the southwest of the Barrow arch.

Subtle tectonic movements are also evident in the Point Lay embayment in the area south of Icy Cape between the Utukok and Kukpowruk Rivers. Here, the outlines of the middle and early Pleistocene strands form two concentric embayments that align with the Hanna structural trough mapped offshore on the Chukchi shelf (Grantz and others, 1982; Fig. 4.23). The Pelukian shoreline distinctively cuts straight across these embayments.

Uplift along the Icy Cape Promontory and portions of the Kuk River embayment is suggested by the northward displacement of the coastal inflection and spit complex from Kuk Inlet toward Peard Bay. Following the early Pleistocene, the spit complex has migrated from near Karmuk Point during the middle Pleistocene, to Asiniak Point during the present interglacial. The uplift responsible for this displacement is difficult to quantify, however, it must be less than 1-2 m $100{,}000^{-1}$ years because the Pelukian shoreline is not noticeably tilted. Long-term uplift rates of less than 1 mm 1000^{-1} years may characterize this area of the coastal plain.

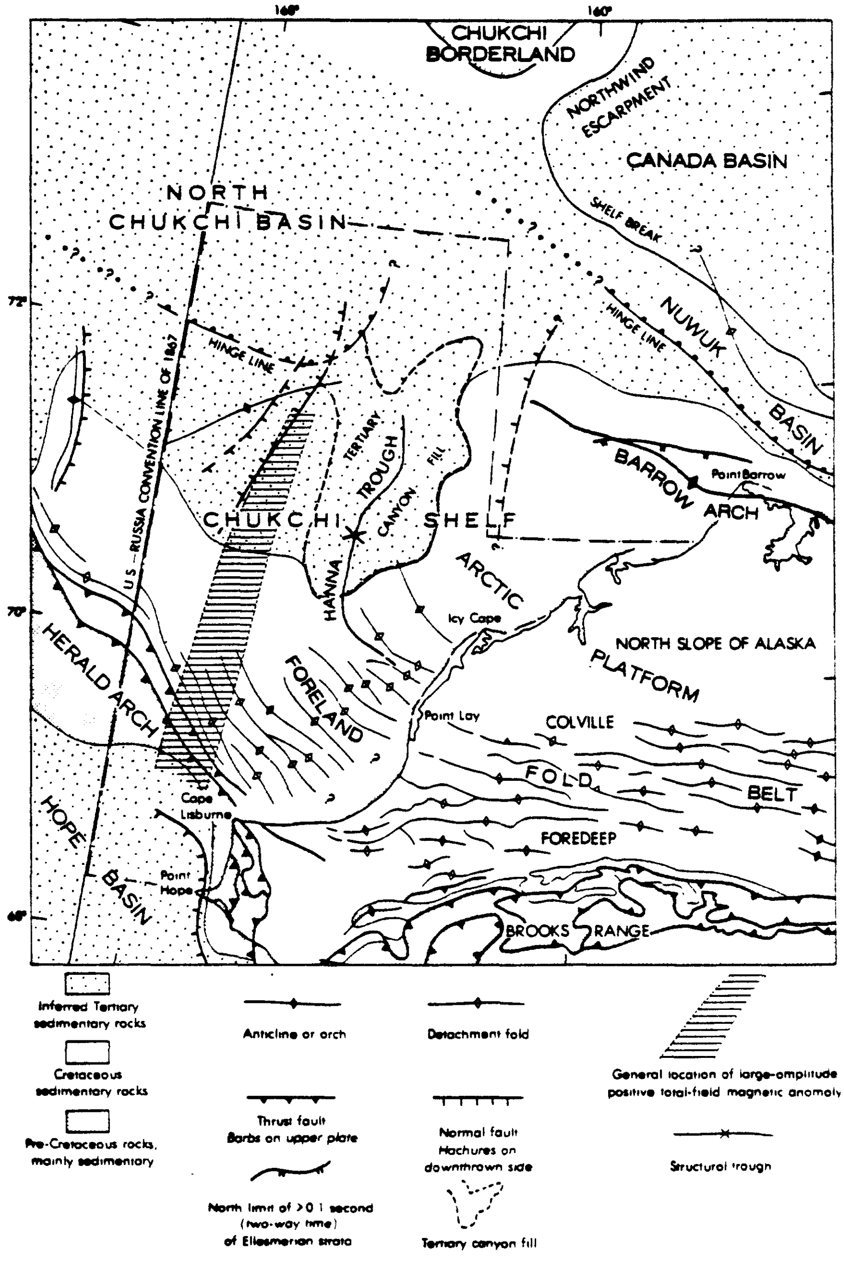


Figure 4.23. Tectonic map of the Chukchi Shelf and the western coastal plain reproduced from Grantz and others (1982). Late Cenozoic movements along the Hanna Trough and south of the Barrow Arch are inferred from the spatial distribution of Quaternary sediments.

CHAPTER V

INVERTEBRATE AND VERTEBRATE PALEONTOLOGY

> The rather striking difference in the fauna from different localities, in spite of the probability that the monotonous Arctic shelf did not offer any great contrasts in environment, may suggest that the Gubik is not one unit but consists of several thin units of slightly different age.
>
> --F.S. MacNeil, 1957

Invertebrate and vertebrate fossils have been collected from the sediments of the Gubik Formation for over 80 years (Dall, 1919). Mollusk assemblages from the Colville River area collected by Leffingwell (1919) and examined by Dall (1920) along with collections described by Meek (1923) from Skull Cliff hinted that perhaps strata of different age were present within the Gubik Formation. MacNeil (1957) synthesized what was known from these and other early collections and hypothesized that the Gubik consisted of several thin units of slightly different age. Later studies of ostracodes (Swain, 1963; Schmidt and Sellman 1966; Schmidt, 1967) and foraminifera (Tappan, 1951; Faas, 1962, 1966) were the first aimed at documenting the microfaunal assemblages of the Gubik Formation. All of these studies, however, lack a firm stratigraphic context.

Recent stratigraphic fieldwork and faunal collections by D.M McCulloch, J.R. Williams, D.M. Hopkins, L.D. Carter, and myself have added to our knowledge of the stratigraphic and paleontological aspects of the Gubik Formation. This chapter lists the invertebrate and vertebrate collections made during the course of this study, documents their stratigraphic context, and draws conclusions concerning their paleoecological significance.

All collections listed here were identified by paleontologists of the U.S. Geological Survey: collections were identified by Louis Marincovich and George Kennedy (mollusks); Elizabeth Brouwers (marine ostracodes); Richard Forrester (freshwater ostracodes); Dr. Kristin McDougall (foraminifera); and Charles Repenning (vertebrates). At the time of writing less than 15% of all my collections had been examined. Because a great deal of paleontological work remains, my conclusions are preliminary.

Mollusks

Modern collections

Mollusks were collected from the modern beach at several localities along Skull Cliff between Barrow and Peard Bay. Several collections were made immediately after storms when waves had delivered debris to the highest level of the swash zone. The faunal list compiled in Table 5.1 is not intended to be a

Table 5.1. Mollusks collected from the Modern Beach between Peard Bay and Barrow Alaska, July - August 1980 - 1981.

Bivalves:	Peard Bay	Nuvalik	Walakpa
Astarte borealis Schumacker	x		
Tellina (Peronidia) lutea alternidentata Broderip & Sowerby	x		
Mytilus edulis Linnaeus	x		
Serripes groenlandicus (Bruguiere)	x		
Macoma (Macoma) lama (Bartsch)	x		x
M. cf. M. brota Dall	x		
Cyclocardia crebricostata (Krause)	x		
Cyrtodaria kurriana Dunker	x		
Musculus niger Gray	x		
Hiatella artica Linnaeus	x		
Mya (Mya) truncata Linnaeus	x		
M. (Arenomya) arenaria Linneaus	x		
M. sp. fragments	x		
Zirfaea cf. Z. pilsbryi Lowe	x		
Clinocardium cf. C. californiense (Deshayes)	x		
Siliqua cf. S. patula (Dixon)	x		
Spilsula polynyma (Stimpson)	x		
Gastropods:			
Velutina undata Brown	x		
Neptunea hero heros (Gray)	x	x	x
N. (Neptunea) cf. N. (N. heros heros (Gray)		x	x
Margarites costalis (Gould)	x		
Buccinum tenellum Dall	x		
B. angulosum (Gray)	x	x	x
B. polare (Gray)			x
B. sp. A	x		x
B. sp. B	x		x
Beringius sp. indet.	x		
Plicifusus kroyeri (Moeller)	x		
Colus spitzbergensis (Reeve)	x		x
Natica (Cryptonatica) clausa Broderip & Sowerby	x	x	x
Amauropsis islandica (Gmelin)	x		
Oenopota sp.	x		x
Acmeaea testudinalis Müller	x		x

*Modern collections were also made along Skull Cliff during this study by G. Kennedy, however, at the time of writing, I have not, received a species list of these collections--U.S.G.S. location numbers M7903, M7904, M7921, M8928, M7932A, M7938, M7941, M7942, M7945, M7946.

comprehensive list of all living fauna. Rather it serves to supplement other faunal lists (MacGinitie 1955, 1959; Mann 1977) including those now being compiled by U.S. Geological Survey personnel.

One of the surprising finds among the modern specimens is the occurrence of Clinocardium cf. C. californiense (Deshayes) in collections at Peard Bay. This species reportedly has a modern range from Norton Sound, near Nome, to Sitka and perhaps Japan, and is therefore considered an extralimital form. It is possible that this specimen is a fossil retransported along the coast. The samples were collected from the beach at the east end of Peard Bay therefore the specimen may have been retransported by the southwestern flowing coastal current from the bluffs containing older sediments just 3 km up the coast. Zirfaea cf. Z. pilsburyi also occurs here beyond its previously known modern range. Based upon previous collections, this species has been considered a southern element that ranges only as far north as Point Lay (L. Marincovich, written comm., May, 1983). Because it is a boring clam, its distribution may be limited by the availability of exposed, soft, borable rocks on the seabed, rather than by climatic parameters.

It is useful to note that where the modern beach lies close to the base of the eroding bluffs, fossil marine shells are readily retransported onto the beach by slumping and small sediment flows. Caution must be exercised when sampling for modern or submodern specimens. Moreover, the modern range of mollusk species between Barrow and Cape Lisburne is poorly known. What is known is based on collections made by D.M. Hopkins and R.E. Nelson between Icy Cape and Peard Bay in 1976. Hence, my two finds may, in fact, expand the known range of Clinocardium californiense and Zirfaea pilsburyi.

Fossil Collections

Fossil mollusks were collected in so far as possible from each lithofacies in all sections. A portion of these samples were used for amino acid analysis to determine the relative age of the strata (Chapter III). Another portion of each collection was submitted for paleontological identification. These identified samples are representative of the total population of samples collected and provide a preliminary documentation of the paleoecology of marine units of different age in the Gubik Formation. Table 5.2 shows the distribution of the fossil mollusks identified thus far in the stratigraphic units defined in Chapter III. Extralimital species are highlighted by an asterisk and extinct species by a cross.

Two major conclusions can be drawn: first, the extinct species, Astarte leffingwellii (Dall) and Neptunea (Neptunea) lyrata leffingwelli (Dall), occur only in the oldest units, represented by aminozones 3, 4, and 5. These species had been previously recovered only from the late Pliocene and early Pleistocene beds along the coast of western Alaska (Hopkins, 1967a)

Table 5.2. Stratigraphic distribution of fossil mollusks collected from the Gubik Formation along the Chukchi Sea coast[1]

Species List	Aminozone 1	Aminozone 2	Aminozone 3	Aminozone 4	Aminozone 5	Found only as float
Gastropods:						
Natica clausa (Broderip & Sowerby)		0				
Natica (Cryptonatica) clausa (Broderip & Sowerby)	0	0	0	0	0	
Natica (Tectonatica) janthostoma (Broderip & Sowerby)			*			
Admete couthouyi (Jay)		0	0			
Admete sp. cf. Admete regina Dall			0			
Admete sp.	0	0	0	0		
Polinices (Euspira) pallidus (Broderip & Sowerby)		0	0	0		
Polinices new sp.?		0	0	0		
Buccinum sp.	0	0	0			
Buccinum sp. cf. Buccinum plectrum Stimpson		0				
Buccinum sp. tenellum Dall						0
Buccinum sp. cf. Buccinum tenellum Dall						0
Buccinum sp. cf. Buccinum normalise Dall, 1885			0			
Aforia circinata Dall, 1873						*
Plicifusus kroyeri (Møller)		0	0			
Neptunea ventricosa (Gmelin, 1791)					0	
Neptunea (Neptunea) lyrata leffingwelli (Dall)		†?	†		†	
Neptunea (Neptunea) heros heros (Gray)			0			
Neptunea sp.	0		0	0		
Neptunea (Neptunea) cf. N. borealis (Philippi, 1950)						0
Oenopota aff. Oenopota declivis (Lovén)			0			0
Oenopota sp.	0	0	0	0		
Oenopota cf. Oenopota arctica (A. Adams)			0			
Amauropsis islandica (Gmelin)			0			
Epitonium groenlandicum (Perry, 1811)			0			
Trichotropis sp. cf. Trichotropis borealis Broderip and Sowerby			0			
Lacuna sp.			*?			
Margarites sp. cf. Margarites helicinus (Phipps)			0			
Margarites sp. cf. Margarites groenlandicus (Gmelin)	0					
Colus sp.			0			
Volutopsius sp.						0
Acteocina sp. (Juveniles)	0					
Antiplanes sp.			0			
Bulbus fragilis (Leach, 1819)		0				
Retusa umbilicata (Montaqu)		0	0			
Hydrobia sp.		0				
Tachychynchus erosus (Couthouy)			0			
Tachychynchus . sp.			0	0		
"Cancellaria" sp.			0			
Boreotrophon sp. cf. Boreotrophon clathratus (Linné)			0	0		
Boreotrophon sp.	0			0		
Boreotrophon truncatus (Strom)					0	
?Cingula sp.						0
Solariella obscura (Couthouy)					0	

Table 5.2. (continued).

Species List	Aminozone 1	Aminozone 2	Aminozone 3	Aminozone 4	Aminozone 5	Found only as float
Pelecypods:						
Cyclocardia crassidens (Broderip & Sowerby)			0			
Cardita crebricostrata (Krause)		0	0	0	0	
Serripes laperousii (Deshayes)		0	0	0	0	
Serripes groenlandicus (Bruquiere)	0	0	0			
Clinocardium ciliatum (Fabricius)	0	0	0			
Clinocardium sp. cf. Clinocardium californiese (Deshayes)			*			
Clinocardium sp.	0		0	0		
Liocyma fluctuosa (Gould)	0	0	0	0		
Zirfaea pilsbryi Lowe			*			
Axinopsida orbiculata Sars			0			
Siliqua sp. cf. Siliqua alta (Broderip & Sowerby)	*		*			
Silqua sp.		*				
Nuculana (Nuculana) radiata (Krause)						0
Nuculana (Nuculana) cf. Nuculana (Nuculana) radiata (Krause)		0				
Nuculana sp.		0				
Thracia (Thracia) devexa Sars						0
Hiatella arctica Linneus	0	0	0	0	0	
Mya (Mya) pseudoarenaria Schlesh						0
Mya (Mya) truncata Linneus	0	0	0	0		
Yoldia (Yoldia) cf. Yoldia (Yoldia) myalis Couthouy		0				
Yoldia (Cnesterium) scissurata Dall, 1897	0	0				
Yoldia sp.		0	0		0	
Pandora (Pandorella) cf. Pandora (P.) gracialis Leach		0				
Pandora sp.			0			
† Cyrtodaria kurrianna (Dunker)			0			
Spisula (Mactromeris) polynyma (Stimpson)			0			
Astarte (Tridonta) borealis (Schumacher)			0	0		
Astarte (Tridonta) bennettii Dall, 1903						0
Astarte (Tridonta) montaqui (Dillwyn)						0
Astarte (Tridonta) cf. Astarte (Tridonta) montaqui Dillwyn			0			
Astarte (Tridonta) cf. A. (T.) leffingwelli Dall 1920				cf.		
Astarte leffingwelli (Dall)				cf.	cf.	
Astarte sp.	0		0			
Macoma balthica (Linneus)	0	0	0			
Macoma (Macoma) cf. M. (M.) obliqua (Sowerby, 1817)	0	0	0			
Macoma (Macoma) lama (Bartsch)			0	0		
Macoma (Macoma) cf. Macoma (Macoma) calcarea (Gmelin)	0					
Macoma (Macoma) cf. Macoma (Macoma) brota Dall		0	0	0		
Macoma sp.	0		0	0	0	

* extralimital species
† extinct species

and are known elsewhere in the Gubik Formation only from strata representing the lower part of the formation along the Colville River and Fish Creek (Carter, unpubl.). One specimen of N. (N.) lyrata leffingwelli was found in an upper unit at sec. 64 (9.7 - 10.5 m a.s.l.) that I had tentatively interpreted as belonging to a younger transgression characterized by aminozone 2 ratios. Based upon the Free alle/Ile ratios on Astarte from the '81 preparation results, this unit could also be reinterpreted as belonging to Aminozone 3.

Secondly, extralimital species, that is, species found as fossils in strata far beyond their modern ranges--occur in several units indicating that during certain past interglacial periods, marine conditions along the Chukchi coast were significantly different than they are today. Extralimital species are particularly common in the beds characterized by aminozone 3 ratios, which are believed to be of early Pleistocene age (Fig. 5.1). Of the species in this unit, Clinocardium californiense apparently has a modern range from Norton Sound in the Bering Sea to near Sitka in southern Alaska and possibly near Japan, Natica (Tectonatica) janthostoma ranges from Kamchatka to Hokkaido, and Vladivostock in the northwest Pacific, and, finally, Zirfaea pilsbryi probably reaches its northern limit near Point Lay, south of Wainwright. Specimens of Aforia circinata, a mollusk that lives today from the Bering Sea to the Straits of Juan de Fuca, Washington, were collected as float at the base of the bluffs near sec. 20 and may have been derived from marine strata of aminozone 3 in the cliffs above.

Marine beds at Fish Creek and Carter Creek (Fig. 1.1) that I consider correlative with the marine beds of Aminozone 3 on the basis of amino acid ratios contain Littorina sitchana Philippi (Carter and others, 1979). Living L. sitchana ranges today from the southern Bering Sea to Puget Sound, Washington, in intertidal waters. All species of Littorina are intolerant of severely subfreezing air temperatures and abrasion by sea ice (L. Marincovich, personal comm., 1983).

Sediments of Pelukian age (i.e., sediments correlative with Aminozone 1) along the Beaufort Sea coast of Alaska, host at least ten extralimital mollusk species, many of which reach their northern ranges at Peard Bay or at Barrow (Hopkins, OCSEAP report). These stratigraphic occurrences suggest that during the Pelukian high sea stand, the marine environment along the northern coast of Alaska was slightly warmer than exists today but was not nearly as warm as during the early Pleistocene.

Most of the gastropod and pelecypod species listed in Table 5.2 are presently found on arctic continental shelves in water less than 100 m deep and several, including Siliqua, Mya, Macoma balthica and Mytilus edulis rarely are found in water deeper than 10 m deep; Mya arenaria is strictly a shallow water species (Macpherson, 1971; Wagner, 1977; Mann, 1977; Bernard, 1979). Macoma balthica is a pelecypod that commonly prefers low-salinity

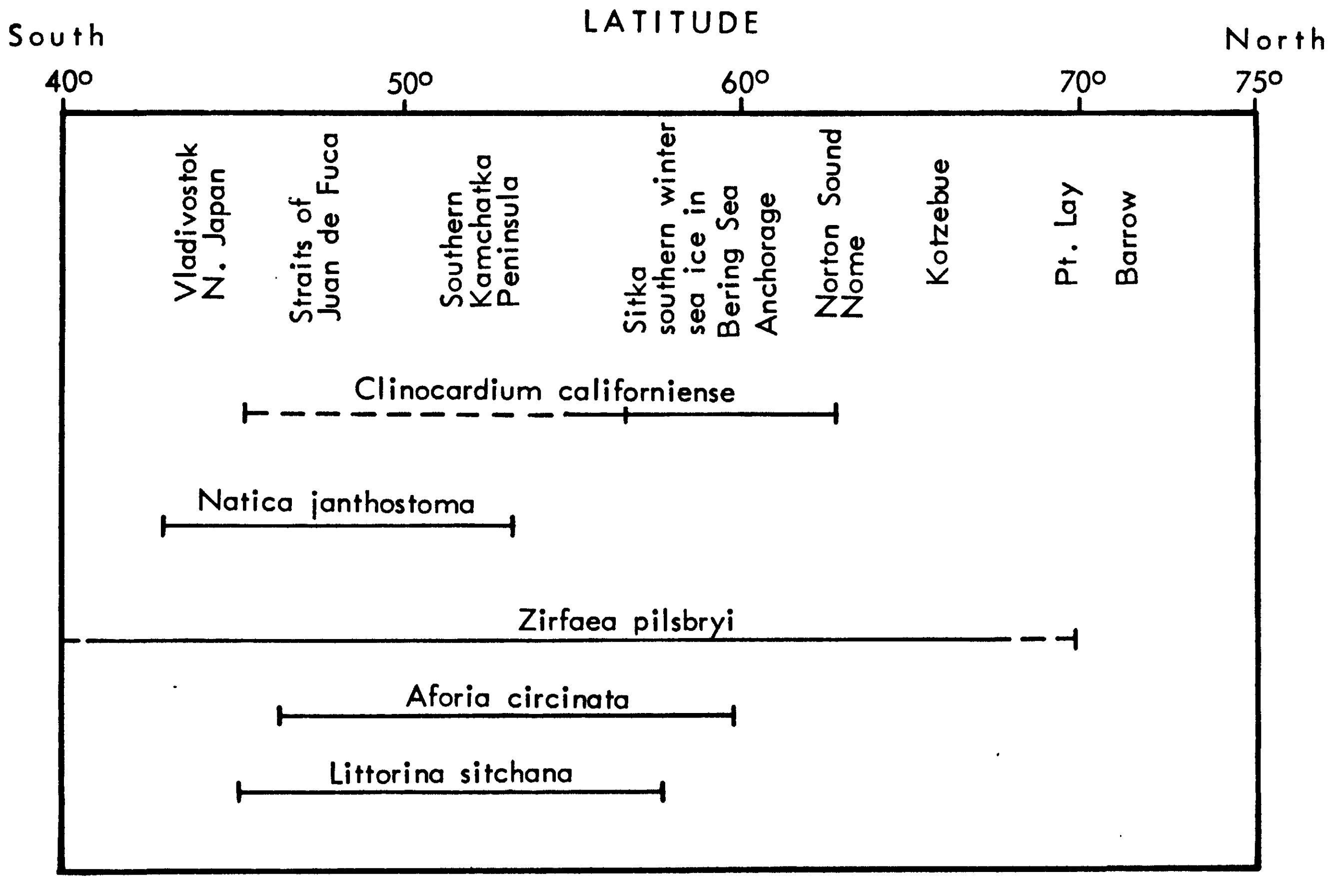


Figure 5.1. Modern latitudinal ranges (Louie Marincovich, pers. comm. 1982) of extralimital mollusk species found the marine units characterized by Aminozone 3.

environments. Empty valves have been recovered in open marine environments in the Beaufort Sea at depths of 270 m (Bernard, 1979) however, this exceptional occurrence may indicate the exposure on the seabottom of old, shallow water deposits. Some species, such as Liocyma fluctuosa and Axinopsida orbiculata, live today at optimum depth ranges of 20-40 m and 13-70 m, respectively.

Ostracodes

Modern collections

Little is known about the modern distribution of marine ostracodes in the northeastern Chukchi Sea. MacGinitie (1959) described the occurrence of a few species from grab samples off Pt. Barrow. The only other studies of ostracodes on the arctic Alaskan shelves are limited to the Beaufort Sea (Painter, 1965; Briggs, 1980, in press; Brouwers, 1982).

Fossil collections

Bulk sediment samples for microfaunal analysis were taken from most, but not all, lithofacies at each stratigraphic section. Forty-three samples were selected for faunal analysis of enclosed ostracode assemblages from representative lithofacies spanning a wide age range; 18 of these samples turned out to be barren of ostracodes. A species list is presented in Table 5.3, along with a chart indicating the stratigraphic distribution of each species. No samples from Killi Creek or Nulavik beds were examined.

Rabilimis paramirabilis Swain, 1964, is unique being the only extinct species found; it is thought to be restricted to late Pliocene/early Pleistocene marine units (Hopkins and others, 1974). This species has been found thus far only in the marine beds of Aminozone 3 but it would be expected in faunas from beds of Aminozone 4 and 5. R. septentrionalis, a related form thought to be a descendent, occurs in marine beds of Aminozone 3 and younger marine units. R. paramirabilis and R. septentrionalis both occur in the marine beds of Aminozone 3 at Skull Cliff, and also in a marine unit recovered in cores from the Beaufort Sea (P.A. Smith, U.S. Geological Survey, personal comm., Jan. 1983).

Cythere cf. C. lutea (Brady, 1968) and Baffinicythere emarginata (Sars, 1865) are the only extralimital forms present suggestive of marine conditions slightly warmer than observed today. These ostracodes now live in subfrigid waters from the southern Chukchi Sea to the Aleutian Islands (Brouwers, written comm., Jan. 1983). Both have been found in the marine beds of Aminozone 2 but not elsewhere.

Most of the ostracode species are characteristic of shallow-shelf and nearshore environments (Brouwers, written comm., 1983). Rabilimis septentrionalis, Cytheromorpha spp., Heterocypriders, Loxoconcha venepidermoidea and Paracyprideis are euryhaline and eurythermal species characteristic of shallow, nearshore conditions. This assemblage is typical of nearshore and

Table 5.3. Stratigraphic distribution of fossil ostracodes in sediments of the Gubik Formation, Skull Cliff, Alaska[1]

Species List	Thaw lake sediments	Eolian sediments	Aminozone 1	Aminozone 2	Aminozone 3	Comment
MARINE SPECIES:						
Eucytheridea bradii (Norman, 1863)			0	0	0	
Eucytheridea punctillata (Brady, 1865)				0		
Eucytheridea macrolaminata (Elofson, 1939)				0		eastern Beaufort, N. Atlantic affinity
Eucytheridea n. sp.				0		
Rabilimus septentrionalis (Brady, 1866)			0	0	0	mid Pleist. to recent
Rabilimus ? paramirabilis (Swain, 1963)					†	upper Pliocene to
Rabilimus paramirabilis (Swain, 1964)					†	lower Pleist.
Cytheromorpha macchesneyi Brady and Crosskey (1870)			0	0		seasonally fluct. salinity or temp.
Cytheromorpha sp.			0			seasonally fluct. salinity or temp.
Paracyprideis pseudopunctillata Swain, 1963			0	0	0	
Paracyprideis ? pseudopunctillata Swain, 1963					0	extinct ??
Cythere cf. Cythere lutea Brady 1968				*		warmer H_2O subfrigid southern Chukchi to Aleutians
"Acanthocythereis dunelmensis" (Norman, 1865)					0	Atlantic affinity
Robertsonites tuberculata (Sars, 1865)				0	0	
Heterocyprideis sorbyana (Jones, 1856)				0	0	
Normanicythere leioderma (Norman, 1869)			0	0	0	Pleist. to recent
Cytheretta teshekpukensis (Swain, 1963)				0		
Baffincythere emarginata (Sars, 1865)				*		warmer H_2O subfrigid
Cytheropteron pyramidale of Swain (1963)			0	0		
Loxoconcha venepidermoidea Swain, 1963				0		
Finmarchinella (Barentsovia) angulata (Sars, 1865)			0			Atlantic affinity
"Paracyprideis" sp.						in Miocene sediments
Palmanella limicola (Norman, 1865)				0	0	
FRESH WATER SPECIES						
Candona sp.	0	0				reworked in eolian sediment[s]
Candona rectangulata Alm, 1914	0					
Limnocythere n. sp.	0			0		reworked in marine sediment[s]
Limnocythere inopinata (Baird, 1843)	0					
Limnocythere platyforma Delorme, 1968	0					
Limnocythere liporeticulata Delorme, 1968	0					
Cypricercus foveata (Delorme, 1968)	0					only in deposits > 25 ka B.P.
Cypricercus sp.	0					
Prinoncyris serrata (Muller, 1900)	0					
Megalocypris sp. indet.	0					
Cypria opthalmica	0					Arctic to Central U.S.
Cytherissa lacustris	0					
Ilyocypris sp.	0					
I. bradii Sar, 1890	0					

* extralimital
† extinct

shoreface sediments of Aminozone 1. Schmidt and Sellman (1966) have described mummified specimens of Paracyrideis pseudopunctillata from Aminozone 1 beds near Barrow. Open marine conditions are characterized by species such as Normanicythere leioderma, Robertsonites tuberculata, and Palmanella limicola (Brouwers and others, 1983).

Individuals of endemic Atlantic species, including "Acanthocythereis" dunelmensis (Norman, 1865), Eucytheridea macrolaminata (Elofson, 1939), and Finmarchinella (Barentsovia) angulata (Sars, 1865), occur in the marine beds of Aminozones 3, 2, and 1, respectively. Because of the small number of ostracode collections and the lack of identified collections from marine beds of Aminozones 4 and 5, the stratigraphic significance of these occurrences remains uncertain.

Fresh-water ostracode faunas were extracted from the sandy thaw lake lithofacies at the top of sections 28 and 29 (Table 5.3). Cypricercus foveata from a thaw lake deposit in Sec. 29 seems to be restricted on the North Slope to lake sediments older than 25 ka B.P. (L.D. Carter and R. Forrester, pers. comm., Jan., 1983). The rest of the assemblage at this site suggests that the ancient thaw lake was less than 3 m deep and had a salinity within the range of 300 ppm to 1,000 ppm. The lake, when active, was either near the ocean or evaporation was high and climate was drier (R. Forrester, written comm., Jan. 1983).

Foraminifera

Modern collections

The modern distribution of foraminifera on the floor of the northeast Chukchi Sea is known only from the survey of Cooper, (1964) and from grab samples recovered off Barrow (MacGinitie, 1963). Recent and ongoing studies in the Beaufort Sea include those by Laqoe (1979a, 1979b), Vilks and others (1979) and McDougall (1983).

Fossil collections

Of 14 microfossil samples from Skull Cliff examined by K.A. McDougall, nine were barren of foraminifera. Table 5.4 lists the stratigraphic distribution of species present in the remaining 5 samples. All are typical shallow water shelf species found in water depths of less than 20 m. The presence of Elphidium orbiculare, E. bartletti and E. incertum and the absence of E. excavatum alba is apparently characteristic of Pelukian sediments (K. McDougall, written comm., Jan. 1983).

Samples of blue-gray plastic clay of Miocene age informally designated in this study as the Papigak clay which crops out below the Gubik Formation in parts of Skull Cliff (Chapter II, Plate I) were also washed and examined. Most of these splits were barren. Three samples, however, yielded a rich fauna of calcareous, arenaceous, and agglutinated foraminifera quite unlike the known

Table 5.4 Stratigraphic distribution of Foraminifera in 5 samples from the Gubik Formation

Species list[1]	Aminozone 1	Aminozone 2	Aminozone 3	Miocene
Buccella frigida (Cushman)	0	0		
Elphidium bartletti (Cushman)	0			
Elphidium incertum (Williamson)	0	0		
Elphidium orbiculare Brady	0	0	0	
Elphidium groenlandica (Cushman)		0	0	
Elphidium cf. Elphidium bartletti (Cushman)		0		
Elphidium clavatum Cushman		0	0	
Elphidium frigidum Cushman		0		
Elphidium sp.		0		
Guttulina glacialis (Cushman and Ozawa)		0		
Polymorphina subolongata (Cushman and Ozawa)		0		
Quinqueloculina seminulum (Linné)		0		
Quinqueloculina sp.			0	
Ammodiscus sp.				0
"Bolivina" sp. of Todd, 1957				0
(probably Textularia)				0
Cibicides perilucidus Nuttail				0
Cyclammina sp.				0
Dentalina sp. cf. D. soluta Reuss				0
? Discorbis sp. (maybe Buliminella curta of Todd, 1957)				0
Eggerella bradyi (Cushman)				0
Eggerella sp. aff. Eggerella elongata Blaisdell				0
Eggerella subconica Parr				0
Gaudryina sp. cf. Gaudryina triangularis Cushman				0
Gryoidina sp. cf. Gryoidina girardana (Reuss)				0
Involutina sp.				0
Lenticulina cultratus (Montfort)				0
Marginulina sp. cf. Marginulina crebicosta sequenza				0
Sarocenaria hantkeni Cushman (Marginulina hantkeni, of Todd 1957)				0
Silicosigmoilina groenlandicus (Cushman)				0
Unidentified arenaceous forms				0

[1]The Karmuk and Tuapaktushak samples are from Sec. 20 at Skull Cliff. The Walakpa sample is from Sec. 37 near Barrow

Arctic Pleistocene assemblage (locations Fig. 2.4). These faunas are most similar to faunas in the Nuwok member of the Sagavanirktok Formation at Carter Creek (Todd, 1957). By comparison, the Papiqak Clay fauna is thought to be correlative with the Nuwok member and to date from the middle Miocene (K. McDougall, written comm., Jan. 1983; pers. comm., June 1983).

The 13 foraminiferal species listed in Table 5.3 may not be representative of Gubik deposits. Tappan (1951) described 17 "index species" and Faas (1962) described 34 species from portions of the Gubik near Barrow. It is unknown how their sites correlate with the stratigraphy presented here.

Vertebrate Fossils

Remains of sea mammals and, less commonly, birds, are sparsely scattered through the Gubik Formation, and remains of terrestrial mammals are common in the overlying non-marine deposits and in continental deposits elsewhere on the coastal plain. Although bones are most commonly found out of stratigraphic context on gully floors, riverbanks, and beaches, vertebrate remains have also been found *in situ* in the Gubik Formation as well as in non-marine sequences on the Arctic coastal plain. Carter (1982) presents radiocarbon dates and stratigraphic settings for several occurrences of land-mammal remains, and Repenning (1983) describes and discusses several important occurrences of fossil marine mammals.

I recovered several fossil mammal bones while excavating stratigraphic sections in the Gubik Formation along Skull Cliff and elsewhere. Their location and stratigraphic positions are described below (Table 5.5). The bird bones listed were identified by Storrs L. Olson, Smithsonian Institution, and the others were identified by Charles Repenning, U.S. Geological Survey.

The sternum of a murre (*Uria* sp.), collected from beach sediments of Aminozone 1 at Wainwright, provides an interesting puzzle. *Uria aalge*, the Common Murre, does not now live in the Arctic Basin, hence had it been this species, the find would be paleoclimatically significant. Nevertheless, the alternative possibility, the Thick-billed Murre (*U. lomvia*), is uncommon in the Arctic north of where the fossil specimen was found. Moreover, Murres do not breed today north of Cape Lisburne (S.L. Olson, written comm. to C.A. Repenning, Nov. 1982). Alternatively, Udvandy (1963, p. 88) reports that *U. lomvia* "is even now a successful pioneer of the Arctic coasts and the only arctic alcid which has colonized the western Canadian arctic". His figure 5 shows this species nesting at Point Barrow and at Cape Bathhurst on the Canadian mainland coast opposite Banks Island. Murres are generally cliff-nesting birds, so their distribution is limited as much by suitable nesting habitat as by climate. Thus, the presence of Murre in sediments of Aminozone 1 suggests that some other factor, other than habitat, allowed these birds to extend their range.

Table 5.5. Location and Stratigraphic Position of Some vertebrate fossils retrieved from the Gubik Formation, 1980 and 1981.[1]

Fossil I.D.	Sec. 27 80 Akb 157	Sec. 50 81Akb 270	Sec. 72 81 Akb 475	Sec. 78 81 Akb (Bfloat)	Sec. 79 81 Akb 534	Sec. 90 588 A + B	Sec. 94 81 Akb 596	Sec. 106 81 Akb 619	Stratigraphic Position or Comment
BIRDS[2]									
coracoid of *Larus* (*Rissa*) *tridactylus* (Kittiwake)		X							from Aminozone 1 sediments at Barrow
sternum of *Uria* sp. (Murre)					X				from Aminozone 1 sediments at Wainwright
MARINE MAMMALS									
(Whale?) cetacean									
bone fragment		X							Aminozone 1 sediments
rib fragment						X			Aminozone
mandibular fragment						X			2 or 3
Erignathus sp., first pedal phalanx I of bearded seal			X						Aminozone 2
Pusa ? sp., fibula of ringed seal				X					possibly Aminozone 1
Odobenus sp., cervical vertebra of walrus						X			inland beach along Kupowruk R.
TERRESTRIAL MAMMALS									
proboscidian, shaft of juvenile humerus							X		mammoth bone in eolian silts overlying marine sediments
Mammuthus sp. cf. *M. columbi* molar fragment								X	collected on modern beach of Kokolik R.
Bison sp., tooth fragment	X								Aminozone 4

[1]Location of stratigraphic sections is indicated in Figs. 2.1a and 2.1b

[2]The bird fossils have been incorporated into the U.S. National Museum collection, Smithsonian Institute, Washington, D.C.

The Erignathus sp. (bearded seal) bone was recovered from shallow, inner-shelf lithofacies of Aminozone 2 thought to be of middle Pleistocene age in Sec. 72. This specimen may be the earliest record of the genus in the Chukchi Sea (C.A. Repenning, written com., Oct., 1982).

A molar fragment of Mammuthus sp. cf. M. columbi was found on a berm of the Kokolik River, some 33 km from the Chukchi Sea (Sec. 106, Fig. 2.1b). Although the specimen was small, the sample is thought to represent an early glacial or pre-glacial fauna that existed during early Nebraskan or pre-Nebraskan time (i.e., >700 ka B.P.; Repenning written comm., Oct. 1982). Although this sample was redeposited, it suggests the presence of very old strata near and upstream from this site.

Repenning (1983) has recently outlined the temporal and stratigraphic significance of fossil marine mammals at several localities on the Arctic Coastal Plain. These specimens included the left mandibular ramus of the fossil sea otter, Enhydra, collected in 1975 by L.D. Carter from strata near Ocean Point along the Colville River and part of the dorsal portion of an Enhydra skull collected in 1955 by K. Hussey and J. O'Sullivan from strata along the sea bluff southwest of Barrow near Walakpa Bay (Fig. 5.2). The sea mammal remains represent valuable biostratigraphic markers and contribute to knowledge concerning the age of different marine units on the coastal plain. Repenning (1983) concluded that the Ocean Point sea otter is of latest Pliocene age dating between 2.2 and 1.7 my B.P.; whereas the Walakpa Bay sea otter is a much later form.

I have differentiated two marine units in the bluffs near the site of the original collection of the Walakpa Bay sea otter; including a lower unit of middle Pleistocene age (i.e., Aminozone 2) and an upper unit of Pelukian or last interglacial age (i.e., Aminozone 1) (Sec. 78, Plate II in Chapter II). Because several valves of Mya cf. M. japonicus were collected in conjunction with the sea otter bone by Hussey and O'Sullivan, the stratigraphic position of the sea otter remains can be established on the basis of amino acid analyses. Assuming that the Mya specimens are contemporaneous with the sea otter, alle/Ile ratios of 0.215 ± .006, Free, and 0.033 ± .002, Total (AAL-3193), indicate that the vertebrate fossil belongs to Aminozone 2 and was collected from the lower unit. These ratios are much lower than ratios determined on Hiatella arctica collected by L.D. Carter with the sea otter at Ocean Point (0.647 ± .006, Free and 0.134 ± .003, Total (AAL-2962).

Though sea otters live today in nearshore waters far south of the Bering Strait, they are quite mobile. Strays have been sighted in Norton Sound, in the Beaufort Sea, and in Chaun Bay on the Siberian coast opposite Wrangell Island (D.M. Hopkins, written comm., Sept., 1983). The occurrence of a sea otter near Barrow in Aminozone 2 marine sediments, therefore, does not necessarily imply warmer marine conditions.

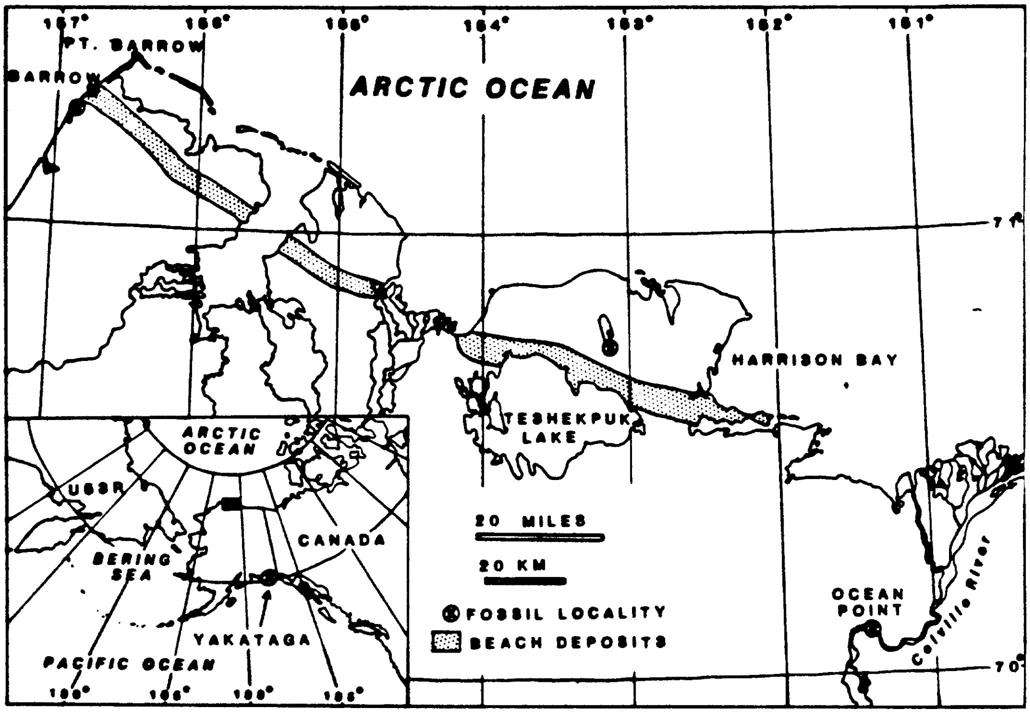


Figure 5.2 Map showing the location of the fossil Enhydra localities at Ocean Point and near Barrow (reproduced from Repenning 1983, after Carter and Robinson, 1981).

Paleoecological Significance of Faunal Collections

The mollusk and ostracode fauna provide the most useful proxy data for interpreting former ecological conditions (Table 5.6). Little faunal material was recovered from the marine beds of either Aminozones 4 or 5. Correlative units with rich faunas are known from the Colville River, however, and these provide some data concerning the paleoclimatic aspects of these marine incursions (e.g., Carter and Galloway, 1982; Carter, unpubl. faunal lists).

Of the three youngest marine units, the Aminozone 3 beds are characterized by the largest number of extralimital species, suggesting that this marine event was warmer than the present interglacial conditions. The modern ranges for three of the four extralimital forms in these beds (C. cf. C. califoniense, Natica (Tectonatica) janthostoma and Aforia circinata) are restricted to regions south of the Bering Strait. Two species of this subset live in areas that support only a few months of winter sea ice and they inhabit warmer waters as far south as Puget Sound. N. (T.) janthostoma lives well south of the southern limit of winter sea ice. Littorina sitchana, found in beds near Ocean Point equivalent to the marine beds of Aminozone 3 is a southern element that also will not tolerate sea ice or subfreezing temperatures, a fact that is incompatible with the severe temperatures and ice conditions that now characterize the Chukchi and Beaufort seas. This molluscan assemblage in Aminozone 3 sediments precludes the existence of any coastal sea ice adjacent to northern Alaska during some portion of the early Pleistocene and may well preclude the existence of perennial ice over the Arctic Ocean. This conclusion bears important implications concerning the climatic evolution of the high Arctic and contradicts the long held conclusions of Clark (1982, and references therein) that the Arctic Ocean has been premanently ice covered since 5 m.y. B.P. Perhaps some early episodes of high Arctic glaciation during the late Pliocene and earliest Pleistocene (Shackleton and Opdyke, 1977) were initiated and nourished by a warm, interglacial Arctic Ocean rather than by southern moisture sources such as the North Atlantic Drift.

In contrast to the marine beds of Aminozone 3, the beds of Aminozone 2 lack extralimital mollusks, although it hosts two extralimital ostracodes, Cythere cf. C. lutea and Baffincythere emarginata, and the remains of a sea otter. Both ostracodes exist today in subfrigid waters from the southern Chukchi Sea to the Aleutian Islands. Sea otters also live south of the Bering Strait today. These elements, although few, suggest that marine conditions during this middle Pleistocene marine event may have been slightly different than today.

Several mollusk species that today reach their northern limit in the northeast Chukchi Sea, extended their range into the Beaufort Sea during the Pelukian (last interglacial) high sea stand suggesting that this interglacial was only slightly warmer than today. The presence of a murre (Uria sp.) in the Wainwright area supports this appraisal.

Table 5.6. Summary of Significant Elements concerning the Paleontology of Gubik Formation, Western Arctic Coastal Plain.

Aminozone (Table 3.5)	Mollusks	Foraminifera	Ostracodes	Vertebrates	Composite Summary
1	modern fauna	modern shelf fauna	modern shelf fauna	possible range extension of Murre	marine conditions like today or slightly warmer
2	modern fauna no extralimitals	modern shelf fauna	includes 2 extralimital forms from S. Bering Sea	Walakpa sea otter beyond normal range; Earliest record of *Erignathus* sp. in Chukchi Sea	marine conditions like today
3	includes 2 extinct forms; and 4(?) extralimital species from S. Bering Sea and N. Pacific	modern shelf fauna	includes extinct form; no extra-limitals		marine conditions warmer than today: little or no sea ice in Arctic Ocean
4	includes at least one extinct species; no extralimital recovered to date	?	?		unclear
5	includes 2 extinct species: no extralimitals	?	?		unclear

The paleoecology of the Alaskan arctic shelves and the paleoclimatic history of interglacial periods in the Arctic is encapsulated stratigraphically in the sediments of the Gubik Formation. The stratigraphic distribution of mollusks and ostracodes provides the most sensitive information concerning changes in the marine environment during sucsessive interglacial periods; additional foraminifera studies may prove useful as well. The evolution of marine mammals and changes in their distributions form valuable geochronologic and paleoclimatic markers within the Gubik Formation. The data presented here provides the first stratigraphic documentation of the Gubik mollusk and microfossil faunas.

CHAPTER VI

PALEOTEMPERATURE ESTIMATES OF THE ALASKAN ARCTIC COASTAL PLAIN DURING THE LAST 125,000 YEARS

The extent of isoleucine epimerization in fossil protein is largely dependent upon temperature (the thermal history of the sample since deposition) and time (since death of the organism), the former being the most sensitive factor affecting the rate of hydrolysis and racemization (epimerization). If one of these two variables is known (or assumed), the second can be derived (Miller and others, 1983). This chapter focuses upon the extent of isoleucine epimerization from mollusks enclosed in marine sediments of the last interglacial high sea stand (Aminozone 1, Chapter III) assumed to be 125 ka old. Based upon analytical results from *Mya truncata* and *Mya arenaria* in these sediments (Table 3.5) the data are used to constrain a series possible paleotemperature models for the North Slope during the last 125,000 years that attempt to accomodate all of the presently known paleoclimatic proxy data.

Paleotemperature Estimates

Equations which express the relationship between the extent of epimerization, absolute time, and diagenetic temperature are reviewed by Schroeder and Bada (1976) and Williams and Smith (1977) and may be expressed as:

$$\ln \frac{1 + D/L}{1 - K' D/L} = (1 + K')\, k_i t + C \qquad (1)$$

where D/L = aIle/Ile ratio of the sample; K' = inverse of the equilibrium constant (0.77); k_i = forward rate constant of isoleucine epimerization; t = time since death of the organism; and C = analytical constant that represents the value of the left side of the equation at death of the organism and accounts for laboratory induced epimerization during preparation (C = 0.0195 in samples discussed here).

The relationship between k_i and temperature is described by the Arrhenius equation

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$$k_i = Ae^{\left(\frac{-Ea}{RT}\right)} \qquad (2)$$

where A = constant for each species; Ea = energy of activation; R = gas constant (1.987); and T = absolute temperature (°K).

Radiocarbon dated control samples and pyrolysis data further define the relationship between the isoleucine epimerization rate constant (k_i) and temperature (T) for *Mya truncata* (Miller and others, 1983; Miller 1983, in press). The relationship between k_i and T for *Mya truncata* is expressed as:

$$\log k_1 = 16.45 - 6141/T \qquad (3)$$

and the energy of activation is calculated to be 28.1 kcal mole. By replacing equation (3) for k_1 in (1) and solving for T, the equation can be written with T, t, and D/L as the only variables.

$$EDT°K = \frac{6141}{16.45 - \log \dfrac{\ln \dfrac{1 + D/L}{1 - .77\,D/L} - 0.0195}{1.77t}} \qquad (4)\ \dagger$$

For a sample with a complex thermal history, T can be considered the effective diagenetic temperature (EDT) that is "the temperature associated with an effective rate constant, which is the time-averaged rate constant estimated by integration through a probable temperature history for a sample," (Wehmiller and others, 1977), p. 62-63). Because of the exponential relationship between the rate constant (k_1) and temperature (T), this thermal integration is more sensitive to periods of high temperatures that accelerate the reactions more than the deceleration effected by an equivalent temperature decrease.

The effects of a variable thermal history on the EDT can be important on both an annual scale and on the scale of long-term climatic shifts. The arithmetic mean annual temperature of the ground will deviate from the effective diagenetic temperature in proportion to the amplitude of temperature about the mean (Miller and others, 1982). Although the mean annual air temperature at Barrow is currently -12.8°C, for example, integrating the monthly temperature cycle over the year calculates into an EDT of -3.6°C, reflecting the greater influence of warm summer temperatures. In permafrost, the value of the EDT is influenced by the amplitude of the annual temperature wave at a given depth and at the depth of zero amplitude approximates the mean annual ground temperature.

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† It is important to caution the reader that the energy of activation for isoleucine epimerization (Ea in Equation 2) is assumed to be 28.10 kcal mole^{-1} throughout the diagenetic history of a shell sample. Recent work on bone protein (Hare 1980) indicates that in bone epimerization only commences after denaturation of the complex collagen moluecule has reduced the protein to substantially lower molecular weight forms, and that the activatioin energy for this reaction may be twice that of epimerization. At cold sites (<-6°C) insufficient energy may be available to complete the denaturation in 10^5 -10^6 yr (P.E. Hare unpublished data). Shell matrix proteins are considerably simmpler than is collagen; nevertheless it is theoretically possible that initial high activation energy reaction may be required to degrade the shell protein before epimerization can occur, although no experimental data to confirm this are yet available. The effect on the temperature calculations reported herein would make them too low.

Figure 6.1a shows the decrease in the EDT with depth calculated from the current mean monthly temperatures measured in permafrost near Barrow (data taken from Lachenbruch and others, 1962). Near the ground surface, the EDT of the permafrost is quite high (≈ -3.0°C at -60 cm) reflecting the high summer temperatures and large amplitude of the annual temperature wave. With depth, however, the amplitude of the annual temperature wave is increasingly attenuated to a point where the EDT in the permafrost (EDT -9.3°C at -18.5 m) is close to the mean annual ground-surface temperature (≈ -10°C).

It is possible to evaluate the effect of the annual temperature wave and hence, the EDT, on the rates of isoleucine epimerization under the present climate and during colder glacial climates. The effect of the present climate (using the EDTs at depth in Figure 6.1a) on rates of epimerization is illustrated for a hypothetical case in Figure 6.1b in which the present climate is maintained for 100,000 years. Using equation (4) and solving for D/L in successively deeper samples, it can be shown that below a depth of approximately -3 m (dashed line in Figure 6.1b), the aIle/Ile ratios resulting from different EDTs differ by 11.5%; this is equivalent to the precision (expressed as coefficient of variation, defined Table 3.5) attainable in isoleucine epimerization analyses (compare with Table 3.5; also Brigham, 1983). For comparison, a series of Holocene samples subjected to the modern climate for 8,000 years would theoretically yield a depth related variation in aIle/Ile ratios of only 2.5% (Figure 6.1b). From this exercise, I conclude that (1) over relatively short time intervals, minor changes in the EDT's below about -6°C effect only minor variations in the observed aIle/Ile ratios, (2) consequently, paleotemperature estimates for low-temperature sites are only precise to within -2° to -3°C, (3) that samples collected for amino acid studies in permafrost areas should come from a depth of at least 2.5 m to minimize the effect of the annual temperature wave, and most importantly (4) these data, in turn, allow one to conclude that the mean permafrost temperature at depth is a good first approximation for the long-term EDT experienced by buried shells during each climatic regime over the last 125,000 yrs. In this study, the EDT expressed by equation (4), represents an integration of the thermal history of the upper few tens of meters of permafrost over an entire time period (t), and places constraints on the magnitude of ground and air temperature changes that can have occurred.

Paleotemperature Reconstructions

Paleoclimatic proxy data and geomorphic evidence demonstrate that Alaska has experienced a number of broad climatic oscillations during the last 125,000 years, including warm and moist interglacials similar to today, cold and dry glacials, and intermediate interstadial conditions (see Hopkins and others,[1982] for reviews). Most paleoclimatic reconstructions for this time period have been qualitative in approach due to a lack of means to quantify temperature and precipitation parameters. Based upon

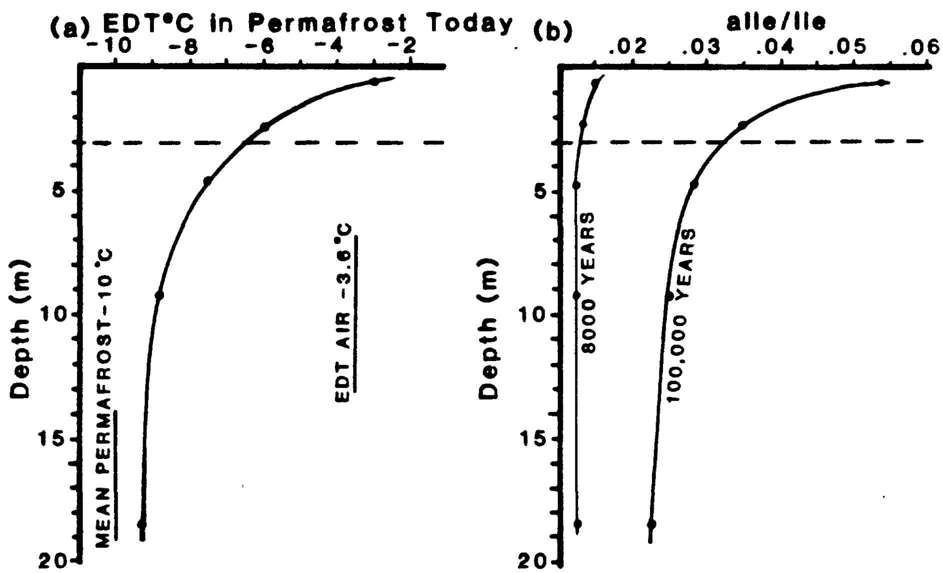

FIGURE 6.1. Effective diagenetic temperatures (EDT) and hypothetical aIle/Ile ratios in permafrost. (a) Calculated EDT at various depths in permafrost under the present climate. (b) Effects of the modern EDT profile on the aIle/Ile ratio in samples 8.0 ka and 100 ka B.P. The differences in ratios observed below 3 m (dashed line) are less than the analytical variation in the amino acid method. For 100 ka, the mean aIle/Ile below -3 m = 0.026 ± .003, c.v. = 11.5%; for 8 ka, the mean is 0.12 ± .0003, c.v. = 2.5%.

amino-acid paleothermometry, a series of paleotemperature reconstructions are presented for the Alaskan Arctic Coastal Plain during the last 125,000 years that attempt to accomodate all of the available paleoclimatic proxy data from the region and to quantify and define limits to the magnitude of temperature change during this time period.

Paleotemperature estimates derived from equation (4) are most accurate when the absolute age of a sample is known. Hence, the most important assumption in this reconstruction is that Aminozone 1 is 125,000 years old; the rationale for this age assignment is outlined later (Chapter VII). Younger age estimates can be tested by solving equation (4) for EDTs using other assumed ages with the observed aIle/Ile ratios. Age assumptions of 80 ka and 35 ka B.P. for Aminozone 1 predict EDTs and thus ground temperatures similar to or significantly warmer than modern values (Table 6.1). In light of paleoclimatic data that suggests the glacial periods in northern Alaska were colder and drier than modern interglacial conditions, I believe that age estimates of 80 and 35 ka are unrealistic and that an age of 125,000 is probably most reasonable for Aminozone 1.

AIle/Ile ratios in *Mya* from last interglacial deposits (Aminozone 1, t = 125,000 years B.P.) require an effective diagenetic temperature for the entire time period of ≈ -14.0°C (Table 6.1). Using the paleoclimatic record as a guide to estimate temperatures during interglacial periods, models can be developed that derive appropriate EDTs for periods lacking proxy data, especially full-glacial episodes. The models employ a computer program that allocates rate constants for intervals of specified climate, thus integrating an effective diagenetic temperature over each interval, and converts the computed rate constant back into a temperature (from equation 3) for an interval of unknown climate (Miller and others, 1983). The equation used for these reconstructions is modified from Miller and others (1983) and written as:

$$x = \sum_{i=1}^{n} \frac{k_a t_a - (k_1 t_1 + k_2 t_2 + \ldots k_n t_n)}{t_2 - (t_1 + t_2 + \ldots t_n)} \qquad (5)$$

(see Figure 6.2)

where, k_x = rate constant for a specific time interval of unknown temperature; k_a = rate constant for the entire assumed 125,000 year period; k_1, $k_2 \ldots k_n$ = rate constant for intervals of known temperature; t_a = length of entire time period = assumed 125,000 years; t_1, $t_2 \ldots t_n$ = length of each climatic interval. The effective diagenetic temperature for the unknown interval is calculated using the calculated k_x in place of k_1 in equation 3.

Paleoecological data from numerous Alaskan sources across Beringia suggest that (1) the last interglacial was slightly warmer

Table 6.1. Effective Diagenetic Temperature in Permafrost predicted by Equation (4) assuming different ages (t) for Aminozone 1.

D/L (Mya)	Absolute age	Predicted EDT°C	Comments
.018	35 ka	-7.8	ca. 4°C warmer than today - not reasonable
.018	80 ka	-11.9	similar to pre-20th century permafrost
.018	125 ka	-14.0	slightly colder than today

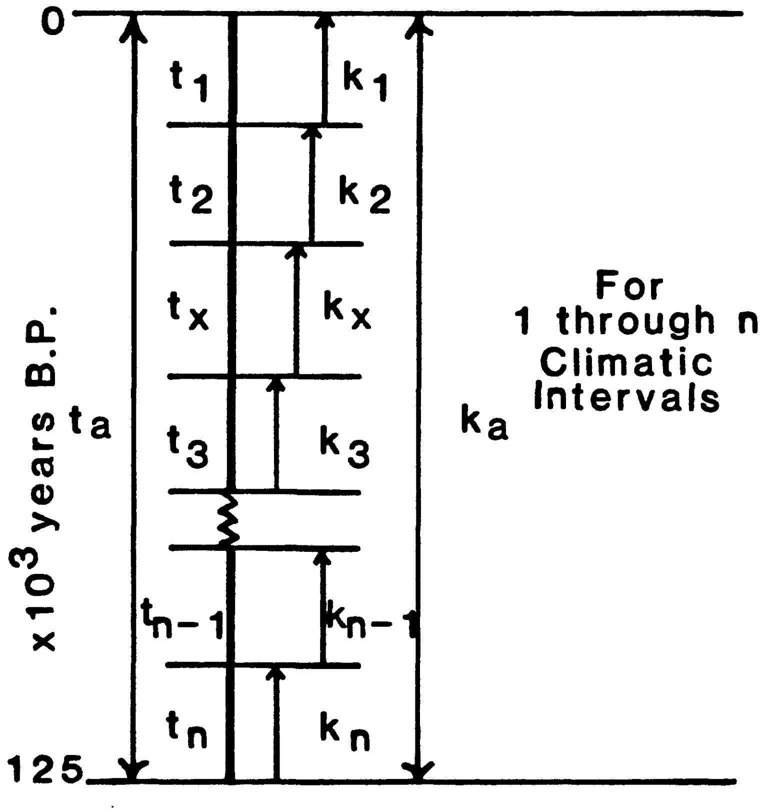

FIGURE 6.2. Schematic diagram illustrating the parameters used in equation 5 for evaluating past temperature histories. The EDT for the unknown interval x is predicted using k_x in place of k_1 in equation 3. See text for the definition of terms shown here. Any number of known and unknown intervals can be included, but the calculated unknown EDT's will be the average of all the unknown time intervals.

than today, (2) an early Wisconsin glacial episode (Happy) occurred prior to 65 ka ago, (3) during the mid-Wisconsin (Boutellier Interval, about 65-30 ka B.P.) climate was intermediate between colder, drier glacials and warmer, more mesic interglacial conditions, (4) the late Wisconsin glacial stade (Duvanny Yar) occurred between 30 and 14 ka ago, and (5) during the late Pleistocene/early Holocene (Birch Interval, 14-8.5 ka years B.P.) summer temperatures were somewhat warmer than during the remainder of the Holocene (compiled by Hopkins [1982]). Analyses of fossil beetles, pollen, and plant macrofossils from the inner Arctic Coastal Plain suggests that about 42 ka B.P., summer temperatures inland from the coast were lower, but within 2°C of modern temperatures (which are 11.6 ± 2.2°C) and that precipitation from June through August was reduced (Nelson, 1982, pg. 81). Nelson (1982) also argued that during the following 20 ka available moisture decreased and summer temperatures were up to 5-6°C lower than at present. These conclusions are corroborated by Carter (1981a, 1981b, 1982), who suggested that alluviation of the Ikpikpuk River between 42 ka and roughly 12 ka years B.P. was caused by sparse vegetation, an increase in eolian activity, and decreasing precipitation. Carter (1981b, p. 6) also has argued that the lack of remains of woody plants on the North Slope dating from 32 ka to 13.5 ka B.P. may reflect extreme moisture stress as well as lower summer temperatures. Sand wedges, rather than ice wedges, dominated the landscape, suggestive of very thin winter snowcover and an abundance of eolian sand (Carter, 1982). Despite the evidence of extreme aridity, glaciation limits were lowered in the Brooks Range during late Wisconsin time indicating that mean annual air temperatures were lower (Hamilton 1981). Balobeau (1978) and Barry (1982) maintain on other grounds that temperatures were depressed at high latitudes.

A major climatic shift on the North Slope is indicated by a dramatic increase in birch pollen in lake cores dated at about 13.5 ka to 8.5 ka B.P. (Hopkins, 1982) and by the presence of Populus beyond its present range between 11.5 ka and 7.3 ka (Hopkins and others, 1981). The so-called "birch zone" (Hopkins, 1982) is also marked by the earliest development of thaw lake deposits and the initiation of ice-wedge formation as opposed to sand-wedge development that prevailed during the full glacial climate of the preceding Duvanny Yar interval (Carter, 1981b, 1982; Hopkins, 1982; this study). Nelson (1982) used the presence of six beetle species, now confined to the boreal forest, to suggest that July temperatures at the inner edge of the coastal plain were about 2°C lower during the time of the birch zone than at present. Thus, the birch period represents a major change in both the moisture and temperature regimes of the North Slope with increased winter snowfall and warmer but still dry summers (Hopkins, 1982). These conditions were followed by the gradual establishment of modern vegetation and climate, tempered by moderate fluctuations, shortly after 8 ka years B.P. (Anderson, 1981).

To model these climatic intervals, assumptions are necessary concerning the conditions in permafrost, as they may have effected propagation in the annual temperature wave. The most important factors influencing the effective diagenetic temperatures in permafrost are (1) the thermal conductivity of the sediment (which has probably been continuously frozen since the mid-Pleistocene), (2) the depth and insulating value of snow cover, and (3) the amplitude of the annual surface temperature wave. The lack of winter snow cover during dry periods probably allowed greater winter ground cooling. Given the more continental situation of the coastal plain, summer temperatures might be thought to have been warmer, but this may not have been true. Sparse vegetation and the high albedo of eolian sand would have inhibited the penetration of summer radiation. The dry soils, however, may have had a lower heat capacity and warmed more quickly at given temperatures. It is clear from studies of D. Hopkins, L.D. Carter and myself that thaw lakes did not develop during the Duvanny Yar glacial period (30 ka to 14 ka B.P.). Because summer temperatures probably did rise above 0°C in the summer months, the lack of these features must indicate that the upper few meters of permafrost were severely dessicated.

Paleoclimatic models assuming mean permafrost temperatures or EDTs at depth for some intervals and calculating them for other intervals are depicted graphically in Figure 6.3. In each case, equation (5) is used to predict the temperature of the unknown (dashed) intervals. In all cases, the temperatures assigned to the intervals are constrained by the fact that an integration of all intervals must yield an average EDT of -14.0°C for the last 125 ka, as calculated earlier.

Case 1 is a simple model which partitions the last 125 ka years into only 6 broad climatic intervals. The pre-20th century meantemperature of -12°C is used to represent the last 8.5 ka years while a temperature of -9°C was chosen for the birch zone (8.5-14 ka years), representing a slight warming of climate, accompanied by inferred thicker, more insulating snowcover. The Boutellier interval (30-65 ka B.P.) is depicted as a period of intermediate climate, and assigned a temperature lower than at present but higher than that of full glacial intervals. Finally, the last interglacial episode is assigned a temperature slightly higher than today, as indicated by the presence of extralimital fossil marine mollusks in deposits of this age along the northern coast. The calculation for case 1 suggests that glacial-age permafrost temperatures (for the Duvanny Yar and Happy intervals) averaged roughly -16°C, or 4°C lower than today. Because the difference between mean annual ground temperature and mean annual air temperature can differ by as much as 1-6°C, depending on snow-cover (Gold and Lachenbruch, 1973), this permafrost temperature suggests past mean annual air temperatures between -17 and -23°C. Because winter snowcover is interpreted to have been thin during the glacial times, permafrost temperatures and mean annual air temperatures were probably similar.

Interpretations of global paleoclimates suggest that the period between 80 and 120 ka years B.P. was marked by oscillating,

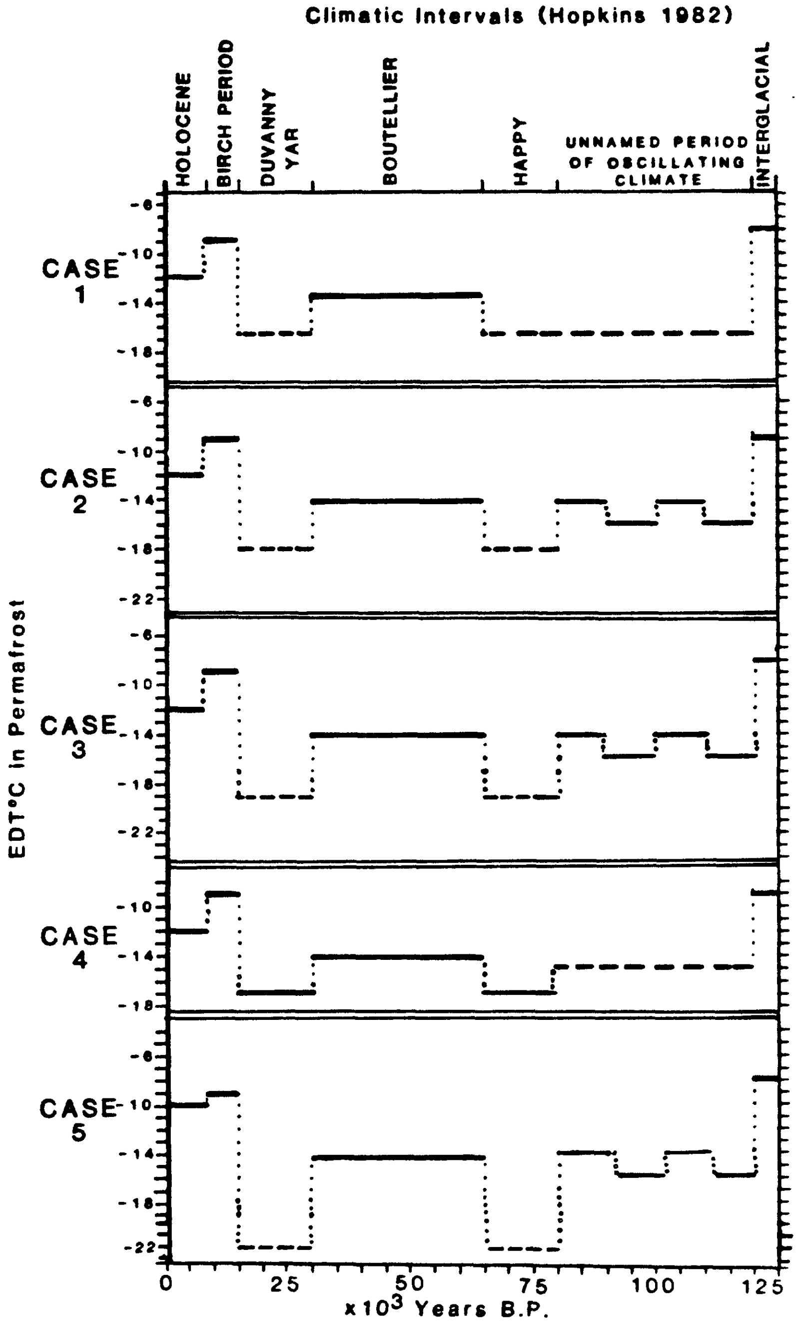

FIGURE 6.3. Schematic diagram illustrating five possible temperature models with glacial temperatures calculated from equation 5. Solid lines represent intervals of "known" climate, whereas dashed lines represent temperatures calculated for intervals of "unknown" climate. All temperatures indicated represent the EDT at depth in permafrost

though not extreme, climatic conditions (cf. Shackleton and Opdyke 1973); hence case 1 is probably oversimplified. In case 2, I have attempted to accomodte these fluctuations in climate with a more complex model by assigning intermediate warm and cold values for the oscillations of $\delta^{18}O$ stages 5a-d. This model effectively provides additional warmth to equation 5 and results in predicted permafrost temperatures (for the Duvanny Yar and Happy glacial intervals of -18°C. This value is colder than in case 1, and is similar to values measured recently in permafrost in the Dry Valleys of Antarctica where the current mean annual air temperature ranges between -18°C and -25°C (Decker and Bucher, 1982). The sensitivity of equation 5 to slight increases in temperature is illustrated in case 3, which is identical to case 2 except that the EDT for the last interglacial interval, assumed to have lasted 5,000 years, is increased by 1°C. This case results in a predicted temperature decrease during the Happy and Duvanny Yar periods, assumed to have had a total duration of 31 ka years, of almost 2°C, reducing permafrost temperatures to -19°C. An attempt to model modern temperature values during $\delta^{18}O$ stages 5a and 5c generated more warmth than could be compensated by any temperature depression during glacial periods.

In case 4, I have attempted to predict an average integrated temperature for the entire time period of $\delta^{18}O$ stages 5a-d (80 ka-120 ka B.P.) assuming permafrost temperatures during the Duvanny Yar and Happy Intervals similar to modern values measured at Alert, N.W.T. (-16.1°C at -15 to -18 m, Washburn 1980, p. 37). This reconstruction predicts an average temperature of -15.7°C for stages 5a-5d, a temperature almost as cold as those assumed for the glacial intervals in this model. Finally, in case 5 I have used the same thermal history as in case 2 but used a Holocene temperature average of -10°C, the modern permafrost temperature, rather than -12°C. This reconstruction predicts glacial temperatures lower than the other models, around -22°C. This case also illustrates the sensitivity of the equations and associated rate constants.

Conclusions

The temperature reconstructions outlined above are merely models, to be refined and tuned as additional paleoclimatic proxy data and information concerning permafrost temperatures under variable climates becomes available. It is not possible to chose which model may be most correct, however, the most complex models are probably most realistic. Data concerning the magnitude and duration of climatic variations during the Boutelliar period especially need to be assessed and added to the models. The most important conclusion to be drawn from this exercise is that the climate of the North Slope during glacial episodes was <u>much</u> colder than today. Very little warmth can be introduced into any part of the model and still accommodate the available data. Any additional warmth (for example, a warmer birch zone) imposed on the models must be compensated by more intense or more numerous cold periods.

It seems likely that conditions during the latter portion of $\delta^{18}O$ stage 5(a-d) were cooler than at present. Modelled glacial temperatures below -25° to -28°C for the Duvanny Yar and Happy intervals are believed to be unreasonable.

From these models it is possible to conclude that permafrost temperatures on the North Slope during glacial intervals probably averaged lower than -17°C, implying glacial mean annual air temperatures of -17° or lower. Moreover, it is doubtful that summer temperatures were as warm as today.

The conditions depicted here for the Wisconsin glacial periods are reminiscent of the harsh polar desert climates of northern Ellesmere Island, N.W.T., northeast Greenland, or the Dry Valleys of Antarctica. Possibly these areas could serve as useful modern analogs for reconstructing the glacial age environment of the Alaskan Arctic Coastal Plain.

CHAPTER VII

MARINE STRATIGRAPHY, GEOCHRONOLOGY, AND SEA LEVEL HISTORY OF THE GUBIK FORMATION

AIle/Ile ratios in fossil mollusks allow the correlation of disjunct stratigraphic sections across the Arctic Coastal Plain and provide a means of defining disconformities between temporally distinct but superposed marine units of similar lithology. In conjunction with detailed stratigraphic study, the aminozones outlined in Chapter III provide a relative chronology for interpreting the sea level history recorded by marine units of the Gubik Formation across the western coastal plain.

The Gubik Formation, by definition, is a lithostratigraphic unit that is distinguished on the basis of lithic character and stratigraphic position. Because the superposed marine units within the Gubik Formation are lithologically indistinguishable and disconformities and paraconformities are not always obvious, subdivision of this complex formation on lithologic criteria alone is inappropriate. Following the guidelines set forth in the new North American Stratigraphic Code (ACSN, 1983), I have adopted to subdivide the Gubik Formation into stratigraphic units using the logic of allostratigraphy but not the nomenclature (see discussion, Chapter 1).

In this study I have used aIle/Ile ratios, as a chemical criterion for recognizing disconformities and paraconformities between units, in addition to traditional stratigraphic methods to distinguish members of the Gubik Formation (otherwise "allomembers", see article 59 in ACSN, 1983).

The nomenclature of the Gubik Formation is currently being revised by U.S. Geological Survey personnel. Specific problems have come into resolution as a result of recent work by David Hopkins, L. David Carter, John Williams and myself. Revisions will eventually redefine the Flaxman Formation as a member of the Gubik Formation, for example. For the purpose of this dissertation, I have included the Flaxman deposits at Barrow as a member of the Gubik Formation (Table 7.1). Whether Pliocene and Pleistocene marine units that extend offshore in the Beaufort Sea should be included in the Gubik Formation has not been formally addressed.

The base of the Gubik Formation across the Arctic Coastal Plain west of Barrow is defined as the disconformity below the lowest marine unit where it rests on Papigak Clay or Cretaceous bedrock. The top is defined as the boundary between the highest marine sediment in a section and the overlying eolian sand, thaw lake sediment, or <u>in situ</u> peat. Temporally distinct marine units within the Gubik Formation have been given the rank of member to facilitate the recognition of units deposited at slightly different times. The boundary between members is defined as the basal transgressive unconformity of each superposed marine unit.

Table 7.1. Stratigraphy of the Gubik Formation, western Arctic Coastal Plain.

Aminozone (Defined in Table 3.5)		Lithostratigraphy		alle/Ile in Ha. $\bar{X} \pm 1\sigma$ Free	Total	Sea level m a.s.l.	Reference Sections	Marine Environment	Age Range B.P.
---	Pleistocene	GUBIK FORMATION	Flaxman member	---	---	≤7 m	Beaufort Sea coast	same as today	80 - 100 ka
1	Pleistocene	GUBIK FORMATION	Walakpa member	Not Detectable	0.014 ± .003	10 m	Sec. 36, 78, 79	same as today or slightly warmer	115-130 ka
2	Pleistocene	GUBIK FORMATION	Karmuk member	0.040 ± .052	0.038 ± .007	23 m	Sec. 20, 80 98	same as today	≈500 ka
3	Pleistocene	GUBIK FORMATION	Tuapaktushak member	0.50 ± .02	0.089 ± .013	>33 m	Sec. 20, 65	warmer than today, no sea ice at the coast	0.7 - 1.7 m.y.
4	Pliocene	GUBIK FORMATION	Killi Creek member	0.58 ± .08	0.149 ± .024	?	Sec. 24, 67	poorly known	1.7 - 2.2
5	Pliocene	GUBIK FORMATION	Nuvalik member	0.75 ± .07	0.235 ± .017	?	Sec. 25, 26, 41	poorly known	>2.2
	Miocene	"Papigak clay" / Nanushuk Group							
	Creta	Nanushuk Group							

Therefore non-marine sediments found between marine units are included with the marine unit below it. Names for each member were derived from the local names of small creeks which have eroded through the coastal bluffs and drain into the Chukchi Sea along Skull Cliff, and for a prominent cape in Kuk Inlet. This nomenclature is best applied below the highest marine limit. How surficial eolian deposits such as the "Foothills or Upland Silt" formally fit in the nomenclature of the Gubik Formation has not yet been resolved. The nomenclature proposed here is intended to supercede the nomenclature of Black (1964).

The reason for assigning names to the individual marine units of the Gubik Formation is to establish an objective, provincial stratigraphy. Although Hopkins (1967a, 1973) proposed a nomenclature for a sequence of transgressions that presumably affected all of the sea coasts of Alaska, the correlation of these events to the stratigraphy on the North Slope remains sketchy. The purpose of this chapter is to provide a summary of the sea level history of the Gubik Formation, describe the characteristics of each stratigraphic unit, assign them pronouncable local names, outline evidence concerning their absolute age, and cite type localities and reference sections for future study by other workers.

The Nuvalik Member

The oldest recognized marine unit of the Gubik Formation that is exposed along Skull Cliff is the Nulavik Member, named after the location of a campsite at the mouth of two small bays along Skull Cliff, about 37 km southwest of Barrow (lat. 71°01'N, 157°18°W, see Fig. 2.1a). AIle/Ile ratios determined from fossil *Hiatella* and *Astarte* from this unit comprise Aminozone 5 (Table 3.5).

Deposits of the Nulavik Member include basal cobbly gravel, cross-bedded ripply sand and silty sand, and interbedded silt and sand, typical of the gravel lag and inner-shelf lithofacies (Chapter II). Erratic cobbles are common at the base of the unit. The paleontology is little known.

The distribution of the Nulavik is patchy. It has been recognized only in sections along Skull Cliff where it rests as a discontinuous bed over Cretacous bedrock of the Nanushuk group or Papigak clay. The stratotype is designated as Sec. 25 and 26 (Plate I) on either side of Oyagatut Creek, 17 km northeast of Peard Bay (Lat. 70°51'N, 157°54'W, Fig. 2.1a). At this site, the deposit is composed of inner-shelf facies, its lower boundary at about 6 m a.s.l. defined by bedrock in Sec. 26 and Papigak clay in Sec. 25. Its upper boundary, at about 7.8 m a.s.l. in both sections, is defined by an abrupt unconformity that is overlain by a younger blue-silver clayey unit containing 5 to 10% dispersed pebbles and believed to be part of the Killi Creek Member.

The Nulavik Member was deposited in an arctic shelf environment when sea level was considerably higher than at present; the actual elevation is not known. Further study will be necessary to draw paleoecological interpretations from the unit.

The Killi Creek Member

Overlying the Nulavik Member is the Killi Creek Member, named after a small creek that intersects the Skull Cliff coast some 62 km southwest of Barrow (Fig. 2.1a). The disconformities above and below this unit bracket aIle/Ile ratios in Hiatella, Mya, Macoma, and Astarte that characterize aminozone 4 (Table 3.5).

Deposits of the Killi Creek Member are, like the Nulavik Member, patchy and discontinuously exposed along Skull Cliff. These deposits include inner-shelf facies of interbedded gravel, sand, silt, silty sand, and some gravel lag deposits. Erratics are also found at the base of the unit in some sections where there is an abrupt unconformity. Locally, bedding planes are folded and contorted within the unit, especially in Sec. 28 and 31.

The stratotype of the Killi Creek Member is designated as Sec. 24 and 67 (Plate I), adjacent sections exposed on either side of Tuapaktushak Creek approximately 30 km southwest of Barrow (lat. 70°54'N; long. 157°38'W, Fig. 2.1a). At these sites, the unit consists of inner-shelf facies composed of interbedded gray silt, clay, and medium sand with a lag gravel at its base. Erratic dropstones also occur near the base of the unit in Sec. 67. This gravel rests on a marked unconformity that delineates the top of the underlying Nulavik Member. The top of the Killi Creek Member is also marked by an unconformity. In Sec. 24, this unconformity is overlain by younger inner-shelf sediments of the Tuapaktushak Member; in Sec. 67 the unconformity is undulating and cut by sand wedges that penetrate into the Killi Creek sediments and are overlain by several meters of eolian sand.

The Killi Creek Member was deposited in an inner-shelf environment when the sea level was much higher than present. During this transgresssion, erratic cobbles were delivered to this portion of the Arctic Shelf. The height of sea level during this period is not known, but marine shells with aminozone 4 ratios were collected in marine beds at Sec. 94 (Fig. 2.1b) at a height of more than 20 m a.s.l. suggesting that the strand during deposition of the Killi Creek Member was somewhat higher. Little is known concerning the paleontology of this unit at Skull Cliff. A specimen of the extinct Astarte leffingwelli was found at one site. This unit also deserves more thorough study.

The Tuapaktushak Member

One of the most extensive marine units exposed along Skull Cliff is the Tuapaktushak Member, named after Tuapaktushak Creek located 30 km southwest of Barrow (Fig. 2.1a). Tuapak is the Eskimo word for "flat rocks"; and "tuapaktushak" is said to mean "big flat rocks" (Frank Akpik, Barrow, pers. comm., 1981). The name is derived from the abundance of large slabs of Cretaceous sandstone that litter the valley floor at the mouth of the creek. This unit is characterized by aIle/Ile ratios of aminozone 3 in Hiatella, Mya, Macoma, and Astarte (Table 3.5).

Deposits of the Tuapaktushak Member consist largely of inner-shelf lithofacies bound by a gravel lag facies at its base. These units include interbedded sand, silty sand, silt, and some clay, and thin pebbly beds are common. Erratic cobbles and boulders are sparsely dispersed throughout the unit but are most common at its base. More commonly than in any other unit, these beds are locally folded and contorted, often into isoclinal and overturned folds. Two extinct mollusks, Neptunea lyrata leffingwelli and Astarte leffingwelli, and one extinct ostracode, Rabilimis paramirabilis, are commonly recovered from these sediments. Three extralimital mollusks, including Natica (Tectonatica) janthostoma, that now live in the southern Bering Sea or in the coastal areas of the north Pacific, are also common. The occurrence of these mollusks suggest that coastal waters were warmer during deposition of the Tuapaktushak beds than they are today and that they may have been covered by sea ice only 2 or 3 months of the year.

The stratotype of the Tuapaktushak Member is Sec. 20, located only 0.9 km north of the Tuapaktushak Creek along Skull Cliff (lat 70°55'N; 157°38'W, Fig. 2.1a, Fig. 7.1). Sections 17 and 65 on either side of Sec. 20 are additional reference sections. In Sec. 20, the Tuapaktushak unit is exposed between 6 and about 12 m a.s.l. and consists of interbedded silty clay, sandy silt and medium-fine sand with a few thin gravel layers. The unit contains more clay at its base. At this site the sediments are isoclinally folded and inland, along the gully exposure, the folds continue with a wave length of less than 50 m. Some folds are overturned toward the north.

The lower boundary of the Tuapaktushak Member is marked by gravel on an unconformity over Cretaceous bedrock, the Papigak clay, or older marine members of the Gubik Formation. In Sec. 20, the unit overlies the Nulavik Member. The top of the Tuapaktushak beds is most commonly marked by a clear, but sometimes undulating, gravelly unconformity at the base of the Karmuk Member. Where the Karmuk beds are missing, the Tuapaktushak Member is overlain by younger eolian or thaw lake facies.

The Tuapaktushak Member was deposited across the Arctic Coastal Plain in an arctic shelf environment when sea level was at least 33 m a.s.l. and erratics were being delivered to the coast. Erosion during this transgression removed portions of the older marine units. Southern elements in the mollusk fauna suggest that coastal waters were slightly warmer than today and sea ice may have persisted during fewer months of the year, if at all.

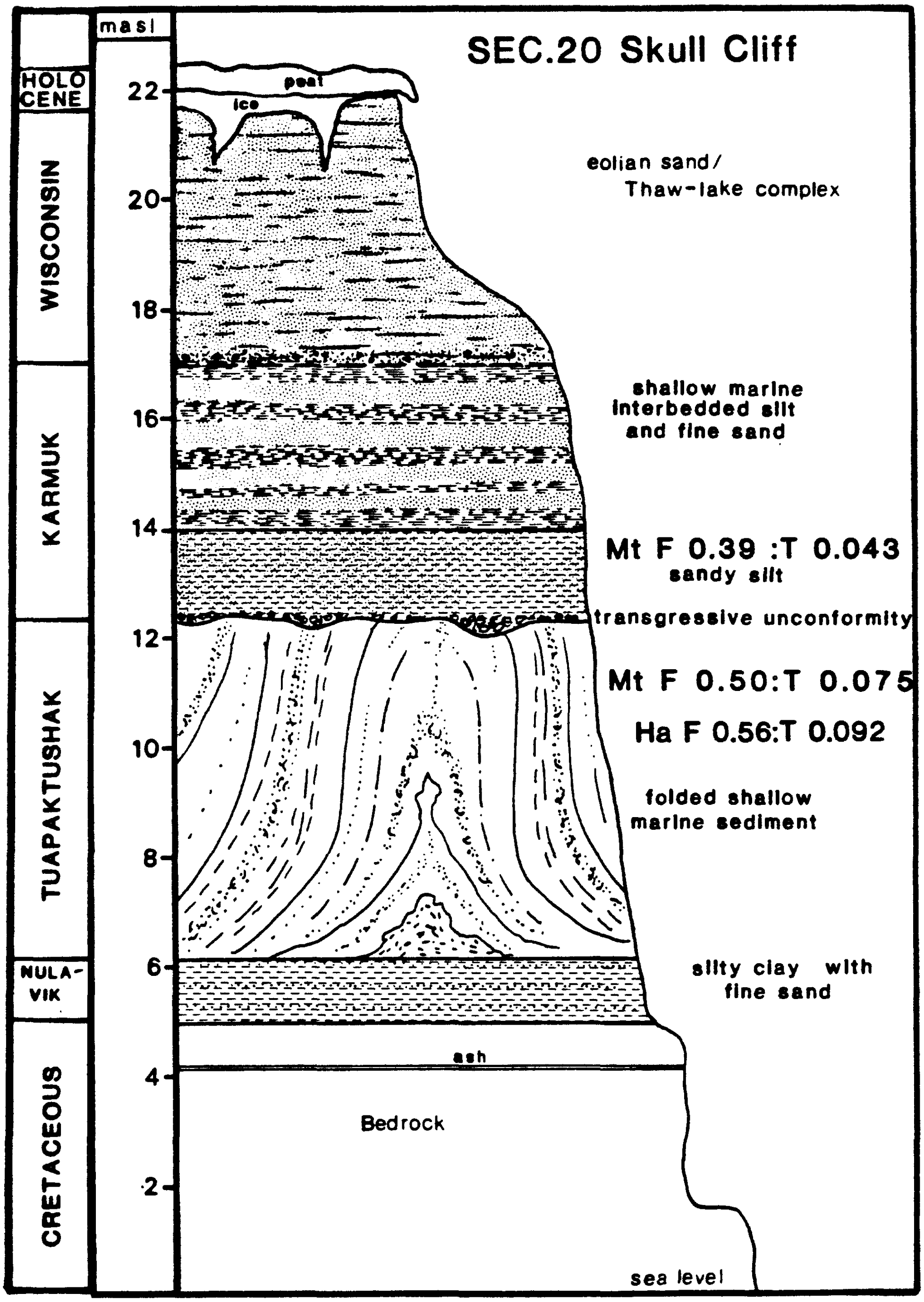

Figure 7.1. Section showing the stratotypes of the Tuapaktushak and Karmuk Members.

The Karmuk Member

The Karmuk Member is nearly as extensive as the Tuapaktushak Member in exposures along Skull Cliff. This unit is named after Karmuk Point which separates Wainwright Inlet from the Kuk Inlet (lat 70 °35'N; long 159°54'W, Fig. 2.1b). Sediments which compose a portion of this unit are exposed in high bluffs 2.5 km south of the Point. The boundaries of this Member across the coastal plain are delineated by unconformities that bracket aIle/Ile ratios in Hiatella, Mya, Macoma, and Astarte of Aminozone 2.

Deposits of the Karmuk Member consist of gravel lag, inner-shelf, barrier beach, and lagoonal lithofacies. It is lithologically the most diverse unit of the Gubik Formation. Lithologies include everything within the range between open-matrix gravel and fine clayey silt. Mollusk, ostracode, and foraminiferal faunas enclosed in these sediments are all similar to those now occupying the Arctic-shelf waters today; no extinct forms and few extralimital forms have been identified; exceptions include the remains of a single fossil sea otter near Walakpa Bay (Sec. 78, Fig. 2.1b) and two subfrigid species of ostracodes (Chapter V).

The stratotype for the Karmuk Member is exposed in Sec. 20 between 12 and 17 m a.s.l. (lat. 70°55'N; 157°38'W, Fig. 2.1a, also Fig. 7.1). Sec. 80, near Karmuk Point, and Sec. 98, on the Kokolik River, serve as reference sections (Fig. 2.1b, Fig. 4.17, and Fig. 4.15). In Sec. 20, the unit consists of inner-shelf lithofacies including interbedded silty sand, sandy silt and sandy clay. Marine shells in growth position are common in the sediments. The unit grades upward over 30 cm into more finely interbedded silty clay and fine sand with wavy crossbeds. In Sec. 80, in contrast, the Karmuk beds consist of interbedded and crossbedded coal-rich sand and gray clayey silt that becomes sandier upward. These beds are overlain by 7 m of coarse crossbedded gravel that mark the migration of an emergent barrier bar or spit over the site. Lagoonal sediments of Karmuk age are exposed in Sec. 98. Sediments thought to represent a tidal-delta (discussed Chapter II) are exposed in Sections 6 and 10 between 7 and ca. 15 m a.s.l. (11 m in sec. 6).

At the type section, the lower boundary over Tuapaktushak beds is marked by a gravelly shoreface erosional unconformity containing marine shells penecontemporaneous with the rest of the unit above. The top of the unit is marked by an unconformity covered by 40 cm of sandy pebble gravel which is overlain, in turn, by an eolian sand and peat. Only in a few sections do sediments of the Karmuk member directly overlie Papigak clay (e.g. Sec. 7, 10, and 32), and rarely does it overlie Cretaceous bedrock along Skull Cliff. Like the Tuapaktushak beds, Karmuk beds are also locally folded and contorted.

The Karmuk Member was deposited by a marine transgression that reached about 23 m a.s.l. Marine conditions during this transgression were similar to the modern conditions found offshore today.

The Walakpa Member

One of the youngest transgressions onto the Coast Plain is represented by deposits of the Walakpa Member, named after Walakpa Bay located 12 km southwest of Barrow (Fig. 2.1a). Sediments of this age are extensive in the bluffs on either side of Walakpa Bay. The boundaries of this unit bracket alle/Ile ratios of Aminozone 1 determined on valves of Hiatella, Mya, Macoma, and Astarte.

The Walakpa Member consists largely of beach and barrier bar lithofacies but includes some lagoonal facies. The beach deposits include interbedded and planar to horizontally crossbedded gravel and sand; the lagoonal deposits consists of horizontal, parallel-bedded dark brown silt and fine sandy silt interbedded with thin 2-3 mm laminae of medium sand containing Macoma balthica. Mollusk, ostracode, and foraminifera faunas from these beds are all similar to those found in the modern Arctic nearshore environment.

The stratotype of the Walakpa Member is designated as Sec. 36 located 2 km southwest of Walakpa Bay where a sequence of fossiliferous beach gravel up to 8 m a.s.l. is extensively exposed (lat. 71°8'N; 157°07'W, Fig. 2.1a). At this site several driftwood logs of spruce crop out from the frozen gravel, one of which yielded a radiocarbon date of greater than 36,000 years B.P. Similar sections include Sec. 46, 47, 49 and 50 all within a km southwest of Barrow (Fig. 2.1a) and Sec. 79, 81, and 82 near Wainwright (Fig. 4.19). Section 79 serves as an accessible reference section of barrier bar sediments. Lagoonal sediments of Walakpa age are exposed between 7 and 9 m a.s.l. in Sec. 78, some 5 km north of Walakpa Bay.

The Walakpa Member was deposited during a marine transgression that rose to somewhere between 8 and 10 m a.s.l. and inundated only the margins of the coast along the Chukchi Sea. As this sea level rise stabilized, a large cuspate spit and barrier bar system developed between present day Barrow and Teshekpuk Lake and a bay mouth bar developed at the mouth of the Kuk River estuary. The lower reaches of the Kugra River were flooded forming a large bay partitioned from the sea by an emergent spit extending out toward modern Asiniak Point. Marine conditions during this transgression were similar to modern conditions, although several mollusks that reach their northern limit in the Chukchi Sea expanded their ranges into the Beaufort Sea at this time.

Flaxman Member

Following the Pelukian transgression that deposited the Walakpa Member, a younger transgression occurred depositing fine-grain sediments and erratic boulders up to 7 m a.s.l. These deposits are known in older literature as the Flaxman Formation and are only exposed along the Beaufort Sea coast. Near Barrow, this transgression resulted in the development of curved spit, similar to the modern spit at Point Barrow today (Fig. 4.21).

Because no fossiliferous sediments of the Flaxman Member are present in the area west of Barrow, it is not possible to define an aminozone for this young unit. Nevertheless, shells from this unit would be expected to yield ratios slightly less than those of Aminozone 1 and probably close to results from modern specimens.

Late Pleistocene and Holocene Sediments

Throughout much of Wisconsin time, the landscape of the Arctic Coastal Plain was drier and colder than today. The Happy and Duvanny Yar glacial intervals from about 65-80 ka and 14 to 30 ka B.P., respectively, are both thought to have been very dry due to the remoteness of moisture sources to the south. These periods were also much colder, with mean annual air temperatures well below -17°C, or more than 4.5°C lower than at present (Chapter VI). The Boutellier period from about 30 to 65 ka B.P., in contrast, was a time of moderate climate, intermediate between colder, drier glacial conditions and warmer, more mesic interglacial conditions (Hopkins, 1982). Periodically, and perhaps continuously, during these intervals eolian sand and loess dominated the terrain forming dunes in areas where sand was plentiful (Carter, 1981a, 1981b). Elsewhere these deposits formed only a discontinuous sand sheet over the dessicated marine sediments and tundra surface (Williams, 1983a, 1983b). Eolian sediments are exposed at the top of many sections along Skull Cliff.

During the final waning stages of the late Wisconsin glaciation, much of Alaska experienced a dramatic warming between 8.5 and 14 ka B.P., an interval known as the birch period (Hopkins, 1982). By about 11,000 years B.P. this warming caused thaw lakes and ice-wedges to become active on the North Slope and by 8,500 years B.P. peat began to accumulate <u>in situ</u>. These deposits cap much of the marine sequence of the Gubik Formation across the western coastal plain.

Absolute Ages

The absolute age of each member of the Gubik Formation is based upon paleontological age estimates of mollusk and marine mammal collections; radiocarbon dates on wood, peat, and shells; uranium series and uranium-trend dates on shells and bulk sediments, respectively; and some preliminary paleomagnetic inclination data. This collection of data, in concert with kinetic model age estimates for isoleucine in <u>Mya</u> (Miller, 1982; in press.) provide the best age estimates bracketing each aminozone.

Paleontological Age Estimates

The edentulous left mandibular ramus of a fossil sea otter (<u>Enhydra</u>) was collected in 1975 by L.D. Carter, U.S.G.S., from fossiliferous marine sediments exposed near Ocean Point, along the Colville River (location in Fig. 5.2). From its morphology, Repenning (1983) deduces that this specimen represents a sea otter form that lived between 1.7 and 2.2 m.y. B.P. By coincidence this age estimate is similar to that proposed by Hopkins (1967a, p. 58) who considered the marine beds at Ocean Point to represent one of the Beringian transgressions based upon the occurrence of the extinct gastropod Neptunea lyrata leffingwelli and other faunal similarities. Supposed Beringian deposits on St. Georges Island are dated by K/Ar to 2.2 m.y. old but the stratigraphy of this site needs to be reevaluated (D.M. Hopkins, pers. comm. 1983). The Ocean Point faunas provide an important key to the age of the oldest units of the Gubik Formation.

In 1979, L. David Carter, U.S.G.S., began submitting mollusk samples from the Colville River region to the INSTAAR Amino Acid Geochronology Laboratory for analysis. These specimens have included numerous samples from the sea otter locality at Ocean Point which provide an important calibration point for the aminozones established west of Barrow. The most recent analyses (Table 7.2) suggests that the marine unit containing the sea otter is correlative with aminozone 4 (of Table 3.5) based upon a comparison of the Free and Total aIle/Ile ratios. The direct comparison of the ratios is justified because of the similarities in ground and air surface temperatures across the Coastal Plain (see Chapter I). If aminozone 4 is between 1.7 to 2.2 m.y. old, then aminozone 5 is inferred to represent a transgression that took place at sometime before or during the late Pliocene, i.e., > 2.2 m.y. ago.

Radiocarbon Dates

Radiocarbon assays on driftwood and shells from marine units of the Gubik Formation (Table 7.3) generally yield infinite dates, or finite dates that D.M. Hopkins and I consider to be minimum age estimates (e.g., those of Sellman and Brown, 1973). All of these samples were taken from marine units yielding aminozone 1 ratios, or are from deposits seaward of these units at Barrow. Of greatest significance is the fact that samples yielding infinite radiocarbon dates are barely hydrolyzed or epimerized above modern sample levels.

Radiocarbon dates from fluvial, eolian, thaw-lake, and ice-wedge-deformed sediments provide a chronology for recent geomorphic changes in the landscape. Basal dates on thaw-lake sediments and from ice wedge casts, in particular, are of interest because as found elsewhere on the Coastal Plain, all of these dates are 11 ka B.P. or younger; most are between roughly 8 to 11 ka B.P. This time period, referred to by Hopkins (1982) as the "birch period", marks a period of warmer conditions across Alaska as the landscape emerged from the glacial climate. This climatic change began about 13.5 ka B.P. and was followed by lake activity and the initiation of thaw accumulation of in situ peat.

Table 7.2. Extent of Isoleucine Epimerization in mollusks from marine beds at Ocean Point, collected and submitted for analysis by L.D. Carter, U.S.G.S.

Lab # AAL-	Genus	aIle/Ile Free	Total
2962	Hiatella	0.647 ± .006	.134 ± .003
2965	Mya	.617 ± .020	.120 ± .026
3124	Hiatella	.783 ± .011	.138 ± .012

Table 7.3. Radiocarbon dates on organic materials of the Gubik Formation, western Arctic Coastal Plain.

Field No.	Section	Lab I.D.	Material	Date	Description
				MARINE SEDIMENTS	
81 Akb 308	55	DIC-2568	Astarte	27,510 ± 500/550	Minimum date on marine transgression to ≥ 6 m a.s.l. Aminozone 1. Peard Bay
81 Akb 261 and 262	49	DIC-2569	Hiatella	31,200 ± 810/900	Minimum date on Aminozone 1 deposits deposits at Barrow. From the same deposits as W-380.
80 Akb 213	36	Beta-1766	Wood	>36,100 B.P.	Dates Pelukian marine transgression to 10 m a.s.l. Aminozone 1 at Walakpa Bay
--	--	W-380	Wood	>38,000	Dates Pelukian gravels at Barrow (Coulter and others, 1960)
--	--	I-3628	Wood	>39,900	Dates Pelukian gravels at Barrow (Sellman and Brown, 1973)
--	--	USGS-676	Wood	> 51,000	Dates Pelukian barrier bar near Teshepuk Lake (Carter and Robinson, 1981)
				FLUVIAL SEDIMENTS	
81 Akb 653	119	USGS-1489	Populus wood	9,370 ± 80	Terrace with Populus wood on Avak River. Sample analyzed for amino acids.
81 Akb 649	118	I-12926	Wood	21,140 ± 450	Wood in gravel terrace along Kokolik River.
81 Akb 634	107	I-12948	Peat	> 38,000	Peaty horizon over fluvial sediments 4 m above Utukok River
81 Akb 619	104	USGS-1425	Wood	45,900 ± 3100/1300	Wood associated with Larix in gravel terrace of Kokolik River. Minimum date.
81 Akb 616	99	USGS-1433	Wood	> 45,500	Wood in alluvial sand of Meade River

Table 7.3. (Continued).

Field No.	Section	Lab I.D.	Material	Date	Description
				THAW LAKE SEDIMENT	
80 Ahp 32b	61	USGS-1336	Peat	8,110 ± 80	Detrital peat at base of same thaw lake this 70 cm lower than USGS 1332.
81 Akb 424	66	I-12951	Peat	8,950 ± 150	Sandy thaw lake deposits--highest in sequence of two or three lakes
80 Ahp 32a	61	USGS-1331	Peat	9,350 ± 80	Basal organic layer in thaw lake deposits north of Kokolik Creek of Skull Cliff
81 Akb 393	64	I-12923	Peat	10,430 ± 150	Basal organic layer in thaw lake
81 Ahp 8d	64	I-12925	Peat	10,630 ± 160	Organics in sandier thaw lake superimposed onto the lake of I-12923.
				EOLIAN SEDIMENTS	
81 Akb 617	99	I-12875	Peat	5,600 ± 110	Dates colonization of cliff head dune along Meade River and retardation of sand accumulation
81 Akb 618	101	I-12874	Peat	7,880 ± 130	Lowest buried turf layer in sand/peat interbeds overlying terrace alluvium Small river between Kuk and Meade River
81 Ahp 62c	107	I-12950	Peat	> 40,000	Mottled medium sand with sparse pebbles peat clots--sand wedges in alluvium?
81 Ahp 62B	107	I-12949	Peat	> 40,000	Peat stringers in fine sand with rare ventifacted pebbles, in exposure along Utukok River
81 Akb 635	107	I-12981	Peat	> 40,000	Clots of silty peat in sand
81 Akb 636	107	I-12952	Peat	> 40,000	Clots of peat in sand 7.8 m above Utukok River
81 Akb 637	107	I-12979	Peat	> 40,000	Same as above

Table 7.3. (Continued).

Field No.	Section	Lab I.D.	Material	Date	Description
				ICE WEDGES	
81 Ahp 6c	20	I-12924	Peat	10,080 ± 150	Deepest tip of cryoturbated peat.
81 Ahp 6g	20	USGS-1420	Peat	11,710 ± 60	Well-bedded peat, in medium sand uplifted by ice wedge
81 Ahp 50c	98	USGS-1422	Wood	46,400 ± 1550/1300	Wood from ice wedge cast filled with Aminozone 2 lagoon sediments
81 Akb 605	98	USGS-1423	Peat	> 51,500	Ice wedge collapse below Aminozone 2 lagoonal sediments. Oldest ice wedges known on the coastal plain.

Uranium Series and Uranium-trend dating

Szabo (1982) has reported the only uranium series dates on mollusks from the Gubik Formation. Unidentified shell fragments from sediments represented by aminozone 3 collected by D.M. Hopkins (Sample S-2) in 1976 from the Skull Cliff area yielded a ^{230}Th age of 108,000 ± 8000 and ^{231}Pa age of 66,000 ± 6000. From the same area, a whale bone (Sample B-47) yielded an open-system ^{231}Pa age of 117,000 ± 25,000. Because of problems with the loss of ^{231} Pa (the daughter product of ^{235}U) in arctic settings, Szabo (1982, p. 5) regards only the ^{230}Th results "as reliable or at least minimum age estimates."

J. Rosholt, U.S.G.S., recently provided a uranium-trend date on bulk sediment from a unit represented by aminozone 2 at Skull Cliff (Sec. 20, Plate I, see also Fig. 7.1, 12.5-14 m a.s.l.). Uranium-trend dating is a technique that attempts to date sediments directly by determining an isochron that describes the distribution of daughter products following the migration of uranium through the sediment (Rosholt, 1980). The sample submitted yielded an age of 540,000 ± 60,000 B.P. (J. Rosholt, pers. comm., Oct., 1983). The unit from which the sample was taken is stratigraphically younger than the unit dated by Szabo (1982) suggesting that his shell dates may be, in fact, minimum estimates.

Paleomagnetic Data

Oriented sediment samples were collected from several sections to evaluate the inclination signature of the depositional remnant magnetism. All of the samples were demagnetized in an alternating field of 75ϕe; inclination data were then determined using a Schonstedt Spinner Magnetometer (Model SSLM-1A) under the direction of Dr. Edwin Larson, Department of Geology, University of Colorado.

Four samples collected from units represented by Aminozone 2 aIle/Ile ratios (83 Ahp 54B) all yielded strong steeply-dipping, positive inclinations suggesting deposition during a normal period hen the magnetic field was similar to today. Older sediments collected from units characterized by aIle/Ile ratios of Aminozones 3 and 5, however, yielded a mixture of steeply-dipping positive (normal) and negative (reversed) inclinations as well as a number of very low angle inclination values. This type of signature is somewhat typical of older sediments that have been overprinted with a strong chemical remnant magnetism parallel to the normal field (E. Larson, pers. comm., Sept. 1983). The low angle inclination values and strongly positive values may represent the vector resulting from such overprinting.

In order to evaluate the samples further, additional demagnetization using a stronger alternating field and thermal demagnetization will be necessary.

Summary of Absolute Ages

Based upon the absolute and relative age dating methods, it is possible to establish a range of ages within which the aminozones and corresponding stratigraphic units probably fall (Table 7.4). The absolute age and extent of amino acid diagenesis in fossil mollusks from aminozone 1 are of critical importance for defining age limits around the older units from the amino acid data. Aminozone 1 (Walakpa Member) which represents one of the last transgressions on the western Coastal Plain forms a prominent shoreline of gravel beach deposits up to 10-12 m a.s.l. that are continuously exposed above older transgressive marine units for almost 2.5 km on the south side of Walakpa Bay. Near Barrow, sediments of Aminozone 1 are composed of a barrier bar system exposed up to 10 m a.s.l. Beach sediments up to 10 m a.s.l. also partition Wainwright Inlet from the Chukchi Sea. This stratigraphic setting, coupled with the fact that the absolute age of the unit is beyond the range of radiocarbon, leads to the conclusion that these deposits represent oxygen isotope stage 5e of Shackleton and Opdyke (1973) when eustatic sea level reached ca. + 7 m a.s.l. at about 125,000. In Alaska, this transgression is known as the Pelukian transgression (Hopkins, 1967a, 1973).

The Flaxman member lies stratigraphically above the Aminozone 1/Walakpa Member. Based upon finite radiocarbon dates between 25 and 37 ka B.P. and stratigraphic relationships observed on the Beaufort Shelf, Hopkins (1982) maintains that this transgression took place between 80 and 100 ka B.P.

Karmuk Member sediments characterized by Aminozone 2 aIle/Ile ratios are estimated to be about 540,000 years old based upon a uranium-trend date of bulk sediment from Sec. 20. This age is consistent with the scanty paleomagnetic data that suggests these sediments are normally magnetized and less than about 730 ka B.P. The micro and macrofauna in this unit are also common, modern arctic forms.

The only absolute age available for Aminozone 3, the Tuapaktushak Member, is a ^{230}Th date of 108,000 on shells. Because of the skepticism surrounding this method (Kauffman and others, 1971; Szabo and others, 1982), this age is considered to be a minimum age estimate.

The absolute age of Aminozone 4, the Killi Creek Member, is based entirely upon the age estimate of 1.7 to 2.2 m.y. for a fossil sea otter by Repenning (1983) found in marine beds with shells at Ocean Point yielding similar aIle/Ile ratios. The absolute age of Aminozone 5, the Nuvalik Member, is unknown and can only be placed at greater than 2.2 m.y. and less than 3-3.5 m.y. B.P. when the Bering land bridge was breached and mollusks of Pacific origin migrated to the North Atlantic (Einarsson and others, 1967; Gladenkov, 1981).

Table 7.4. Summary of knowledge concerning the absolute age of each aminozone of the Gubik Formation.

Aminozone (Table 3.5)	Amino Age	Other Evidence for Age	Best Possible Age Range
1	125 ka	Beyond the range for radiocarbon; represents a sealevel event to about 10 m a.s.l.; associated with a warmer fauna than present in Beaufort Sea; correlation with the oxygen isotope record	115 - 130 ka
2	475 ka	Uranium trend date of 540,000 ± Normally magnetized	≈ 500 ka
3	1.0 - 1.4 m.y.	Maybe reversely magnetized Contains 2 extinct mollusks and one extinct ostracode (as outlined in Chapter V); uranium series date of > 108,000 at Skull Cliff	.73 - 1.7 m.y.
4	< 2.4	Correlative with a marine unit yielding a 1.7-2.2 m.y. old sea otter at Ocean Point; contains an extinct mollusk	1.7 - 2.2 m.y.
5	< 3.8	May be reversely magnetized; Most poorly preserved.	> 2.2 m.y.

Kinetic Model Age Estimates

Using equations that define the relationship between the extent of epimerization, age, and the effective diagenetic temperature (or thermal history) of a sample (defined in Chapter VI) it is possible to calculate the thermal term for samples in Aminozone 1. Using equation 4 from Chapter VI, an aIle/Ile ratio of 0.018 for Mya, and an age (t) of 125,000, the effective diagenetic temperature (EDT) for Aminozone 1 is calculated to be about -14.0°C. This temperature refers to the "temperature associated with an effective rate constant, which is the time averaged rate constant estimated by integration through a probable temerature history for a sample" (Wehmiller and others, 1977, p. 62-63). The integration of this history compensates for periods of high temperatures that accelerate the reactions more than the deceleration effected by an equivalent temperature decrease. Assuming that the proportion of warm and cold periods for the last 700,000 years is similar to that which occurred over the last 125 ka (c.f. Shackleton and Opdyke, 1973), age estimates for aminozones 2 and 3, corresponding to the Karmuk and Tuapaktushak members, can be made by solving equation 4 for t, such that,

$$t = \frac{\ln\left[\frac{1+D/L}{1-0.77}\right] - 0.0194}{1.77 \times 10^{[16.45-6141/T]}},$$

where t = Age of the fossil in years
D/L = aIle/Ile ratio in the Total Fraction
T = absolute temperature (°K) (assumed EDT)

Using -14.0° C, aminozone 2 is estimated at about 475,000 years B.P., an age that agrees well with the uranium-trend age estimate. Using the same EDT, the age of aminozone 3 lies within the range of 1.0 to 1.4 m.y., however, this estimate is much more spectulative because the thermal history that far back is so poorly known. Age estimates by this method for the older aminozones must also be regarded as highly speculative because so little is known concerning their thermal history. Calculations using -14.0° C for aminozones 4 and 5, for example, predict ages in the range of 2.4 m.y. and 3.8 m.y. B.P., respectively. On paleontological grounds it is thought that even the oldest deposits of the Gubik Formation post-date the development 3-3.5 m.y. ago of a seaway between the Pacific and Arctic oceans. Hence, the oldest date of 3.8 m.y. for Aminozone 5 is older than one would expect according to our present knowledge of the paleogeography. Paleontological evidence suggests an age of 1.7-2.2 m.y. BP for aminozone 4. The fact that these age estimates are at least close to reasonable values indirectly suggests that the high Arctic already was quite cold entering the latest Pliocene. This broad statement does not contradict the evidence presented in this study and elsewhere (Nelson, 1979) that some interglacial periods during the late Pliocene and early Pleistocene were warm enough to sustain a coniferous forest across much of the Arctic Coastal Plain and preclude the presence of arctic sea ice as it exists today. The fact that high arctic glaciation was underway in some manner by at least 2.5 m.y. B.P. is well demonstrated by the occurrence of ice-rafted debris in the north Pacific and north Atlantic oceans and by oxygen-isotope evidence (Shackleton and Opdyke, 1977; Shackleton and others, 1984, and references therein). Slightly warmer EDT is, perhaps caused by warmer interglacials during Pliocene/Pleistocene time, could be used to estimate ages greater than 1.7 m.y. but less than 3.0-3.5 m.y. B.P. for aminozones 4 and 5.

Summary

The Pliocene/Pleistocene Gubik Formatioin that mantles the Arctic Coastal Plain includes wedges of marine sediment that represent one of the most complete records of high sea level stands in Arctic North America. These marine sequences have been subdivided into five Members, each of which corresponds to a specific aminozone representing a single depositional time period when sea level was above its present level. The Nulavik and Killi Creek Members are only locally preserved along Skull Cliff and represent high sea level stands that took place >2.2 and between 1.7 and 2.2 m.y. ago, respectively. An early Pleistocene transgression to at least 33-36 m a.s.l. and represented by deposits of the Tuakaptushak Member took place between 0.7 and 1.7 m.y. B.P. Extralimital mollusk faunas in these deposits suggest that coastal conditions along northern Alaska were warmer than they are today and that the Arctic Ocean may have lacked perennial sea ice during a portion of the early Pleistocene. The Karmuk Member consists of marine sediments deposited when sea level reached about 23 m a.s.l. about 500 ka B.P. and is characterized by marine conditions similar to today. About 125 ka B.P. sea level reached between 8 and 10 m a.s.l. and beach and lagoonal sediments of the Walakpa Member were deposited. Marine conditions in the Chukchi Sea at that time were similar to today, although, several mollusk species that today reach their northern limits in the Chukchi Sea expanded their ranges into the Beaufort Sea.

The youngest Plesistocene transgression onto the coastal plain took place 80 to 100 ka B.P. and deposited the Flaxman Member. These sediments are preserved only along the Beaufort Sea coast. At Barrow, Alaska, this transgression resulted in the formation of a curved spit about 7 m a.s.l. that parallels the modern spit at Point Barrow. Eolian sand was periodically deposited and redistributed as a discontinuous sand sheet across portions of Coastal Plain throughtout the Wisconsinian. Finally, a dramatic warm interval, known as the birch period, initiated the active growth of thaw-lakes and ice wedges and by 8.5 ka B.P. *in situ* peat began to accumulate.

CHAPTER VIII

CORRELATIONS AND CONCLUSIONS

Stratigraphic relationships, amino acid ratios, micro- and acrofaunal characteristics, and radiometric dates provide a means f r rrelating the marine record of the Gubik Formation with evi of similar high sea level stands documented around Alaska and areas adjacent to Alaska. Further studies, particularly those us acid geochronology in areas of the Arctic, should strengthen or refine the correlations discussed here. Because the age of d cores from the Arctic Ocean (Herman, 1974, Clark and others, 1981) is now being seriously questioned (Sejrup and others, 1984), the study and correlation of coastal marine records may be especially critical to interpret properly the climatic evolution of the Arctic Basin during the late Cenozoic. Such studies will also improve upon what is known concerning the timing of major mar faunal migrations across Arctic Ocean between the Pacific and the Atlantic oceans.

Correlations across the Alaskan Arctic Coastal Plain

The somewhat uniform and extremely low mean annual ground and air temperatures across the Arctic Coastal Plain allow the direct correlation of marine units by amino acid geochronology. Studies by L.D. Carter (Carter and others, 1979; Carter and Galloway, 1982) of the Gubik Formation east of Barrow and along the Colville River document the existence of marine deposits that represent at least five high sea stands. Alle/Ile ratios on Hiatella and Mya from these units cluster similarly with ratios in the aminozones of the western Coastal Plain (L.D. Carter, unpublished[1]) suggesting they record sea level events recognizable across the entire coastal plain (Table 8.1). The principle difference between Carter's sequence and that exposed at Skull Cliff is that marine sediments correlative with those of the Karmuk Member have not yet been recognized in the Colville River region. Mollusks from marine sediments of the Colvillean transgression exposed at Ocean Point yield ratios similar to the Killi Creek Member. The extinct mollusks Neptunea lyrata leffingwelli and Astarte leffingwelli, are present as well as a fossil sea otter (Enhydra) believed to be between 1.7 and 2.2 m.y. old (Repenning, 1983). Based upon the entire mollusk fauna, Carter and Galloway (1982) proposed that these beds are correlative with deposits of the Anvilian transgression of western Alaska (Hopkins, 1967a).

Pollen collected from deposits of the Colvillean transgression at Ocean Point indicate that a coniferous forest

[1]Amino acid data of L.D. Carter completed under contract for the U.S. Geological Survey at the Amino Acid Geochronology Laboratory, INSTAAR, University of Colorado, 1979 through 1982.

Table 8.1 Preliminary correlation of marine strata of the Gubik Formation across the Alaskan Arctic Coastal Plain

		western coastal plain (this study)	eastern coastal plain (Carter & others, 1979, Carter & Robinson, 1981, Carter & Galloway, 1982)
Pleistocene	GUBIK FORMATION	Flaxman member	Flaxman Formation
Pleistocene	GUBIK FORMATION	Walakpa member	deposits of the Pelukian transgression
Pleistocene	GUBIK FORMATION	Karmuk member	no recognized equivalent
Pleistocene	GUBIK FORMATION	Tuapaktushak member	beds at Fish Creek
Pliocene	GUBIK FORMATION	Killi Creek member	deposits of the Colvillean transgression
Pliocene	GUBIK FORMATION	Nulavik member	beds on Milliveach Creek
Miocene		Papigak clay	Nuwok member of Sagavanirktok Fm.

dominated by spruce and a considerable amount of tree birch covered the Arctic Coastal Plain during deposition of the sediments (Nelson, 1979, 1981). Hardwoods and hemlocks apparently were not present in the coastal plain forest, and pines and true firs were rare. By analogy, Nelson (1979, 1981) claims that the overall environment was similar to modern-day Anchorage. This comparison suggests that mean annual temperatures were nearly 12° C higher and precipitation was perhaps 50 percent higher than the conditions that exist today on the North Slope (Nelson 1981, p. B10). Nelson (ibid, p. B10) concludes from this analogy that the Arctic Ocean was "largely ice free in the late Pliocene, and permafrost, if present, was discontinuous and probably limited to north-facing slopes."

Elsewhere in the area of the Colville River, cones of the extinct larch, Larix minuta, were found in alluvium that is younger than deposits of the Colvillian transgression at Ocean Point but older than the early Pleistocene marine sediments at Fish Creek (L.D. Carter, pers. comm., October, 1982). Larix minuta is also present and evidently formed treeline during an early Pleistocene interglacial recorded at Cape Deceit on the Seaward Peninsula (Matthews, 1974; Hopkins and others, 1981).

Marine sediments older than the deposits of the Colvillian transgression may be present in bluffs upstream from Ocean Point and are apparently present in exposures along Milliveach Creek. Shells yielding alle/Ile ratios similar to those in the molluskan Tuapaktushak Member have been collected from marine deposits along Fish Creek now assigned to the informally named Fish Creekian transgression (Marincovich and others, 1983) that also contain specimens of Littorina sitchana, an intertidal mollusk that is intolerant of sea ice or subfreezing air temperatures. The presence of this mollusk, in conjunction with the extralimital species from the Tuapaktushak beds suggests that the coastal areas of the Arctic Ocean were probably ice free throughout much of the year and that only seasonal ice may have existed offshore in the Arctic Ocean.

Some of the youngest marine deposits exposed on the Coastal Plain east of Barrow are represented by patches of gravel representing an ancient and extensive barrier bar system between Barrow and the Kogru River (Carter and Robinson, 1981). These deposits are thought to have been deposited during the Pelukian transgression and consequently form an extension of the gravel of the Walakpa Member exposed at Barrow. Several mollusk species including Mytilus edulis, Tellina lutea, Macoma balthica, and Siliqua patula that presently reach their northern limits in the Chukchi Sea extended their range into the Beaufort Sea during this transgression.

Though rare or perhaps absent on the Chukchi Sea coast, deposits of the Flaxman "Formation" are ubiquitous along the Beaufort Sea and are conformable on sediments of Pelukian age in some sections. This transgression is apparently older than a drastic reduction of sea level on the Beaufort continental shelf that occurred less than 40,000 years B.P. (Hopkins, 1982, p. 13). Although the absolute age is unknown, the Flaxman Formation is thought to represent a marine transgression only meters above present sea level that took place during the latter portion of oxygen isotope stage 5 (Hopkins, 1982).

Banks Island, N.W.T.

Vincent (1982; Vincent and others 1983; 1984) recently reported on the glacial and sea level history of unconsolidated deposits on Banks Island and the surrounding areas of the Canadian Arctic Archipelago. This area was inundated by glacial ice from the southeast at least three times during the late Cenozoic. Marine sediment representing 3 interglacial periods and interbedded with fossiliferous glacial tills and glaciomarine sediment in coastal sections bear evidence concerning the chronology of these events. Amino acid ratios on shells and wood were used to correlate disjunct stratigraphic sections.

The oldest interglacial bed on Banks Island, the Worth Point Formation, contains trunks of Larix laricina among other floristic elements that suggest than an open subarctic, forest-tundra environment once existed on Banks Island. Larix laricina is restricted today to the central Yukon River valley and the lower reaches of the Mackenzie River valley (see map in Hopkins and others, 1981). The age of the Worth Point Formation is not clearly known, but it is thought that the terrestrial beds that comprise it were laid down over a long period of time, perhaps since the Miocene and into the early Quaternary (Vincent and others, 1983, p. 1698). Sediments near the base of Worth Point Formation are normally magnetized and may record one of the normal events of the otherwise reversed Matuyama epoch; that is, either the Jaramillo, Gilsa, or Olduvai events (Vincent and others, 1984, p. 141). The upper part of the Worth Point Formation is reversely magnetized and is overlain by reversely magnetized marine sediments and till of the Duck Hawk Bluffs Formation. Vincent and others (1984) conclude that the Duck Hawk Bluffs Formtion must be older than 730 ka and was deposited during the Matayama epoch.

The Bruhnes/Matayama boundary is thought to occur within the Morgan Bluffs Formation, a deposit that records the next younger interglacial transgression--the Morgan Bluffs Interglacial. This transgression is found associated with evidence of a climate warmer than today based upon the occurrence of plant macrofossils, including needles, seeds, and buds of Larix sp., and insect remains of species that are now restricted to the mainland near treeline (Vincent, 1982; Vincent and others, 1983). Sea level during the Morgan Bluffs Interglacial stood 20 to 30 m higher than modern sea level suggesting to Vincent (1982) that this interglacial is correlative with the Kotzebuan transgression of Hopkins (1967a). The Morgan Bluffs Formation is overlain by the Big Sea marine sediments and the Baker till which comprise the Nelson River Formation. These sediments are also of normal polarity.

The youngest interglacial period represented in the area is the Cape Collinson Interglacial recorded by the Cape Collinson Formation that is believed to represent the last interglacial (125 ka B.P.). These sediments contain the remains of a tundra biota indicative of a climate slightly warmer than today as indicated by the presence of Betula nana L. and B. glandulosa Michx and several species of Coleoptera which are today restricted to treeline on the mainland. These beds are overlain, in places, by till and marine sediments of the early Wisconsin (?) M'Clure Stade and the late Wisconsin Russell Stade of the Prince of Wales Formation.

The correlation of interglacial sediments on Banks Island with those on the Arctic Coastal Plain is difficult at present because of discrepant opinions concerning the absolute age of the deposits. Published alle/Ile ratios on Hiatella arctica from several of the stratigraphic units on Banks Island (Vincent, 1982) are difficult to evaluate next to results from the Coastal Plain because the Banks Island samples were prepared using the '81 prep which produced anomously high Total ratios (see discussion of preparation methods, Chapter III). Miller (in press) has proposed a conversion factor of 0.78 ± .19 for Hiatella as a first approximation to convert '81 prep values to values one would expect from the presently used '76 prep. After multiplying by the mean of this factor, Hiatella arctica from Big Sea sediments (upper Nelson River Formation) yield ratios of 0.125 to 0.156 (formerly 0.16-0.20); shells from East Coast Sea sediments of the M'Clure Stade (Prince of Wales Formation) yield ratios of 0.031-0.070 (formerly 0.04-0.09); and the shells from Schuyter Point Sea sediments of the Russell Stade (upper Prince of Wales Formation) yield ratios of 0.016 (formerly 0.02).

Because the modern mean annual air temperature of -14° C on Banks Island is only slightly lower than at Barrow (-12.6°C) one might at first expect that deposits of similar age from the two areas might yield similar alle/Ile ratios. To the contrary, what is apparent is that there is a large discrepancy in the absolute age of similar ratios from the two areas. For example, the shells yielding ratios of 0.016 from the Schuyter Point Sea sediments also yielded radiocarbon dates of 10.2 and 11.2 ka B.P. (Vincent, 1982). Higher ratios of 0.031-0.070 from the East Coast Sea sediments are thought to post-date the early Wisconsin M'Clure Stade. In contrast, shells with a lower ratio of 0.014 from the Walakpa Member of the Gubik Formatioin are from deposits yielding radiocarbon dates on wood of > 36 and > 38 ka, and thought to be 125 ka old. This difference emphasizes further that ground temperatures across the Arctic Coastal Plain were, in fact, low enough to inhibit much fossil diagenesis througout the Wisconsin (see Chapter VI).

Discrepancies are also apparent in the data from older stratigraphic units. The ratios of 0.125 to 0.156 from the Big Sea sediments of Banks Island are thought to be younger than the Bruhnes/Matayama boundary whereas Killi Creek beds characterized by similar ratios in the Gubik Formation are thought to be of late Pliocene age.

If the age estimates of the lithostratigraphic units on Banks Island are correct, then discrepancies between Banks Island and the Alaskan Arctic Coastal Plain might be explained by differences in the history of ground temperatures in the two areas. It is important to emphasize that the stratigraphy on the Arctic Coastal Plain records a eustatic history, whereas, the stratigraphy on Banks Island primarily records a glacioisostatic history. Although Banks Island is today colder, the presence of glacial ice for relatively long periods of time (with basal temperatures around 0° C) and the persistence of isostatically, as well as eustatically, induced marine cover (with temperatures at 0° C or slightly lower), may have radically accelerated the racemization reactions. In contrast to Banks Island, the sediments on the Arctic Coastal Plain have been emergent and exposed to temperatures probably less than -10° C throughout most of the Quaternary. If we assume that the East Coast Sea sediments with ratios of 0.031-0.071 are, in fact, early Wisconsin in age (e.g., ca. 70 ka) then the effective diagenetic temperature (EDT) experienced by these sediments estimated from a mean ratio of 0.05 is close to -2.7° C (using the temperature equation for Mya which epimerizes at a rate similar to Hiatella discussed in Chapter VI). This suggests that since deposition, ground temperatures within the East Coast sea sediments have been more than 8° C warmer than current permafrost temperatures in the area. This calculation does not seem completely reasonable since the collection site for these samples ("surface of 36 m a.s.l. delta.... located 25 km south of Jesse Harbor"; Vincent 1982, Table III) must have been subaerially exposed since the mid-Wisconsin. The site was apparently some 11 m above the late Wisconsin Schuyter Point Sea that reached a height less than 25 m a.s.l. (Vincent, 1982, fig. 9) and was outside the ice margin of the Russell Stade. Additional amino acid data will need to be assembled from Banks Island before thorough conclusions can be made.

Using characteristics other than amino acid ratios, it remains tempting to suggest possible correlations between Banks Island and the Arctic Coastal Plain (Table 8.2). Based upon the extremely warm character of the Worth Point Formation and its estimated Pliocene age, I proposed that this interglacial or preglacial unit may be equivalent, in part, to the Killi Creek Member of the Gubik Formation (Colvillian transgression of Carter). Future work on the paleoenvironmental aspects of the older Nuvalik Member may also suggest that this interglacial was also quite warm and equivalent to some portion of the Worth Point Formation.

The Morgan Bluffs Formation also represents an interglacial episode warmer in character than those of the middle or late Pleistocene. The warmth of this interglacial coupled with fact that its associated deposits contain Larix sp. broadly suggests that the Morgan Bluffs Formation may be correlative with the Tuapaktushak Member of the Gubik Formation. Sea level is thought to have stood 20-30 m higher during the Morgan Bluffs Ingerglacial; similarly, sediments of the Tuapaktushak member are found up to

Table 8.2. Possible Correlation of Lithostratigraphic units from western Arctic Coastal Plain and Banks Island.[1]

Gubik Formation Alaskan Arctic Coastal Plain (this study)		Banks Island (Vincent, 1982)
	--------	Schuyter Point Sea (0.016)
	--------	East Coast Sea (0.031-0.071)
FLAXMAN M.	?	tills
	--------	Pre-Amundsen Sea
WALAKPA MEMBER	------------	CAPE COLLINSON FORMATION
(0.014)		Big Sea Sediments (0.125-0.156)
KARMUK -- ?		till
(0.038)		Pre-Thomsen Sea Sediments
TUAPAKTUSHAK M.	----- ? ----	MORGAN BLUFFS FORMATION
(0.089)	----	Post Banks Sea Sediments
additional interglacial		till
		Pre-Banks Sea Sediments
KILLI CREEK M.	-------------	WORTH POINT FORMATION
(0.149)		
	----- ? ----	
NUVALIK M.		
(0.235)		

[1]Numbers in parentheses are mean Total aIle/Ile ratios in *Hiatella arctica*. Banks Island ratios have been converted to 76b prep results.

33-36 m a.s.l. Alternatively, the Morgan Bluffs Formation may record an older warm interglacial episode not recorded on the Coastal Plain between the Tuapaktushak and Killi Creek Members. Cones of Larix minuta, an extinct species, have been found in alluvium of such an intermediate age on the Coastal Plain (Carter, pers. comm., Oct., 1982) and this same species exclusively formed treeline during the early Pleistocene at Cape Deceit on Seward Peninsula. Hopkins (pers. comm., Nov., 1983) claims that the Cape Deceit section may well be of intermediate age between the Tuapaktushak and Killi Creek sediments.

The Cape Collinson Formation records an interglacial episode characterized by a climate slightly warmer than today. Based upon its stratigraphic position, this terrestrial unit is thought to represent the last interglacial episode. If so, then it is correlative with the Walakpa Member of the Gubik Formation (Pelukian transgression). Interestingly, marine deposits of this age have not been reported from Banks Island. Moreover, the stratigraphy of Banks Island lacks any record of middle Pleistocene interglacials such as that recorded by the Karmuk Member of the Gubik Formation.

The erratic boulders that characterize the early Wisconsinan Flaxman "formation" along the cost of the Beaufort Sea are thought to have been delivered to the area, at least in part, by icebergs calving from an ice shelf in the Amundsen Gulf (Roderick, 1975). Ice surrounded Banks Island through the Amundsen Gulf and M'Clure Strait during the early Wisconsinan M'Clure Stade and presumably provided the source for much of the Flaxman material. Although the absolute timing of events is not known, the Flaxman Member (Flaxman Formation) is probably correlative with deposits in the lower half of the Prince of Wales Formation.

Refinement of the correlations proposed here will require further work with the paleontology of the Banks Island stratigraphy coupled with further amino acid analyses of in situ mollusks from available marine units.

Western Alaska

The transgressive marine units of the Gubik Formation are, in part, 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, 1967a). More recently, amino acid analyses (by G.H. Miller, unpub., 1976) of samples of Hiatella arctica and Mya truncata provided by D.M. Hopkins from the Nome Region shed a chemical perspective onto these correlations (Table 8.3). The samples analyzed illustrate the continuous progression of isoleucine epimerization with aIle/Ile ratios approaching racemic equilibrium in the Free fraction of the oldest unit, the Beringian transgression. The higher absolute values of the ratios at Nome reflect the warmer effective diagenetic temperatures at that latitude, hence a more rapid racemization rate. The current mean

Table 8.3. Aminostratigraphy of marine transgressions at Nome, western Alaska (CMAT, -4.7° C).[3]

Unit	Species[1] analyzed	No. of valves	aIle/Ile[2] FREE	TOTAL (81 prep)	TOTAL[4] (76 prep)
Pelukian --	Ha	2	0.16 ± 0.001	0.040 ± 0.002	0.031 ± .002
	Mt	4	0.28 ± 0.03	0.042 ± 0.005	0.026 ± .005
Kotzebuan--	Mt	2	0.46 ± 0.04	0.093 ± 0.001	0.057 ± .005
Anvilian --	Ha	3	0.90 ± 0.02	0.47 ± 0.03	(little or no conversion required for samples > 0.4)
Beringian--	Ha	3	1.04 ± 0.03	0.56 ± 0.07	
	Mt	2	0.94 ± 0.03	0.56 ± 0.04	

[1]Ha, _Hiatella arctica_; Mt, _Mya truncata_;

[2]Values represent the mean ±δ for peak-height measurements.

[3]Data from G.H. Miller and D.M. Hopkins, unpublished.

[4]Conversion factor of 0.78 for _Hiatella_ and 0.61 for _Mya_ taken from Miller (in press).

annual air temperature at Nome is -4.7° C, with a July mean of about +10° C.

The correlation of high sea level stands at Nome and along the western Arctic Coastal Plain is partially based upon stratigraphic position but is strongly reinforced by paleontological elements outlined elsewhere (Hopkins, 1967a) (Table 8.4 this study). Theoretical age estimates can be calculated from the amino acid data assuming linear kinetics of the isoleucine epimerization reaction up a Total ratio of about 0.3 (Wehmiller and Belknap, 1978; Miller, 1982). For example, an age of 125,00 years B.P. for the Pelukian transgression at Nome (ratio of 0.026 ± .005 for Mya) requires an effective diagenetic temperature (EDT, Chapter VI) below the ground surface between -9° and -12° C with a mean of ca. -10.4° C for the entire time period. If we assume that the climatic cycles of the last 500 k to 700 ka have experienced the same proportion of warm and cold intervals as has the last 125 ka, then -10.4° C represents at least a first approximation of the EDT throughout that time period. Hence, an aIle/Ile ratio of 0.057 ± .005 on shells derived from sediments thought to be equivalent to the Kotzebuan transgression at Nome (Table 8.3) predicts an age between 350 ka and 430 ka B.P. Even if we consider the possible temperature range of -9° to -12° C, the age estimates still fall within the range of 290 k to 550 ka B.P. This age range based on the amino acid data includes the K/Ar date of 320 ka ± 70 ka on a lava flow directly beneath sediments of the Einahnutan transgression (now known to belong to the Kotzebuan transgression, Miller and Hopkins, 1980) on St. Paul Island of the Priblof group (Hopkins, 1973), but is older than uranium-series dates of 170 k and 175 ka on shells from the type locality in Kotzebue Sound (Blanchard, 1973).

This principal can also be applied to samples from the North Slope. As demonstrated in Chapter VI, an age of 125 ka B.P. for the Walakpa member requires an EDT of about -14° C. This diagenetic temperature predicts an age for the Karmuk member of ca. 475 ka, an age slightly younger than the uranium-trend date of ca. 540 ka B.P. on these sediments but within the age range predicted by the amino acid data for the Kotzebuan transgression. Hopkins (1973) however, believes that the Kotzebuan transgression is closer to 250 ka B.P. Until further evidence of other middle Pleistocene transgressions can be found, my present bias is to suggest that the Karmuk Member was deposited during the Kotzebuan transgression and that this transgression occurred sometime between 250 k and 500 ka B.P.

I recognize three pre-middle Pleistocene events at Skull Cliff, recorded by sediments of the Tuapaktushak, Killi Creek, and Nulavik Members, whereas Hopkins (1967, 1973) recognizes only two events at Nome, the Anvilian and the Beringian transgressions. Based primarily upon their stratigraphic position and the relative differences in the AIle/Ile ratios from shells in these deposits, I propose that the Killi Creek Member was deposited during the Anvillian transgression and the Nulavik Member was deposited during

Table 8.4. Aminostratigraphic correlation of late Cenozoic marine units across the Bering and Chukchi seas.

Estimated Absolute Age (m.y.)	Kamchatka, U.S.S.R.	Chukotka, U.S.S.R.	Nome, W. Alaska	Coastal Plain, NW. Alaska
0.08-0.100	-------------	--------	--------	Flaxman Member
0.125	Attarmen beds	Val'katlen	Pelukian	Walakpa Member
0.150	-------------	Kresta	--------	-------
0.25-0.50	Karagin beds	Pinakul'	Kotzebuan	Karmuk Member
?	Upper Olkhovaya (?) suite	(older	---------	Tuapaktushak Member
1.7-2.2	Tusatuvayam (?) beds	Pinakul'	Anvilian	Killi Creek Member
2.5-3.5	--------------	localities?)	Beringian	Nuvalik Member
>3.0	Lower Olkhovaya suite		---------	-------

the Beringian transgression. Moreover, it is possible that deposits of an early Pleistocene transgression, correlative with the Tupaktushak Member on the North Slope were subsequently eroded or have not yet been recognized in the coastal stratigraphy at Nome and Kotzebue.

Kamchatka and Chukotka Peninsulas, USSR

Sequences of marine beds representing Pliocene and Pleistocene transgressive events have been described from the western shores of the Bering and Chukchi seas on Kamchatka and Chukotka 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 stratigraphic units were analyzed to determine the progression of amino acid diagenesis in the series (Table 8.4). The youngest units analyzed are the Val'katlen Beds of Chuktoka Peninsula and the Attarmen Beds of Kamchatka Peninsula, (Fig. 8.1), both believed to mark the last interglacial high sea level event (Khoreva, 1974; Petrov, 1976, 1982). The oldest unit analyzed from Kamchatka, the lower Ollkovaya Suite of late Pliocene age, is racemic (that is, at chemical equilibrium) in the FREE fraction with an aIle/Ile ratio of 1.38 and nearly racemic in the Total hydrolysate. The percentage of free amino acids in shells from these beds also reflect a continuous increase with age (Fig. 8.2).

The generally the higher aIle/Ile ratios and more extensive hydrolysis in mollusks from the Kamchatka deposits reflects the warmer thermal regime affecting Kamchatka Peninsula (CMAT -0.9° to -15.°C with a July mean of about +11°C), particularly the presence of a maritime influence which persisted during interglacial and interstadial episodes but may have been lost or less intense during glacial episodes. Sancetta and Robinson (1983) have shown that sea ice persisted up to 9 months of the year without significant summer warming across the northern Aleutian Basin during the late Wisconsin maximum. Present temperatures along the Siberian shore average 2 - 6° C lower than the Alaskan coast at points of similar latitude with increasing divergence southwestward (Fig. 8.3). All of the sites sampled in this study, with the partial exception of the Kamchatka sites, lost most of their maritime influence during the glacial episodes of low sea level and presumably experienced dramatically lower mean annual temperatures (e.g., Brigham and Miller, 1982, 1983).

A possible correlation of transgressive deposits along Kamchatka and Chukotka peninsulas with those along the Alaskan coast is suggested by the amino acid data in Table 8.5 and Fig. 8.4. This scheme varies in detail from those previously published (Hopkins and others, 1965; Hopkins, 1973; Gladenkov, 1981; Petrov, 1966, 1982), all of which include the recently retracted Einahnutan transgression of western Alaska (Miller and Hopkins, 1980; see Chapter I).

Of the samples analyzed from either side of the Bering and Chukchi seas, the aIle/Ile ratios determined on shells from the

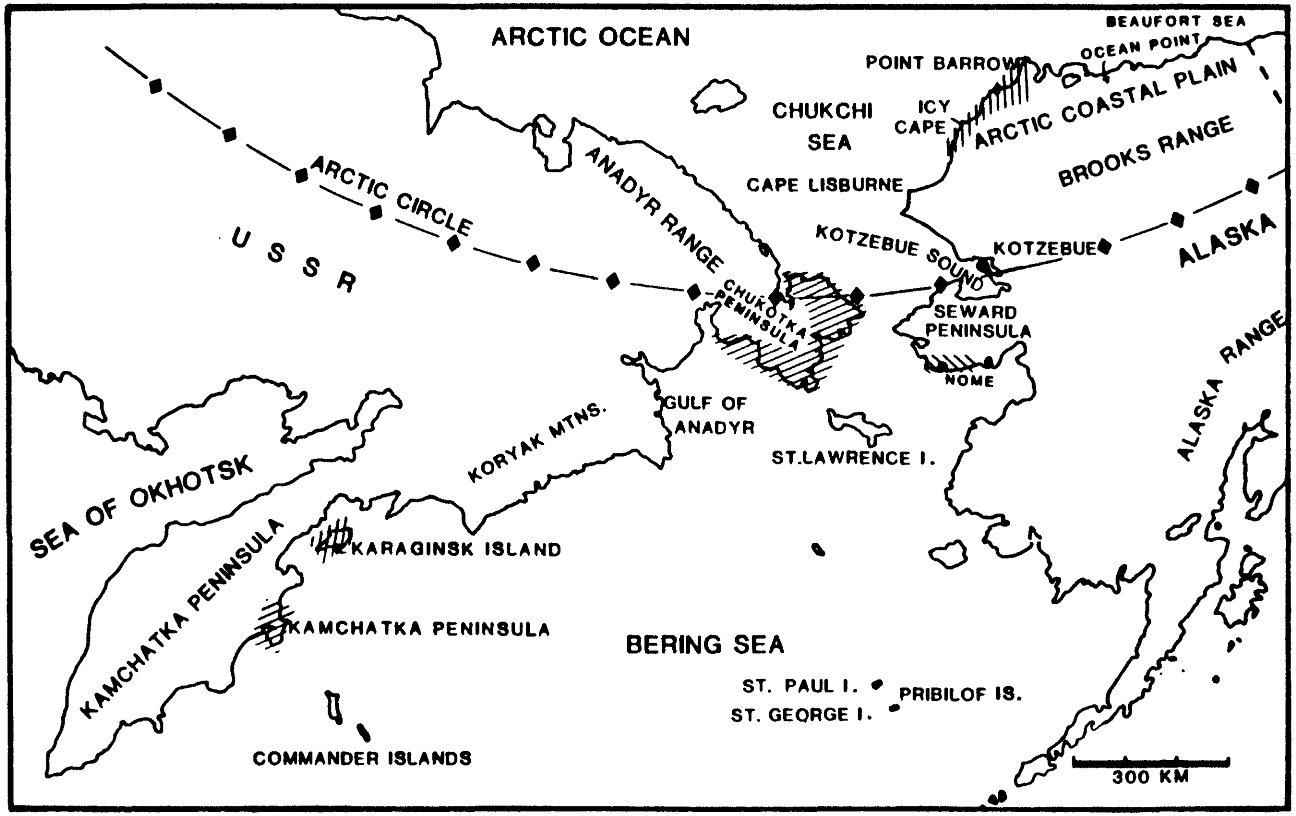

Figure 8.1. Map of northeastern Siberia and Alaska showing localities mentioned in the text. Sample localities (shaded) include the regions of northwestern Arctic Coastal Plain and Nome, Alaska, Karaginsk Island and Kamchatka Peninsula, Kamchatka, and the coasts of Anadyr, Kolyuchin and Lawrence bays, Chukotka Peninsula.

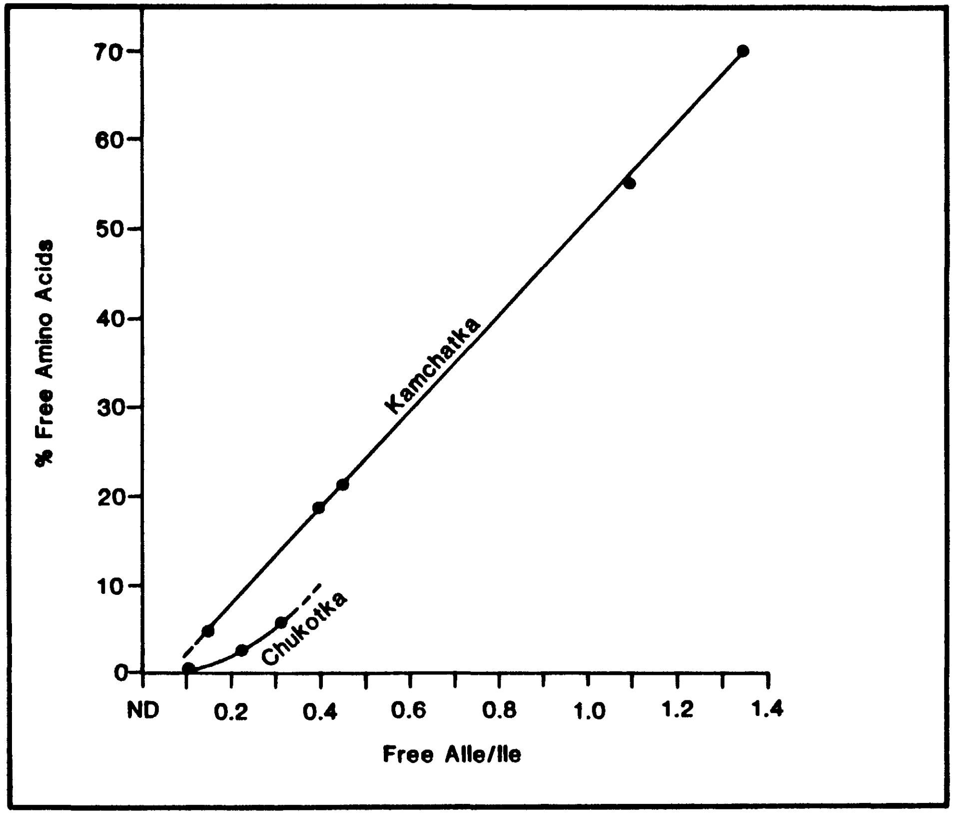

Figure 8.2. The hydrolysis of peptide bonds causes the gradual increase in the percentage of free amino acids. This progress is well illustrated in the Siberian samples.

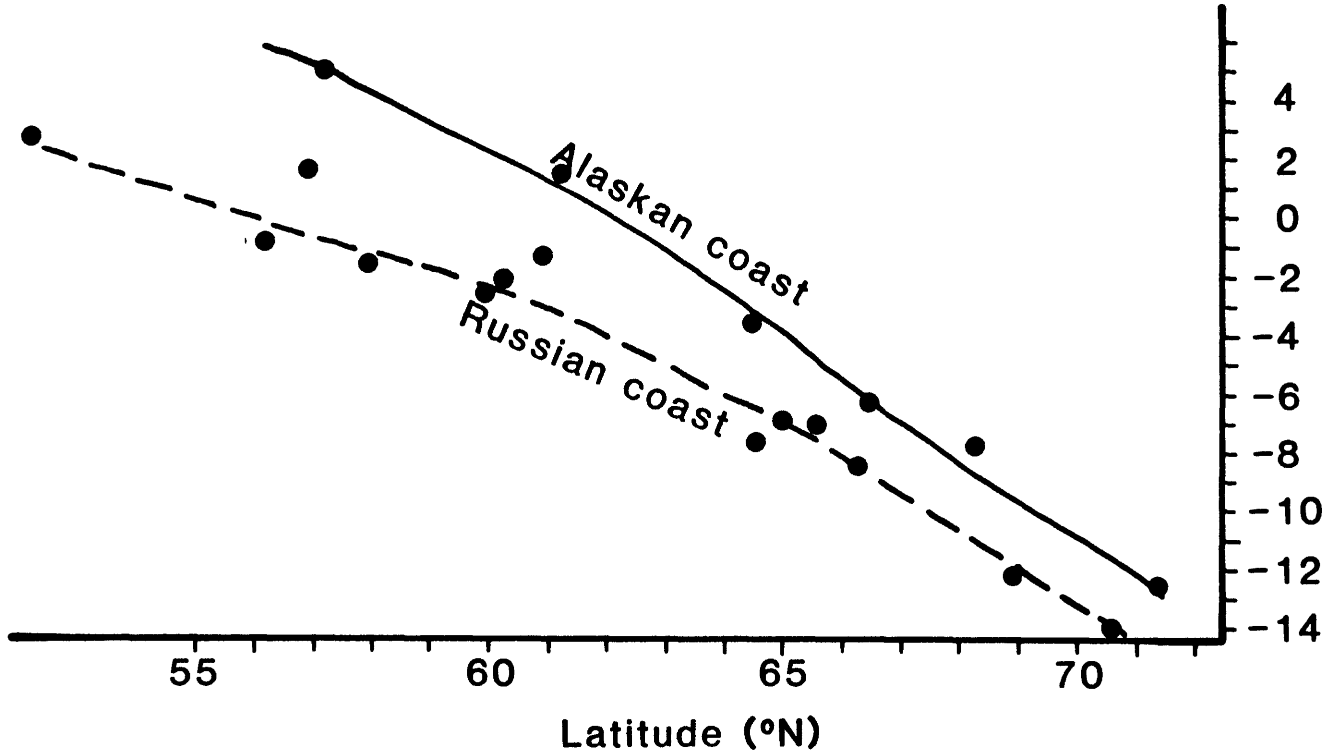

Figure 8.3. Diagram illustrating the contrast in current mean annul air temperatures (CMAAT) at similar latitudes on the Alaskan and Siberian coasts ofthe Berig and Chukchi seas. Siberian coastal temperatures are 2° to 6° colder than temperatures at the same latitude on the Alaskan coast.

Table 8.5. Aminostratigraphy of some marine transgressions on Kamchatka and Chukotka Peninsulas, U.S.S.R.[3]

Unit	Species[1]	No. of valves analyzed	aIle/Ile[2] FREE	TOTAL
		Chukotka Peninsula		
Val'katlen Beds	Mt	3	0.17 ± .011	0.018 ± .002
Kresta Suite	Ha.	1	0.23	0.022
	Mt.	2	0.22 ± .02	0.020 ± .002
Pinakul Suite	Ha	1	0.26	0.031
	Mt	1	0.21	0.035
		Kamchatka Peninsula		
Attarmen Beds	Ha	1	0.17	0.038
	Mt	1	0.14	0.027
Karagin Beds	Mt	3	0.43 ± .05	0.115 ± .013
Upper Olkhovaya Suite	Mt	3	0.44 ± .01	0.22 ± .012
Tusatuvayam Beds	Mt	2	1.16 ± .03	0.577 ± .004
Lower Olkhovaya Suite	Ha	2	1.38 ± .03	1.04 ± .06
	Mt	2	1.28 ± .03	1.04 ± .05

[1]Ha = Hiatella arctica, Mt = Mya truncata
[2]Values represent the mean ± 1δ for peak height measurements.
[3]Data from Brigham, 1982.

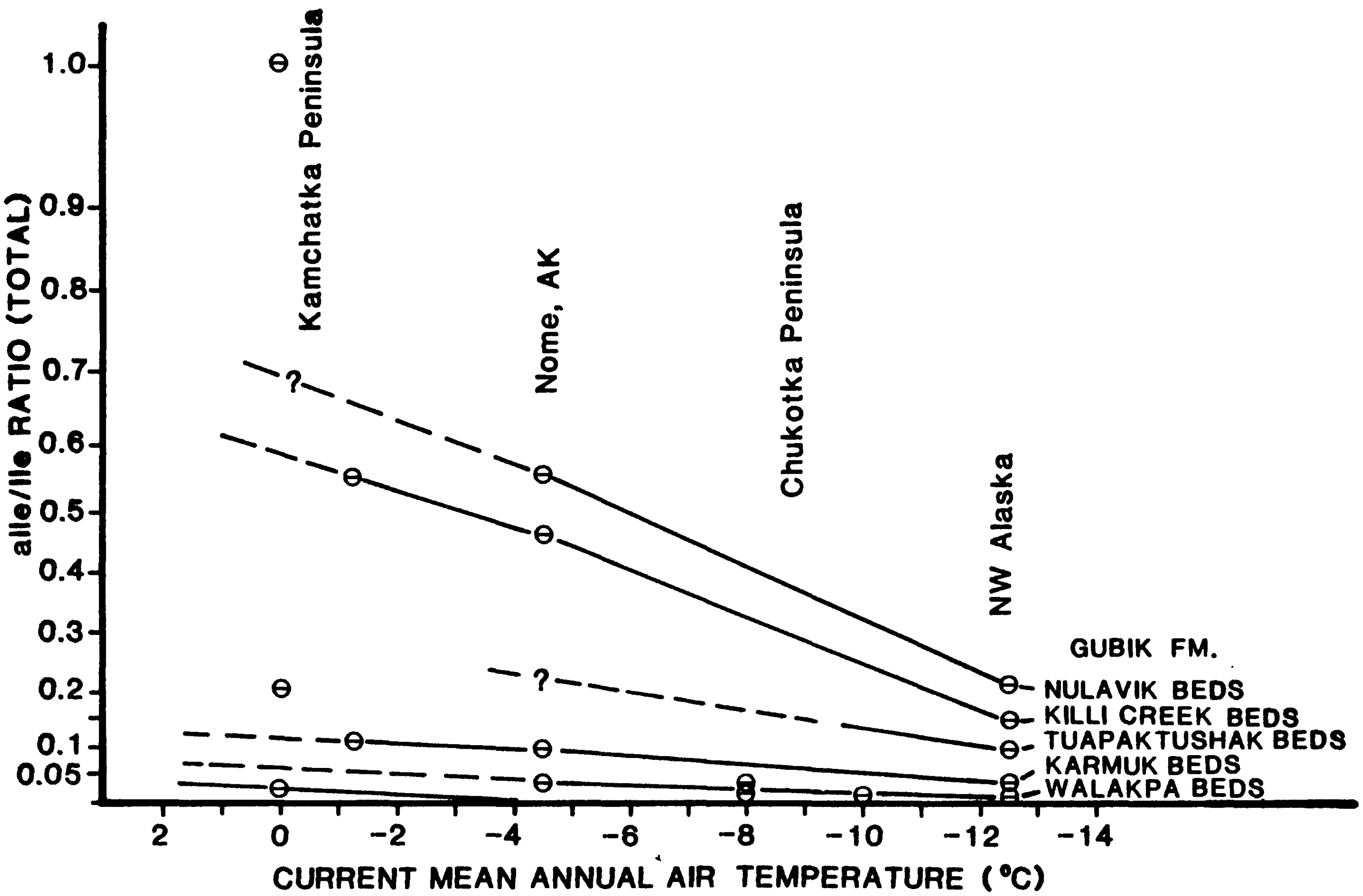

Figure 8.4. Correlation of transgressive, marine sediments between Alaska and northeastern Siberia plotted against current mean annual air temperature.

Attarmen beds, Kamchatka, are anomolously low to be 125 ka old, especially at that latitude and adjacent to the Bering Sea. Assuming an age of 125 ka B.P., the ratio of 0.027 determined for Mya truncata from these sediments requires an effective diagenetic temperature (EDT, Chapter VI) of about -10° C, or roughly 9° C lower than the mean annual air temperature in the area today. This temperature calculation indirectly requires that coastal temperatures along eastern Kamchatka were drastically lower (<< -10° C) during the full glacial periods. Such a conclusion conflicts with CLIMAP reconstructions for the north Pacific that suggest that sea surface temperatues off Kamchatka at 18 ka B.P. were only ca. 2° C lower than present during both the summer and winter months (Moore and others, 1980). On the other hand, this calculation may provide key land-based evidence corrabative with the work of Sancetta and Robinson (1983) who conclude that during the early and late Wisconsin glacial maximum sea ice persisted up to 6 and 9 months of the year, respectively, across the northern Aleutian Basin with only a short ice-free season and no summer warming. If correct, my data would suggest further that sea ice was probably present across much of the western Aleutian basin and all of the neighboring Komandorsky Basin as well, during glacial episodes, and that sea surface temperatures during open water periods of the Wisconsin were never nearly as warm as those of the Holocene.

A major middle Pleistocene interglacial transgression has been documented along numerous reaches of the Bering and Chukchi Sea coast. Based upon stratigraphic position, faunal characteristics, and an evaluation of the thermal history across the region, the Karaginsk beds on Karaginsk Island, Kamchatka, the Kotzebuan sediments at Nome, Alaska, the Pinakul' beds on Chukotka Peninsula and the Karmuk member in northwestern Alaska, might be considered penecontemporaneous representing one or several eustatic transgressions between 200,000 and 500,000 yr. B.P. The best dated deposits within this group are those of the Kotzebuan transgression (Hopkins, 1967a, 1973) which suggest that at least one event may be correlative with either $\delta^{18}O$ stage 7 or 9 of the open-ocean deep sea core records (Shackleton and Opdyke, 1973).

This correlation scheme of mid-Plesitocene marine units proposed here is in contrast to the scheme of Khoreva (1974), who based her correlations upon similarities in the foraminiferal assemblages of these units. She suggested that the Pinakul' suite was correlative with the Anvilian transgression in Alaska (Hopkins, 1967a) and that the Kresta suite, a glaciomarine unit, was correlative with the Kotzebuan transgression. The very low alle/Ile ratios in shells from the Pinakul' suite furnished by Petrov (0.035, Total) precludes the correlation of that particular collection locality with mollusks from the Anvilian deposits that are highly empimerized (0.47, Total), despite any assumed differences in their thermal history. It is important to emphasize that the Pinakul' beds at some localities includes reversely-magnetized sediments suggesting that parts of the Pinakul' complex may, in fact, be considerably older than indicated by the results from Petrov's locality (Hopkins, pers. comm., Nov. 1983). Moreover, aIle/Ile results on shells from the Val'katlen, Kresta and Petrov's Pinakul' locality are quite similar, suggesting that any age differences between these beds are small. The fact that ale/Ile ratios in the Kresta suite are nearly identical to ratios from the Val'katlen supports the conclusion of Hopkins (1973) and Petrov (1966) that at least a portion of the Kresta suite represents a glaciomarine facies associated with glaciation on Chukotka peninsula ca. 150,000 yr. B.P. At Arctic temperatures (CMAT <-7°C), the epimerization reaction proceeds too slowly to distinguish depositional units which differ in age less than ca. 50,000 yr.

Petrov (1982) interpreted the Karaginsk beds of Kamchatka and the Pinakul' beds on Chukotka peninsula as the same transgressive event along the Siberian coast; however, he correlated both of these units with the Einahnutan transgression (Hopkins, 1967a, 1973) that Hopkins now considers to be of Kotzebuan age at the Einahnutan type locality (see Chapter I).

The correlation of transgressive deposits older than middle Pleistocene is less certain, due to (1) our limited knowledge of the complexities of the stratigraphy in various regions (2) the nonlinear rates of the epimerization reaction beyond a Total of .3, and (3) our lack of knowledge concerning climate change (i.e., thermal history) during the late Pliocene and early Pleistocene. As noted earlier, a considerable amount of data has emerged to indicate that the regional climate in northern Beringia during interglacials and perhaps at all times was considerably warmer during Pliocene and early Pleistocene time than during the middle and late Pleistocene (Hopkins, 1972; Nelson, 1979; Brigham and Miller, 1982; Vincent, 1982). In addition, as we have thus far not been able to acquire a complete suite of samples from all of the recognized Pliocene and Quaternary stratigraphic units on Kamchatka and Chukotka penisulas, we are left with an incomplete aminostratigraphy.

AIle/Ile ratios from the upper (0.22, Total) and lower (1.04, Total) portions of the Olkhovaya suite on Kamchatka peninsula suggest that this unit spans a great deal of Pliocene and Pleistocene time. Results from the lower Olkhovaya suite, in particular, suggest that these sediments are much older than the Beringian transgression in western Alaska, especially when compared with data from other Pliocene stratigraphic units. Specimens of Mya sp. from the Coralline Crag of Pliocene age in the British Isles , an area where the CMAT is ca. +10°C, gave aIle/Ile ratios of 1.20 ± .002 in the Total hydrolysate (Miller and others, 1979); the lower Olkhovaya should be considerably older. Paleontologically, however, the lower Olkhovaya suite does not appear to be that old. My interpretation contradicts the correlations proposed by Gladenkov (1981, pg. 20) who equated the entire Olkhovaya suite with marine deposits younger than the Anvilian transgression. Further amino acid work through the Olkhovaya suite and discussions with Gladenkov will be needed to resolve this discrepancy.

The Tusatuvayam beds on Karagin Island, Kamchatka, with a Total aIle/Ile of 0.58 may be correlative with the Anvilian transgression along the Alaskan coast, which at a cooler site yields a lower Total ratio of 0.47. This correlation concurs with Gladenkov (1981). Similarly, the upper Olkhovaya beds may be correlative with the early Pleistocene Tuapaktushak member in northwestern Alaska; however, such a correlation is spectulative based on the available data.

Northern Greenland

Svend Funder, and colleagues at the University of Copenhagen, have recently begun investigations of a Pliocene or early Pleistocene marine sequence known as the Kap Kobenhavn Formation in Independence Fjord along northeast Greenland (Funder and Hjort, 1980; Funder and others, in press). This sequence consists of two marine units including (1) a lower deeper water facies of laminated silt and clay with isolated pockets of ice-rafted detritus, and (2) an upper shallow water facies composed mainly of sand and overlain in some sections by very fine grain sediments suggesting a return of deeper water facies. In contrast to the permanently ice bound coast that exists today in the area, sedimentary structures and in situ mollusks in the Kap Kobenhavn Formation indicate that these sediments were deposited when the coastal waters were at least seasonally ice free and offshore bars were allowed to form. On biostratigraphic criteria, this is thought to have occurred at sometime between 0.7 and 3.0 m.y. B.P. (Funder, 1984). (Unpublished amino acid ratios on shells from the Kap Kobenhavn Formation yield total values similar to those from the Killi Creek Member of the Gubik Formation (S. Funder, pers. comm., January, 1984).

Rich collections of in situ marine mollusks and allochtonous plant and insect macrofossils from nearby limnic and terrestrial environments have been studied for paleoenvironmental considerations (Funder and Hjort, 1980; Fredskild and Roen, 1982; Mogensen, in press). The marine mollusk fauna consists of arctic and high arctic species characteristic of cold polar waters. In contrast, fossil flora and fauna remains from land are characteristic of open boreal forest and low arctic tundra with summer temperatures of at least 10° C and mild winters not colder than -5° to -10° C (Funder, 1984; Funder and others, in press). Although the boreal forst elements are dominated by *Picea* and *Larix*, the vegetation was dominated by a variety of arctic dwarf shrubs. Funder (1984) believes that the environment along the northern cost of Labrador, Canada, may serve as a modern analog to conditions formerly present along northeast Greenland.

Based upon floral and faunal similarities, Funder and others (1984) suggest that the Kap Kobenhavn Formation may be correlative with the Worth Point Formation on Banks Island (Vincent, 1982; Vincent and others, 1983) and the Tjørnes beds of northern Iceland (Einarsson and others, 1967; Albertsson, 1981). Based upon the warm terrestrial environment suggested by elements within the Kap Kobenhavn Formation coincident with coastal conditions that were at least seasonally ice-free, I suggest that the Kap Kobenhavn Formation could be correlative with the Tuapaktushak Member, characterized by warm marine conditions at sometime between 0.73 and 1.7 m.y., but is more likely correlative with the Killi Creek Member of the Gubik Formation that is correlative elsewhere on the Arctic Coastal Plain with a time 1.7 to 2.2 m.y. B.P. when a coniferous forest existed next to a relatively warm Arctic Ocean. Further paleobotanical and amino acid data may clarify this suggestion.

World Ocean Record

The marine lithofacies of the Gubik Formation are thought to represent the infrequent high sea stands that reached above present sea level during interglacial intervals of the late Cenozoic. What is known from the Gubik Formation is that deposits of two transgressions of unknown extent occurred during the Pliocene (Nulavik and Killi Creek members), that deposits of an early Pleistocene transgression (Tuapaktushak Member) are present above 33 - 36 m a.s.l., that a middle Pleistocene transgression (the Karmuk Member) left deposits now up to +23 m a.s.l., and that the Pelukian transgression, or last interglacial high sea stand (Walakpa Member), reached +8 to +10 m a.s.l. The youngest transgression deposited the Flaxman Member up to 7 m a.s.l. from Barrow eastward along the Beaufort Sea.

The idea that sea levels were higher during Pliocene and early Pleistocene time was hypothesized by Zeuner (1959) and later reiterated by Fairbridge (1961, 1971) who postulated that sea level has dropped from early Pleistocene elevations in a series of oscillations superimposed on a long-term trend. This work, however, was primarily based on marine deposits in the Mediterranean where local tectonics have deformed paleoshorelines significantly (Hey, 1978). Vigdorchik (1980) has recently revived the ideas of Fairbridge and proposed further that cold transgressions along the Siberian coast were caused by the physical isolation of the Arctic Ocean from the world ocean during glacial episodes of the middle Pleistocene.

How much tectonics have contributed to the elevation of the older units on the Northern Alaska coastal plain is not known. The Point Lay embayment of marine deposits that straddle the Hanna trough might suggest that this area is, in fact, subsiding. Certainly the consistent height of the Pelukian shoreline, represented by the Walakpa Member at about +10 m a.s.l. across the field area, is close to the mean world sea level height at about 7 m a.s.l., suggesting that little uplift has occurred in the last 125,000 years.

Marine deposits from many different parts of the world suggest that sea level was higher than present during the early Pleistocene. For example, Belknap (1978) describes marine deposits between 15 and 30 m a.s.l. along the east cost of the United States that are believed to be 1 to 2 m.y. old, and Hoxnian marine deposits of middle Pleistocene age are found up to +30 m a.s.l. on the southern coast of England (Bowen, 1979). Stearns (1978) has a K/Ar date that documents sea level at 30 m a.s.l. about 500,000 years B.P. on Oahu, Hawaii. Nunn (1984) provides an extensive review of high-level late Tertiary and early Pleistocene shorelines, some 40 to 300 m a.s.l., in coastal areas of the South Atlantic Ocean. Even in these areas, it is not known to what extent tectonic movements have influenced the present height of the deposits.

Prior to 125,000 years B.P., the absolute height of sea level is unknown. Uncertainties in the oxygen isotope record (Shackleton and Opdyke, 1973, 1977) make it inappropriate to interpret sea level from the deep sea record (Aharon, 1983). Further evidence from coastal areas is needed to evaluate the older sea level record.

Summary

Late Cenozoic high interglacial sea level events are documented in many areas of Beringia and elsewhere on the arctic coasts of North America and Greenland. Based upon biostratigraphic criteria and a variety of relative and absolute dating methods, these events are thought to be penecontemporaneous. Table 8.6 represents a summary of the correlations discussed in this study.

One of the most important conclusions that can be drawn from this correlation scheme is that late Pliocene and early Pleistocene interglacial periods were dramatically warmer than those that characterized the middle and late Pleistocene. Moreover, many of these interglacial periods were accompanied by sea ice conditions much different than that of the present interglacial; perennial sea ice did not exist during several of these periods and some lack evidence that sea ice existed at all. No doubt differences that characterize late Pliocene and early Pleistocene deep-sea ^{18}O record from the middle and late Pleistocene record reflect the influence of an atleast partially ice-free Arctic Ocean during the late Pliocene and early Pleistocene.

Further work along western Alaska will be necessary to extablish whether or not more marine events are recorded there. Additional amino acid analyses in conjunction with some stronger absolute age control will be necessary to determine how the Banks Island history relates to the sea level history of the North Slope. Finally, further work across the coastal plain will help to determine to what extent the shorelines there have been deformed.

Suggestions for Future Work

This dissertation has only scratched the surface of the Gubik Formation, but it does establish a provincial stratigraphical framework for future studies. Further work on the micro- and

Table 8.6. Suggested correlation of stratigraphic units from Kamchatka Peninsula to northern Greenland.

Age m.y.	Kamchatka	Chukotka	W. Alaska	W. Arctic Coastal Plain		Colville R.	Banks Is.	N. Greenland
late Pleistocene					GUBIK FORMATION		Prince of Wales Fm.	
				Flaxman Member		Flaxman F.		
— 0.125 — middle	Attamen Beds	Val'katlen Kresta	Pelukian T.	Walakpa Member		deposits of Pelukian T.	Cape Collinson Fm	
	Karagin Beds	Pinakul	Kotzebuan T.	Karmuk Member			Nelson River Fm.	
Pleistocene								
— 0.73 — early	Upper Olkhovaya Suite	Older Pinakul localities ?		Tuapaktushak M.		Beds at Fish Creek	Morgan Bluffs Fm.	
Pleistocene							Duck Hawk Bluffs Fm.	
— 1.8 —	Tusatuviam Beds		Anvilian T.	Killi Creek M.		Deposits of Colvillian T.		Kap København
							Point Worth Formation	
2.2 Pliocene								
			Beringian T.	Nulavik M.		Beds on Milliveach Ck		
3.5	Lower Olkhovaya Suite							

macrofaunal assemblages in each Member will improve upon the ideas suggested here and may provide additional evidence concerning the absolute age of these deposits.

It goes without saying that many key areas require detailed study. Of critical importance are the sediments exposed between Walakpa and Nunavak bays, near Barrow, where I believe the offlap of the middle Pleistocene beds (Karmuk Member) might be traced. Priority should also be given to tracing the patchy distribution of older deposits above 35 m a.s.l. and linking the deposits of the Gubik Formation across the Western Coastal Plain with the Gubik stratigraphy emerging from the Colville River region (Carter's work). More sections need to be studied to determine how and where regressive marine sequences are preserved.

The isoleucine epimerization reaction is so slow in high Arctic areas that only those deposits more than 100,000 years old can be distinguished. The racemization of L-aspartic acid to its enantiomer, D-aspartic acid, is a faster reaction that may better serve the stratigraphic needs of Quaternary researchers in some areas, particularly those interested in late Pleistocene sea level changes.

The following quote by C.F. Hall (1865) seems as appropriate today as it was over 100 years ago:

> That argument and discussion may arise from portions of what I advance is very probable, but, if so it will be better to enter upon such in another form than this. Readers very naturally expect to be entertained, as well as, perchance, instructed in what a voyager or traveler puts before them. Long, prosy dissertations are seldom wanted. All that most people require is a truthful report of personal doings in strange lands and a faithful record of incidents, discoveries, and interesting events connected with them. Such, then, is the task I have taken in hand, with hope that a ready excuse will be granted for all of those imperfections necessarily consequent upon the mode and manner of my carrying on the work in which I was engaged.

BIBLIOGRAPHY

Aharon, P., 1983, 140,000-yr. isotope climatic record from raised coral reefs in New Guinea, Nature, v. 304, p. 720-723.

Albertson, K.J., 1981, On Tertiary tillites in Iceland, in Hambrey, M.J., and Harland W.B., eds., Earth's pre-Pleistocene glacial record: Cambridge University Press, Cambridge, p. 562-565.

American Commission on Stratigraphic Nomenclature (ACSN), 1970, Code of Stratigraphic nomenclature, (2nd ed.): American Association of Petroleum Geologists, Tulsa, Oklahoma, 45 p.

_______________, 1983, North American Stratigraphic Code: American Association of Petroleum Geologists Bulletin, v. 67, p. 841-875.

Anderson, P.A., 1981, Reconstructing the past: The synthesis of archaeological and palynological data, northwestern Alaska and northwestern Canada: Ph.D. dissertation, Brown University, 600 p.

Andrews, J.T., Miller, G.H. and Clark, P.C., 1982, Instrumental Calibration and Standards in Amino Acid Geochronology Laboratory Report of Current Activities, January 1981 - May 1982, Miller, G.H., ed.,: Institute of Arctic and Alpine Research, p. 45-47.

Andrews, J.T., Miller, G.H., Davies, D.C., and Davies, K.H., submitted, Identification of the Genus of Fragments of Quaternary Fossil Molluscan Genera from amino acid chromatography: Geological Society of America Bulletin.

Bada, J.L., Steinberg, S., Ho, M.-S., and Ruttenburg, K., 1982, Amino Acid diagenetic reactions in calcareous fossils and their geochronological implications: Geological Society of America, Abstracts with Programs, v. 14, n. 7, p. 437.

Balobeav, V.T., 1978, Reconstruction of paleoclimate from present day geotherma data, in Proceedings of the Third International Conference on Permafrost, v. 1: Ottawa, National Research Council of Canada, p. 10-14.

Barnes, P.W., and Reimnitz, E., 1974, Sedimentary processes on Arctic shelves of the northern coast of Alaska, in The Coast and Shelf of the Beaufort Sea, J.C. Reed and J.E. Sater (eds.) Arlington: Arctic Institute of North America, p. 439-476.

Barnes, P.W., and Hopkins, D.M., 1978, Geological Sciences Summary, in Environmental Assessment of the Alaskan Continental Shelf, Interim Synthesis: Beaufort and Chukchi, August: U.S. Dept. of Commerce, National Oceanic and Atmospheric Administration and Bureau of Land Management, p. 101-133.

Barry, R.G., Approaches to reconstructing the climate of the steppe-tundra biome, in Hopkins, D.M., Matthews, Jr., J.V., Schweger, C.E., and Young, S.B., eds., Paleoecology of Beringia: New York, Academic Press, p. 195-204.

Belknap, D.F., 1979, Application of Amino acid Geochronology to Stratigraphy of Late Cenozoic Marine units of the Atlantic Coastal Plain: Ph.D. thesis University of Delaware, 550 p.

Benson, J.R., and Hare, P.E., 1975, O-Phthalaldehyde: fluorogenic detection of primary amines in the picomole range: comparison with fluorescamine and ninhydrin: Proceedings of the National Academy of Sciences, U.S.A., v. 72, p. 612-622.

Benard, F.R., 1979, Bivalve mollusks of the western Beaufort Sea: Contributions in Science, No. 313, Natural History Museum of Los Angeles County, 80 p.

Bernard, H.A., LeBlanc, R.J., and Major, C.F., Jr., 1962, Recent and Pl s e geology of southeast Texas: Geology Gulf Coast and Central Texas and guidebook of excursion, Houston Ge p. 175-225.

Black, R.F., 1951, Eolian deposits of Alaska: Arctic, v. 4, n. 2, p. 89-111.

______________, 1964, Gubik Formation of Quaternary age in northern Alaska: U.S. Geological Survey Professional Paper 302-C, p. 59-91.

______________, 1974, Ice-wedge polygons of northern Alaska, in Glacial Geomorphology, ed. D.R. Coates, Chapter 9: Binghamton, N.Y., Publications in Geomorphology, SUNY, p. 247-275.

Blanchard, R.L., 1963, Uranium decay series disequilibrium in age determination of marine calcium carbonates: Ph.D. thesis, Washington University, St. Loius, 164 p.

Bowen, D.Q., 1979, Quaternary Geology: Pergamon Press, New York, 221 p.

Briggs, W.M., 1980, The modern ostracode fauna of the continental shelf, slope, and rise of the southeastern Beaufort Sea: American Quaternary Association, 6th Biennial Meetings, Abstracts and Programs, p. 49.

__________, in press, Ostracoda, in Atlas of the Beaufort Sea, ed. B. Pelletier, Geological Survey of Canada.

Brigham, J.K., 1980, Stratigraphy, amino acid geochronology and genesis of Quaternary sediments, Broughton Island, E. Baffin Island, Canada: Masters' thesis, University of Colorado, Boulder, 199 p.

__________, 1982, Amino Acid Geochronology of Quaternary transgressions on the Alaskan Arctic Coastal Plain and the potential for correlations with Quaternary Events on Kamchatka and Chuktoka peninsulas, USSR: IX INQUA Congress, Moscow, August, Abstracts, v. 1, p. 50.

__________, 1983a, Stratigraphy, amino acid geochronology, and correlation of Quaternary sea-level and glacial events, Broughton Island, arctic Canada: Canadian Journal of Earth Science, v. 20, n. 4, p. 577-598.

__________, 1983b, Intrashell variations in amino acid concentrations and isoleucine epimerization ratios in fossil Hiatella arctica: Geology, v. 11, p. 509-513.

Brigham, J.K. and Miller, G.H., 1982, Late Cenozoic History of High Sea Level Stands and Pleistocene temperature estimates, northwestern Alaska: Geological Society of America Abstracts wi h ograms, v. 14 (7), p. 451-452.

__________, 1983, Marine stratigraphy of the Gubik Formation and paleotemperature estimates of the Alaskan Arctic Coastal Plain during the last 125,000 years, in Permafrost -- Fourth International Conference, Fairbanks: Washington, D.C., National Academy of Sciences, v. 1, p. 80-85.

Britton, M.E., 1957, Vegetation of the arctic tundra, in Hansen P. ed., Arctic Biology: Corvallis, Oregon State Univ. Press, p. 26-61.

Brouwers, E.M., 1982, Analysis of Inner Neritic Ostracode Assemblages from the Beaufort Sea: Implications for interpreting analogous Quaternary Sediments, Institute of Arctic and Alpine Research: 11th Annual Arctic Workshop, Abstracts, p. 5-7.

Bruum, P., 1962, Sea level rise as a cause of shore erosion: Proceeding ASCE, Journal of Water W. Harbors Division, v. 88, p. 117-130.

Cady, J.W., 1984, Erosion surfaces Offset by thrust faults? Comparison of Side-looking Radar, Landsat, and reflection seismic Interpretations, National Petroleum Reserve in Alaska, Journal of Geophysical Research, submitted.

Cantilon, J.E., 1961, Plant cover in relation to macro-, meso-, and micro-relief: Final Report, Office of Naval Research, Grants No. ONR-208 and 216, 128 p.

Carter, L.D., 1981a, A Pleistocene sand sea on th Alaskan Arctic Coastal Plain: Science, v. 211, p. 381-383.

__________, 1981b, Middle Wisconsin through Holocene climate in the Ikpikpuk River region, Alaska: Institute of Arctic and Alpine Research: Tenth Annual Arctic Workshop, Abstracts, p. 5-9.

__________, 1982, Late Wisconsin desertification in northern A : Geological Society of America Abstracts with Programs, v. 7, p. 461.

Carter, L.D., 1983a, Fossil sand wedges on the Alaskan Arctic Coastal Plain and their paleoenvironmental significance: Proceedings of the Fourth International Conference on Permafrost, July 171-22, 1983, Fairbanks, Alaska, Washington, D.C., National Academy Press, pp. 109-114.

__________, 1983b, A Pleistocene sand desert in Arctic Alaska, Polar Symposium: U.S. Geological Survey Circular 911, p. 36.

Carter, L.D., and Robinson, S.W., 1978, Eolian sand and interbedded organic horizons at Keolok Creek on the Arctic Coastal Plain of Alaska--possible regional implications: U.S. Geological Survey Open-File Report 78-320, 26 p.

__________, 1981, Minimum age of beach deposits north of Teshekpuk Lake, Alaskan Arctic Coastal Plain: U.S. Geological Survey Circular 823-B. p. B8-B9.

Carter, L.D., Marincovich, Jr., L., Brouwers, E.M., and Forester, R.M., 1979, Paleogeography of a Pleistocene coastline, Alaskan Arctic Coastal Plain: U.S. Geological Survey Circular 804B, p. B39-B41.

Carter, L.D. and Galloway, J.P., 1982, Terraces of the Colville River Delta region, Alaska: U.S. Geological Survey Circular 844, p. 49-51.

Chapman, R.M., and Sable, E.G., 1960, Geology of the Utukok-Corwin region, northwestern Alaska: U.S. Geological Survey Professional Paper 303-C, 164 p.

Clark, D.L., 1982, Origin, nature, and world climate effect of Arctic Ocean ice-cover: Nature, v. 300, p. 321-325.

Clark, D.L., Whitman, R.R., Morgan, K.A., Mackey, S.D., 1980, Stratigraphy and Glacial-Marine Sediments of the American Basin, Central Arctic Ocean: Geological Society of America Special Paper 181, 57 p.

Cooper, S.C., 1964, Benthonic foraminifera of the Chukchi Sea: Cushman Contribution, Laboratory Foraminifera Research, v. 15, n. 3, p. 79-104.

Coulter, H.W., Hussey, K.M., and O'Sullivan, J.G., 1960, Radiocarbon dates relating to the Gubik Formation, northern Alaska: U.S. Geological Survey Professional Paper 400-B, p. B350-B351.

Cox, A., Hopkins, D.M., and Dalrymple, G.B., 1966, Geomagnetic polarity epochs, Pribilof Islands: Geological Society of America Bulletin, v. 77, p. 883-909.

Creager, J.S., and McManus, D.A., 1967, Geology of the floor of Bering and Chukchi Seas-American Studies, in D.M. Hopkins, ed., The Bering Land Bridge: Stanford University Press, Stanford, Calif., p. 7-31.

Cronin J.R., Pizzarello, S., and Grandy, W.E., 1979, Amino acid analysis with o-phthailaldehyde detection: effects of reaction temperature and thiol on fluorescence yields: Analytical Biochemistry v. 93, p. 174-179.

Dall, W.H., 1920, Pliocene and Pleistocene fossils from the Arctic coast of Alaska and the auriferous beaches of Nome, Norton Sound, Alaska: U.S. Geological Survey Professional paper 125-C, p. 1-37.

Dall, W.H. and Harris, G.D., 1892, Correlation papers - Neogene: U.S. Geological Survey Bulletin 84, 349 p.

Decker, E.R., and Bucher, G.J., 1982, Geothermal studies in the Ross Island - Dry Valley region, in Craddock, C., ed., Antarctic Geoscience: Madison, Wisconsin, University of Wisconsin Press, p. 887-894.

Detterman, R.L., Reiser, H.N., Brosge, W.P., and Dutro, J.T., Jr., 1975, Post-Carboniferous stratigraphy, northern Alaska: U.S. Geological Survey Professional paper 886, 46 p.

Dillon, W.P., 1970, Submergence effects on a Rhode Island barrier and lagoon and inferences on migration of barriers: Journal of Geology, v. 78, p. 94-106.

Dingman, S.L., Barry, R.G., Willer, G., Benson, C.S., LeDrew, E.F., and Goodwin C.W., 1980, Climate, snow cover, microclimate, and hydrology, in Brown, J., and others eds., An Arctic Ecosystem: the coastal tundra at Barrow: Alaska, Strowdsburg, Pennsylvania, Dowden, Hutchinson and Ross, p. 30-65.

Dinter, D.A., 1983, Quaternary sedimentation on the Alaskan Beaufort Shelf: Sediment Sources, Glacioeustatism, and regional tectonics: Polar Symposium, U.S. Geological Survey Circular 911, p. 37-38.

Dutro, J.T., Jr., 1981, Geology of Alaska bordering the Arctic ocean, in Nairn A.E.M., Churkin, M., Jr., M. Stehli, F.G., eds., The Ocean Basins and Margins, the Arctic Ocean, v. 5: New York. Plenum Press, p. 21-36.

Easterbrook, D.J. and Rutter, N., 1982, Amino Acid analyses of wood and shells in development of chronology and correlation of Pleistocene sediments in the Puget Lowland, Washington: Geological Society of America, Abstracts with Programs, v. 14, n. 7, p. 480.

Echols, R.J., 1974, Benthic foraminifera of the Alaskan shelf and slope of the Beaufort Sea (Abstract), in The Coast and Shelf of the Beaufort Sea, Reed, J.C., and Sater, J.E., eds.: Arctic Institute of North America, 750 p.

Einarsson, T., Hopkins, D.M., and Doell, R.R., 1967, The stratigraphy of Tjörnes, northern Iceland, and the history of the Bering Land Bridge, in The Bering Land Bridge, D.M. Hopkins, ed.: Stanford University Press, Stanford California, p. 313-325.

Elliot, T., 1978, Clastic shorelines, in H.G. Reading, ed., Sedimentary Environments and Facies: Elsevier, New York, p. 143-177.

Ervin, L.D., 1981, The geologic significance of compaction gradient plots of wells in the National Petroleum Reserve - Alaska: Report prepared for Husky Oil NPR Operations, Inc. under contract to U.S. Geological Survey Office of National Petroleum Reserve in Alaska Contract No. GC-80-11, 22p.

Faas R.W., 1962, Foraminiferal paleocology of the Gubik (Pleistocene) Formation of the Barrow Area, northern Alaska: Iowa Academy of Science Proceedings, v. 69, p. 354-361.

__________, 1966, Paleoecology of an arctic estuary: Arctic v. 19, p. 343-348.

Fairbridge, R.W., 1961, Eustatic changes in sea-level: Physics and Chemistry of the Earth, v. 4, p. 99-185.

__________, 1971, Quaternary shoreline problems at INQUA: Quaternaria, v. 15, p. 1-17.

Fischer, A.G., 1961, Stratigraphic record of transgressing seas in the light of sedimentation on the Atlantic coast of New Jersey: American Association of Petroleum Geologists Bulletin, v. 45, p. 1656-1666.

Fredskild, B., and Roen, U., 1982, Macrofossils in an interglacial peat deposit at Kap Kobenhavn, north Greenland: Boreas, v. 11, p. 181-185.

Funder, S., 1984, On an arctic beach one million years ago: YMER-80 Symposium Abstracts, April, Voksenasen, Oslo.

Funder, S., and Hjort, C., 1980, A reconnaissance of the Quaternary geology of eastern north Greenland: Rapp. Gronlands geol. Unders. 99, p. 99-105.

Funder, S., Bennike, O., Morgensen, G.S., Noe-Nygaard, B., Pedersen, S.A.S., Petersen, K.S., in press, The Kap Kobenhavn Formation, a late Cainozoic sedimentary sequence in north Greenland: Rapp. Gronlands geol. Unders. 120, p. 9-18.

Galloway, J.P., 1982, Grain-size analyses of 20 eolian sand samples from northern Alaska: U.S. Geological Survey Circular 844, p. 51-53.

Gladenkov, Y.B., 1981, Marine Plio-Pleistocene of Iceland and problems of its correlation: Quaternary Research, v. 15, p. 18-23.

Gold, L.W., and Lachenbruch, A.H., 1973, Thermal conditions in permafrost--A review of North American literature, in Permafrost--The North American contribution to the Second International Conference, Yakutsk: Washington, D.C., National Academy of Sciences, p. 3-23.

Grantz, A., Eittreim, S., Whitney, O.T., 1981, Geology and Physiography of the Continental Margin north of Alaska and Implication for the origin of the Canada Basin, in Nairn, A.E.M., Churkin, Jr., M., Stehli, F.G., eds., The Ocean Basins and Margins, the Arctic ocean, v. 5: New York Plenum Press, p. 439-492.

Grantz, A., Dinter, D.A., Hill, E.R., Hunter, R.E., May, S.D., McMullin, and Phillpis R.L., 1982, Geologic Framework, Hydrocarbon Potential, and Environmental conditions for Exploration and Development of Proposed Oil and Gas Lease Sale 85 in the central and northern Chukchi Sea: U.S. Geological Survey Open-File Report 82-1053, 84 p.

Gryc, G., Patton, W.W., Jr., and Payne, T.G., 1951, Present Cretaceous stratigraphic nomenclature of northern Alaska: Washington Academy of Science Journal. v. 41, p. 159-167.

Hall, C.F., 1865, Arctic Research and Life among the Esquimaux: Harper and Brothers Publishers, New York, 595 p.

Hamilton, T.D., 1981, Multiple moisture sources and the Brooks Range glacial record: Tenth Annual Institute of Arctic and Alpine Research Workshop, p. 16-18.

Hare, P.E., 1967, Amino acid composition of the extrapallial fluid in mollusks: Carnegie Institution of Washington Yearbook, v. 65, p. 364-365.

__________, 1975, Amino acid composition by column chromatography, in Needleman, S.B., ed., Molecular biology, biochemistry and biophysics, v. 8, p. 205-231.

__________, 1980, Organic geochemistry of bone and its relation to the survival of bone in the natural environment, in Behrensmeyer, A.K. and Hill, A., eds., Taphonomy and vertebrate paleontology: University of Chicago Press.

Hare, P.E. and Mitterer, R.M., 1969, Laboratory simulation of amino acid diagenesis in fossils: Carnegie Institute of Washington Yearbook, v. 67, p. 205-208.

Hare, P.E., Hoering, T.C., and King, K., 1980, Biogeochemistry of amino acids: New York, Wiley Press, 558 p.

Haugen, R.K. and Brown, J., 1980, Coastal-inland distributions of summer air temperature and precipitation in northern Alaska: Arctic and Alpine Research, v. 12, n. 4, p. 403-412.

Hey, R.W., 1978, Horizontal Quaternary shorelines of the Mediterranean: Quaternary REsearch, v. 10, p. 197-203.

Hopkins, D.M., 1959, Cenozoic history of the Bering Land Bridge (Alaska): Science, v. 129, p. 1519-1528.

__________, 1967a, Quaternary marine transgressions in Alaska, in Hopkins, D.M., ed., The Bering Land Bridge, Stanford, California: Stanford Univ. Press, p. 47-90.

__________, 1967b, The Cenozoic History of Beringia - A Synthesis, in Hopkins, D.M., ed., The Bering Land Bridge, Stanford, California: Stanford Univ. Press, p. 451-484.

__________, 1972, The paleogeography and climatic history of Beringia during the late Cenozoic time: Internord, v. 12, p. 121-150.

__________, 1973, Sea level history in Beringia during the past 250,000 years: Quaternary Research, v. 3, p. 520-540.

__________, 1982a, Aspects of the paleogeography of Beringia during the late Pleistocene, in Hopkins D.M., Matthews, Jr., J.V., Schweger, C.E., and Young, S.B., eds., Paleoecology of Beringia: New York, Academic Press, p. 3-28.

__________, 1982b, Abortive Glaciations at High Latitudes indicated by Glaciomarine Deposits, Gubik Formation, northern Alaska: Geological Society of America Abstracts with Programs, v. 14, n. 7, p. 518.

__________, in press, Whale biogeography and the history of the Arctic Basin, Proceedings, Groningen Symposium on Whales and Whale Conservation on the Arctic: Groningen Rijksuniversitet Press.

Hopkins, D.M., MacNeil, F.S., and Leopold, E.B., 1960, The coastal plain at Nome, Alaska - a late Cenozoic type section for the Bering Strait region: Report, 21st International Geological Congress (Copenhagen), part 4, p. 46-57.

Hopkins, D.M., MacNeil, F.S., Merklin, R.L., Petrov, O.M., 1965, Quaternary correlations across the Bering Strait: Science, v. 147, p. 1107-114.

Hopkins, D.M., Matthews, J.V., Wolfe, J.A., and Silberman, M.L., 1971, A Pliocene flora and insect fauna from the Bering Strait region: Paleogeography, Paleoclimatology, and Paleoecology, v. 9, p. 211-231.

Hopkins, D.M., Rowland, R.W., Echols, R.E., and Valentine, P.C., 1974, An Anvilian (Early Pleistocene) Marine Fauna from Western Seward Peninsula, Alaska: Quaternary Research, v. 4, p. 441-470.

Hopkins, D.M., Smith, P.A., and Matthews, Jr., J.V., 1981, Dated wood from Alaska and the Yukon: Implications for forest refugia in Beringia: Quaternary Research, v. 15, p. 217-249.

Hopkins, D.M., Matthews, Jr., J.V., Schweger, C.E., and Young, S.B., 1982, Paleoecology of Beringia: New York, Academic Press, 489 p.

Hopkins, D.M., Carter, L.D., Brigham, J.K., Brouwers, E., McDougall, K.A., Marincoirch, Jr., L.N., Nelson, R.E., Repenning, C.A., and Williams, J.R., in prep., The Flaxman member of the Gubik Formation of northern Alaska: in Changes in stratigraphic nomenclature by U.S. Geological Survey, U.S. Geological Survey Bulletin.

Hughes, O.L., Harrington, C.R., Janssen, J.A., Matthews, Jr., J.V. Morlan, R.E., Rutter, N.W., Schweger, C.E., 1981, Upper Pleistocene Stratigraphy, Paleoecology, and Archaeology of the Northern Yukon Interior, Eastern Beringia 1. Bonnet Plume Basin: Arctic, v. 34, n. 4, p. 329-365.

Hughes, T., Denton, G.H., and Groswald, M.G., 1977, Was there a late-Würm Arctic Ice Sheet?: Nature, v. 266, p. 596-602.

Hume, J.D., Schalk, M., 1967, Shoreline processes near Barrow, Alaska: a comparison of normal and the catastrophic: Arctic, v. 20, no. 2, p. 86-103.

International Subcommission on Stratigraphic Classification (ISSC), 1976, International Stratigraphic Guide, H.D. Hedberg, ed.: John Wiley and Sons, New York, 200 p.

Jensen, H., Graham, L.C., Porcello, L.J., Leith, E.N., 1977, Side-looking Airborne Radar: Scientific American v. 237, n. 4, p. 84-95.

Kauffman, E.G., 1969, Cretaceous marine cycles of the western Interior: Mountain Geologist, v. 6, p. 227-245.

Kaufman, A., Broecker, W.S., Ku, T.-L., and Thurber, D.L., 1971, The status of U-series methods of mollusk dating: Geochimica Cosmochimica Acta, v. 35, p. 1155-1181.

Kennedy, W.J., Taylor, J.D., and Hall, A., 1969, Environmental and Biological controls on bivalve shell mineralogy: Biological Reviews, v. 44, p. 499-534.

Khoreva, I.M., 1974, Stratigraphy and Foraminifera of the Marine Quaternary Deposits of the west shore of the Bering Sea: Akademy Nauk USSR. Geologic Institute, v. 225, 152 p.

Kraft, J.C., 1971, Sedimentary environment, facies pattern, and geologic history of a Holocene marine transgression: Geological Society of America Bulletin, v. 82, p. 2131-2158.

Kriaulsakul, N., and Mitterer, R.M., 1978, Isoleucine epimerization in peptides and proteins: kinetic factors and applications to fossil proteins: Science, v. 201, p. 1011-1014.

Kriaulsakul, N., and Mitterer, R.M., 1983, Epimerization of COOH-terminal isoleucine in fossil dipeptides: Geochimica et Cosmochimica Acta, v. 47, p. 963-966.

Kvenvolden, K.A., 1975, Advances in the geochemistry of amino acids, in Donald, F.A., ed.: International Annual Reviews of Earth and Planetary Sciences, v. 3, p. 183-212.

Lachenbruch, A.H., Brewer, M.C., Greene, G.W., and Marshall, B.V., 1962, Temperature in permafrost: Temperature--Its measurement and control in science and industry, v. 3, pt. 1, p. 791-803.

Lagoe, M.B., 1979a, Modern benthic foraminifera from Prudhoe Bay, Alaska: Journal of Paleontology, v. 53, p. 258-262.

__________, 1979b, Recent benthonic foraminiferal biofacies in the Arctic Ocean: Micropalentology, v. 25, p. 214-224.

Leffingwell, E. de K., 1919, The Canning River region, northern Alaska: U.S. Geological Survey Professional Paper 109, 251 p.

Loeblich, A.R., Jr. and Tappen, H., 1953, Studies of Arctic Foraminifera: Smithsonian Miscellaneous Collections, v. 121, n. 7, 150 p.

MacCarthy, G.R., 1958, Glacial boulders on the Arctic coast of Alaska: Arctic, v. 11, p. 71-85.

MacGinitie, G.E., 1955, Distribution and ecology of the marine invertebrates of Point Barrow, Alaska: Smithsonian Miscellaneous Collections, v. 128, no. 9, 201 p.

MacGinitie, N., 1959, Marine Mollusca of Point Barrow, Alaska: Proceedings of the U.S. National Museum, v. 109, no. 3412., Smithsonian Institute, p. 59-209.

MacPherson, E., 1971, The Marine Molluscs of Arctic Canada: Publications in Biological Oceanography, No. 3, National Museum of Canada, Ottawa, 149 p.

Major, R.P., and Matthews, R.K., 1983, Isotopic composition of bank margin carbonates on Mid-way Stoll: Amplitude constraint on post early Miocene eustasy: Geology, v. 11, p. 335-338.

Mann, D.M., 1977, Shelled Benthic Fauna of the eastern Chukchi Sea: U.S. Geological Survey Open-File Report M-77-672, 112 p.

Matthews, J.V. Jr., 1974, Quaternary environments at Cape Deceit, Seward Peninsula, Alaska: evolution of a tundra environment: Geological Society of America Bulletin, v. 85, p. 1353-1384.

Matthews, J.V., Jr., 1976, Insect fossils from the Beaufort Formation geological and biological significance: Geological Survey of Canada Paper 76-1B, p. 217-227.

McCulloch, D.S., 1967, Quaternary geology of the Alaskan shore of Chukchi Sea, in Hopkins, D.M., ed., The Bering land bridge: Stanford California, Stanford University Press, p. 91-120.

McCulloch, D.S., Taylor, D.W., and Rubin, M., 1965, Stratigraphy, nonmarine mollusks, and radiometric dates from Quaternary deposits in the Kotzebue Sound area, western Alaska: Journal of Geology, v. 73, p. 442-453.

McDougall, K.A., 1983, Quaternary microfossils on the Arctic Shelf: U.S. Geological Survey Circular 911, p. 39-40.

McManus, D.A., Kelley, J.C., and Creager, J.S., 1969, Continental Shelf Sedimentation in an Arctic Environment: Geological Society of America, v. 80, p. 1961-1984.

McManus, D.A., Creager, J.S., Echols, R.J., and Holmes, M.L., 1983, The Holocene transgression on the Arctic flank of Beringia: Chukchi Valley to Chukchi Estuary to Chukchi Sea, in P.M. Masters, and N.C. Flemming, (eds.), Quaternary Coastlines and Marine Archaelogy: New York, Academic Press, p. 365-388.

McNeil, F.S., 1957, Cenozoic megafossils of northern Alaska: U.S. Geological Survey Professional Paper 294-C, p. 99-125, pls. 11-17.

Meek, C.E., 1923, Notes on stratigraphy and Pleistocene fauna from Peard Bay, Arctic Alaska: California University Department of Geological Sciences Bulletin, v. 14, n. 13, p. 409-422.

Miller, G.H., 1982, Quaternary Depositional Episodes, western Spitsbergen, Norway: Aminostratigraphy and Glacial History: Arctic and Alpine Research, v. 14, n. 4, p. 321-340.

__________, 1983, Unpublished results of pyrolysis experts on Mya truncata.

__________, in press, Aminostratigraphy of Baffin Island shell-bearing Deposits, in Quaternary of Baffin Island, Baffin Bay, and west Greenland, J.T. Andrews, ed.: Allen and Unwin, England.

Miller, G.H., Andrews, J.T., and Short, S.K., 1977, The last interglacial/glacial cycle, Clyde Foreland, Baffin Island, N.W.T.,: Stratigraphy, biostratigraphy, and chronology: Canadian Journal of Earth Science, v. 14, no. 12, p. 2824-2857.

Miller, G.H., Hollin J.T., and Andrews, J.T., 1979, Aminostratigraphy of UK Pleistocene deposits: Nature, v. 281, p. 539-543.

Miller, G.H., and Hare, P.E., 1980, Amino acid geochronology: Integrity of the carbonate matrix and potential of molluscan fossils: in Hare, P.E., Hoering, T.C., and King, K., eds., Biogeochemistry of Amino Acids: New York, Wiley Press, p. 415-444.

Miller, G.H., and Hopkins, D.M., 1980, Degradation of Molluscan shell protein by lava induced transient heat flow, Pribilof Islands, Alaska: Implications for Amino Acid Geochronology and Radiocarbon Dating, in Biogeochemistry of Amino Acids, Hare, P.E., Hoering, T.C. and King, K., Jr. eds.: John Wiley and Sons, New York, p. 445-452.

Miller, G.H., Hare, P.E., Haynes, C.V. and McCoy, W.D., 1982, The influence of environmental parameters on amino acid racemization: Unpublished National Science Foundation Progress Report, Grant No. EAR-78-08222.

Miller, G.H., Brigham, J.K. and Clark, P.C., 1982, Alteration of the Total aIle/Ile ratio by different methods of preparation, in Amino Acid Geochronology Laboratory Report of Current Activities, January 1981 - May 1982, Miller, G.H., ed.: Institute of Arctic and Alpine Research, p. 9-20.

Miller, G.H., Sejrup, H.-P., Mangerud, J., Anderson, B., 1983, Amino acid ratios in Quaternary mollusks and foraminifera from western Norway: Correlation, geochronology and paleotemperature estimates: Boreas, v. 12, p. 107-124.

Miller, R.D., and Dobrovolny, E., 1959, Surficial geology of Anchroage and vicinity, Alaska: U.S. Geological Survey Bulletin 1093, 128 p.

Mitterer, R.M., 1982, Comparison of kinetics and degree of isoleucine epimerization in pure and fossil dipeptides: Geological Society of America, Abstracts with Programs, v. 14, n. 7, p. 567.

Moore, D.G., 1964, Acoustic-reflection reconnaissance of continental shelves: eastern Bering and Chukchi Seas, in R.L. Miller, ed., Papers in marine geology, Shepard commemorative volume: Macmillan, New York, p. 319-362.

Moore, T.C., Jr., Burckle, L.H., Geitzenauer, K., Luz, B., Molina-Cruz, A., Robertson, J.H., Sachs, H., Sancetta, C., Thiede, J., Thompson, P., Wenkam, C., 1980, The reconstruction of sea surface temperatures in the Pacific Ocean of 18,000 B.P.: Marine Micropaleontology, v. 5, p. 215-247.

Morgensen, G.S., in press, Subfossil bryophytes from Plio-Pleistocene Kap Kobenhavn, northeast Greenland: Lindbergia, v. 10.

Moritz, R.E., 1977, On a possible sea-breeze circulation near Barrow, Alaska: Arctic and Alpine Research, v. 9, p. 427-431.

Naidu, A.S., Burrell, D.C. and Hood, D.W., 1971, Clay mineral composition and geologic significance of some Beaufort Sea sediments: Journal of Sedimentary Petrology, v. 41, p. 691-694.

Naidu, A.S., and Mowatt, T.C., 1983, Sources and dispersal patterns of clay minerals in surface sediments from the continental shelf areas off Alaska: Geological Society of America Bulletin, v. 94, p. 841-854.

Nelson, A.R., 1978, Quaternary glacial and marine stratigraphy of the Qivitu Peninsula, northern Cumberland Peninsula, Baffin Island, Canada: unpublished Ph.D. thesis, University of Colorado, Boulder, 299 p.

__________, 1982, Aminostratigraphy of Quaternary marine and glaciomarine sediments, Qivitu Peninsula Baffin Island: Canadian Journal of Earth Science, v. 19, p. 945-961.

Nelson, R.E., 1979, Quaternary environments of the arctic slope of Alaska: MSc. thesis, University of Washington, 141 p.

_________, 1981, Paleoenvironments during deposition of a section of the Gubik Formation exposed along the lower Colville River, North Slope, in The United States Geological Survey in Alaska: Accomplishments during 1979, N.R.D. Albert and T. Hudson, eds., U.S. Geological Survey Circular 823-B, p. B9-B11.

_________, 1982, Late Quaternary environments of the western slope, Alaska: Ph.D. dissertation, University of Washington, 90 p.

Nunn, P.D., 1984, Review of evidence for Late Tertiary shorelines occurring on south Atlantic coasts: Earth-Science Reviews, v. 20, p. 185-210.

Osterkamp, T.E., 1982, Potential impact of a warmer climate on permafrost in Alaska: Proceedings of Conference on Potential effects of CO_2 Induced Climatic Change in Alaska, University of Alaska, April 7-8, University of Alaska, Fairbanks.

Osterkamp, T.E., and Payne, M.W., 1981, Estimates of Permafrost thickness from well logs in northern Alaska: Cold Regions Science and Technology, v. 5, p. 13-27.

O'Sullivan, J.B., 1961, Quaternary Geology of the Arctic Coastal Plain, northern Alaska: Doctoral dissertation, Iowa State University of Science and Technology, Ames, 191 p.

Payne, T.G., and others, 1951, Geology of the Arctic Slope of Alaska: U.S. Geological Survey Oil and Gas Map OM-126, 3 sheets.

Petrov, O.M., 1966, Stratigraphy and Mollusk Faunas of Chukotka Peninsula: Akademy Nauk SSSR, Geological Institute Trudy 155, Izd-vo Nauka, Moscow, 288 p.

_________, 1967, Paleogeography of Chukotka During Late Neogene and Quaternary time, in The Bering Land Bridge, D.M. Hopkins, ed.: Stanford University Press, p. 144-171.

_________, 1982, Marine mollusks of the Antrohopogene from the northern region of the Pacific: Nauka, v. 357, 144 p. [In Russian]

Pewe, T.L., 1975, Quaternary Geology of Alaska: U.S. Geological Survey Professional Paper 835, 145 p.

Phillips, R.L., Reiss, T.E., Kempema, E., Reimnitz, E., 1984, Nearshore marine goelogic investigations, northeast Chukchi Sea, Wainwright to Skull Cliff, U.S. Geological Survey Open File Report 84-108, 33 p.

Repenning, C.A., 1983, New Evidence for the Age of the Gubik Formation, Alaskan North Slope: Quaternary Research, v. 19, p. 356-372.

Roderick, C.A., 1975, The origin, distribution, and depositional history of gravel deposits on the Beaufort Sea continental shelf, Alaska: San Jose State University, M.S. thesis, 76 p.

Rosholt, J.N., 1980, Uranium-trend Dating of Quaternary Sediments: U.S. Geological Survey Open-File Report 80-1087, 76 p.

Rutter, N.W., Crawford, R.J., and Hamilton, R., 1980, Correlation and relative age dating of Quaternary strata in the continuous permafrost zone of northern Yukon with D/L ratios of aspartic acid of wood, freshwater mollusks and bone, in Biogeochemistry

Ryer, T.A., 1977, Patterns of Cretaceous shallow-marine sedimentation, Coalville and Rockport areas, Utah: Geological Society of America Bulletin, v. 88, p. 177-188.

Sancetta, C., and Robinson, S.W., 1983, Diatom Evidence on Wisconsin and Holocene Events in the Bering Sea: Quaternary Research, v. 20, n. 2, p. 232-245.

Schmidt, R.A.M., 1967, New Generic Assignments for some Pleistocene Ostracoda from Alaska: Journal of Paleontology, v. 41, n. 2, p. 487-489.

Schmidt, R.A.M., and Sellman, P.V., 1966, Mummified Pleistocene ostracods in Alaska: Science, v. 153, p. 167-168.

Schmoll, H.R., Szabo, B.J., Rubin, M., and Dobrovolny, E., 1972, Radiometric dating of marine shells from the Bootlegger Cove Clay, Anchorage area, Alaska Geological Society of America Bulletin, v. 83, n. 4, p. 1107-1114.

Schrader, F.C., 1904, A reconnaisance in northern Alaska: U.S. Geological Survey Professional Paper 20, 139 p.

Schroeder, R.A.,, and Bada, J.L., 1976, A review of the geochemical applications of the amino acid racemization reaction: Earth-Science Reviews, v. 12, p. 347-391.

Sejrup, H.P., Miller, G.H., Brigham-Grette, J., Lovlie, R., Hopkins, D.M., 1984, Amino acid epimerization implies rapid sedimentation rates in Arctic Ocean cores: Nature, v. 310, p. 772-775.

Sellman, P.V., and Brown, J., 1973, Stratigraphy and diagenesis of perennially frozen sediment in the Barrow, Alaska, region, in Permafrost--the North American contribution to the second International Conference, Yakutsk: Washington, D.C., National Academy of Sciences, p. 171-181.

Sellman, P.V., Brown, J., Lewellen, R.I., McKim, H., and Merry, C., 1975, The classifiction and geomorphic implications of thaw lakes on the Arctic coastal Plain, Alaska: CRREL Research Report 344, 21 p.

Shackeleton, N.J., and Updyke, N.D., 1973, Oxygen isotope and paleomagnetic stratigraphy of equatorial Pacific core V28-238: Oxygen isotope temperatures and ice volumes on a 10^5 and 10^6 year scale: Quaternary Research, v. 3, p. 39-55.

Shackleton, N.J., and Opdyke, N.D., 1977, Oxygen isotope and paleomagnetic evidence for early Northern Hemisphere glaciation: Nature, v. 270, p. 216-219.

Shackleton, N.J., Backman, J., Zimmerman, H., Kent, D.V., Hall, M.A., Roberts, D.G., Schnitker, D., Baldauf, J.G., Desprairies, A., Homrighausen, R., Huddlestun, P., Keene, J.B., Kaltenback, A.J., Krumsiek, K.A.O., Morton, A.C., Murray, J.W., and Westburg-Smith, J., 1984, Oxygen isotope calibration of the onset of ice-rafting and history of glaciation in the North Atlantic region: Nature, v. 307, p. 620-623.

Sharma, G.D., 1979, The Alaskan Shelf: Hydrographic, Sedimentary and Geochemical Environment: Springer-Verlag, New York, 498 p.

Short, A.D., 1975, Offshore bars along the Alaskan Arctic Coast: Journal of Geology, v. 83, n. 2, p. 209-221.

__________, 1979, Barrier Island Development along the Alaskan Yukon coastal plains: Summary: Geological Society of America, pt. 1, v. 90, p. 3-5.

Short, A.D., Coleman, J.M., and Wright, L.D., 1974, Beach dynamics and nearshore morphology of the Beaufort Sea Coast, Alaska, in The Coast and Shelf of the Beaufort Sea, J.C. Reed and J.E. Sater, (eds.): Arctic Institute of North America Procceding, p. 477-488.

Skelkregg, L.L., ed., 1975, Alaska Regional Profiles: Arctic Region: University of Alaska Arctic Environmental Information and Data Center, Fairbanks, p. 10-24.

Smith, P.S., Mertie, J.B., Jr., 1930, Geology and mineral resources of northwestern Alaska: U.S. Geological Survey Bulletin 815, 351 p.

Stearns, H.T., 1978, Quaternary Shorelines in the Hawaiian Islands: Bernice T. Bishop Museum Bulletin, v. 237, 57 p.

Swain, F.M., 1963, Pleistocene Ostracoda from the Gubik Formation, Arctic Coastal Plain, Alaska: Journal of Paleontology, v. 37, n. 4, p. 798-835.

Swift, D.J.P., 1968, Coastal erosion and transgressive stratigraphy Journal of Geology, v. 76 p. 444-456.

__________, 1975, Barrier island genesis: evidence from the Middle Atlantic Shelf of North America: Sedimentary Geology, v. 14, p. 1-43.

Szabo, B.J., 1982, Age Determinations of Late Pleistocene Marine transgressions in western Alaska: Marine Geology, v. 46, p. M1-M8.

Szabo, B.J., Miller, G.H., Andrews, J.T., and Stuiver, M., 1981, Comparison of uranium-series, radiocarbon, and amino acid data from marine mollusks of Baffin Island, Arctic Canada: Geology, v. 9, p. 451-457.

Tappan, H., 1951, Northern Alaska index foraminifera: Contributions from the Cushman Foundation, v. 2, pt. 1, p. 1-8.

Thom, B.G., 1973, The dilemma of high interstadial sea levels during the last glaciation: Progress in Geography, v. 5, p. 170-246.

Todd, R., 1957, Foraminifera from Carter Creek, northeastern Alaska: U.S. Geological Survey Professional Paper 294-F.

Udvandy, M.D.F., 1963, Zoogeographic study of the Pacific Alcidae, in, J.L., Greissitt, ed., Pacific Basin Biogeography: Bishop Museum press, p. 85-111.

Vail, P.R., Mitchum, R.M., Jr., and Thompson, S., III, 1977, Seismic stratigraphy and global changes of sea level part 4: Global cycles of relative changes of sea level, in Payton, C.E., ed., Seismic stratigraphy - Applications to hydrocarbon exploration: American Association of Petroleum Geologists Memoir 26, p. 83-97.

Vigdorchik, M.E., 1980, Arctic Pleistocene history and the development of submarine permafrost: Westview Press, Boulder, Colorado, 286 p.

Vilks, G., Wagner, F.J.E., and Pelletier, B.R., 1979, The Holocene marine environment of the Beaufort Shelf: Geological Survey of Canada Bulletin 303, 43 p.

Vincent, J.-S., 1982, The Quaternary History of Banks Island, N.W.T., Canada: Geographie physique et Quaternaire, V. XXXVI n. 12, p. 209-232.

Wagner, F.J., 1977, Recent mollusc distribution patterns and paleobathymetry, southeastern Beaufort Sea: Canadian Journal of Earch Sciences, v. 14, p. 2013-2028.

Wahrhaftig, C., 1965, Physiographic provinces of Alaska: U.S. Geological Survey Professional Paper 482, 52 p.

Walker, D.A., 1982, Vegetation and Terrain Chapter III, in Proposed Oil and Gas Exploration within the Coastal Plain of the Arctic National Wildlife Refuge, Alaska: Draft Environmental Impact Statement, U.S. Fish and Wildlife Service, Dept. of Interior, p. III 10-III36.

Washburn, A.L., 1980, Cryogeology: Academic Press, New York, p. 406.

Webber, E.J., 1947, Stratigraphy and structure of the area of the Meade and Kuk Rivers and Point Barrow, Alaska: U.S. Geological Survey, Geological Investigations of the Naval Petroleum Reserve No. 4, Alaska, Report No. 6, p. 9.

Webber, P.J., Miller, P.C., Chapin, F.S., III, McCown, B.H., 1980, The Vegetation: Pattern and Succession, in Brown, J., Miller, P.C., Tiezin, L.L., and Bunnell, F.L., An Arctic Ecosystem: The Coastal Tundra at Barrow, Alaska Stroudsburg, Pennsylvania, Dowden, Hutchinson and Ross, Inc., p. 186-218.

Wehmiller, J.F., 1980, Intergeneric Differences in Apparent Racemization Kinetics in Mollusks and Foraminifera: Implications for Models of Diagenetic Racemization, in Biogeochemistry of Amino Acids, Hare, P.E., Hoering, T.C., and King, K., Jr. ed.: John Wiley and Sons, New York, p. 341-355.

Wehmiller, J.F., 1982a, A review of Amino Acid Racemization Studies in Quaternary Mollusks: Stratigraphic and Chronologic Applications in coastal and interglacial sites, Pacific and Atlantic Coasts, United States, United Kingdom, Baffin Island, and Tropical Islands: Quaternary Science Reviews v. 1, p. 83-120.

__________, 1982b, Interlaboratory comparison of amino acid enantiomeric ratios in fossil mollusks: Geological Society of America Abstracts with Programs v. 14, n. 7, p. 644.

Wehmiller, J.F., and Belknap, D.F., 1978, Alternative Kinetic Models for the Interpretation of Enantiomeric Ratios in Pleistocene Mollusks: Examples from California, Washington and Florida: Quaternary Research, v. 9, p. 330-348.

Wehmiller, J.F., Lajoie, K.R., Kvenvolden, K.A., Petersen, E. Belknap, D.F., Kennedy, G.L. Addicott, W.O. Vedder, J.G., and Wright, R.W., 1977, Correlation and chronology of Pacific coast marine terrace deposits of continental United States by fossil amino acid stereochemistry-technique evaluation, relative ages, kinetic model ages, and geologic implications: U.S. Geological Survey Open-File Report 77-780, 200 p.

Wiggins, I.L., 1951, The distribution of vascular plants on polygonal ground near Point Barrow, Alaska, Contributions, Dudley Herbarium Stanford University, v. 4, 41-56.

Wiggins, I.L., and Thomas, J.H., 1962, A Flora of the Alaskan Arctic Slope, Toronto, Ontario, University of Toronto Press. 425 p.

Williams, D.F., Moore, W.S., Fillon, R.H., 1981, Role of glacial Arctic Ocean ice sheets in Pleistocene oxygen isotope and sea level records, Earth and Planetary Science Letters, v. 56 p. 157-166.

Williams, J.R., 1979, Stratigraphy of the Gubik Formation at Skull Cliff, northern Alaska, U.S. Geological Survey Circular 804-B, p. B31-B33.

__________, 1983a, Geologic map of the Meade River quadrangle, northern Alaska, U.S. Geological Suvey Open-File Report 83-294, 50 pp.

__________, 1983b, Engineering-geologic maps of northern Alaska, Meade River quadrangle: U.S. Geological Survey Open-File Report 83-294, 30 p. 1 pl., scale 1:250,000.

Williams, J.R., Yeend, W.E., Carter, L.D., and Hamilton, T.D., 1977, Preliminary surficial deposits map of National Petroleum Reserve--Alaska: U.S. Geological Survey Open-File Report 77-868, 2 sheets, scale 1:500,000.

Williams, J.R., and Yeend, W.E., 1979, Deep thaw lake basins of the inner Arctic Coastal Plain, Alaska, U.S. Geological Survey in Alaska--Accomplishments during 1978: U.S. Geological Survey Circular 804-B, p. B35-B37.

Williams, K.M., and Smith, G.G., 1977, A critical evaluation of the application of amino acid racemization to geochronology and geothermometry, in International symposium on Origins of Life, 8: Dordrecht, Netherlands, Reidel Publishing Company, p. 91-144.

Wiseman, W.J., Jr., Coleman, J.M., Gregory, A., Hsu, S.A., Short, A.D., Suhayda, J.N., Walters, C.D., Jr., and Wright, L.D., 1973, Alaskan Arctic coastal processes and morphology; Louisianna State University, Coastal Studies Institute Technical Report No. 149, 171 p.

Young, F.G., and McNeil, D.H., 1982, Cenozoic stratigraphy of Mackenzie Delta, Northwest Territories, NTS 107B, 107C: Geological Survey of Canada Open-File Report 872.

Zeuner, F.E., 1959, The Pleistocene Period: its climate, chronology and faunal successions, 2nd ed.: London Press.

APPENDICES

APPENDIX A.1

MOLLUSK CLEANING PROCEDURE

For Samples Larger Than 100 mg

1) The sample is identified to the generic or specific level. Distinguishing characteristics of each specimen are noted (large, thin, abraded, paired, etc.).

2) The shells are cleaned of adhering sediment and periostracum either mechanically, or in an ultra-sonic clearner.

3) A shell fragment (between 100-300 mg) is weighed to the nearest 10 mg and placed in a clean 18 X 150 mm culture tube. The container is labelled with an AAL#, sample species, and weight.

4) Enough 2N HCL is added to dissolve 1/3 of the shell ("X" mg x $0.0033 \frac{ml}{mg}$ = "n" ml to add). Distilled water is added to totally cover the sample in the solution.

5) The samples are placed in a vacuum centrifuge for 5 to 10 minutes to ensure the reaction goes to completion.

6) The solution is poured-off and the shell is rinsed 3 to 4 times in distilled water with vigorous shaking. A final rinse is made in ultra-pure water. (The samples are then air dried.)

For Samples Between 90-110 mg

The cleaning procedure above is used, but 0.001 ml/mg of 2N HCL is added to the sample instead of leaching 1/3 of the shell.

For Samples Less than 90 mg

1) The sample is placed in a clean plastic 7 Dram snap-cap vial and identified. The vial is labelled with species and AAL No.

2) Enough Sodium Metaphosphate ($(NaPO_3)_{13}$) solution is added to cover the shell with 10 mm of solution. This is placed in an ultrasonic cleaner for 30 seconds. This is repeated until the shells are clean.

3) The shell is then rinsed twice in distilled water and twice in ultra-pure water in the ultrasonic cleaner.

4) The samples are dried in a vacuum desiccator.

APPENDIX A.2

1981 SAMPLE PREPARATION FOR MOLLUSCS

1) An approximtely 50 mg fragment is weighed accurately to the nearest 0.1 mg. It is then placed in a sterile 18 X 150 mm culture tube.

2) A solution of 2N HCL and Norleucine is added by syringe and sterile pipette (0.014 ml per mg of shell). The culture tube is placed in the vaccum centrifuge for 10-15 minutes.

3) The sample is sonicated for 20 seconds and split into two 0.20 ml aliquots using an automatic pipette, and placed in two 3.7 ml vials, one labelled with the AAL number, species and FREE and then one labelled with the AAL number, species and HYD. The HYD cap should have a teflon liner.

4) The FREE vials are placed in the vacuum desiccators to dry.

5) Exactly 200 µl 12N HCL are added to each HYD vial with an automatic pipette. The vial is flushed with N_2 gas for 15 seconds and sealed tightly.

6) The HYD vials are pyrolysed for 22 hours at 110°C.

7) The HYD's are spun-down and the caps removed. The vials are placed in the vacuum desiccator to dry. FREE's and HYD's should not be in the same desiccator.

8) When dry, the FREE's and HYD's are stored together in numerical order.

9) When samples are ready to be analyzed, both FREE and HYD's vials re rehydrated with 0.57 ml of pH2 solution.

For samples Weighing 30-40 mg After Cleaning

1) The same preparation is used as above except only 0.15 ml are transferred for FREE and HYD fractions and only 0.15 ml 12N HCL is added to the HYD. Both are rehydrated with 0.43 ml pH2 solution.

For Samples Weighting 20-30 mg

1) The same preparation as above is used, except only 0.1 ml is transferred to the HYD and FREE fraction. Only 0.1 ml 12N HCL is added to the HYD fraction. Both HYD and FREE are rehydrated with 0.28 ml pH2 solution.

<u>For Samples 10-20 mg</u>

1) The same preparation is used as above, but only 0.1 ml is transferred to a HYD fraction only (no FREE). 0.10 ml 12N HCL is added. The sample is rehydrated with 0.28 ml pH2 solution.

<u>For Samples Less than 10 mg</u>

1) 0.14 ml 2N HCL + Norleucine is added to the sample. All of the liquid is transferred into a HYD vial, 0.24 ml of 12N HCL is then added. The vial is labelled with AAL No., species, HYD and the exact weight. Rehydrate with $\left(\frac{\text{sample eight}}{25}\right)$ ml of pH2.

APPENDIX A.3

"76B PREP" FOR MOLLUSKS

1) A cleaned fragment is fractured. The sample is weighed out in two pieces (or groups of pieces); each about 25 mg. Both are weighed to the nearest 0.1 mg.

2) The groups are transferred to two vials. One vial is labelled with AAL No., species and FREE. The other is labelled with the AAL No., species and HYD. The shell fragment's weights are labelled on their respective vials.

3) For both FREE and HYD vials:

$$\left(\frac{\text{.02 ml 7N HCL + Norleu}}{\text{1 mg shell}}\right) \times \text{Sample Weight in mg} = \text{ml 7N HCL + Norleu Added}$$

4) (a) FREE vials are dried in a vacuum desiccator

   (b) HYD vials are flushed with N_2 gas and then sealed. They are pyrolyzed for 22 hours at 110°C. They are then dried in a vacuum desiccator.

5) Both vials are rehydrated with pH2 solution at a ratio of $\left(\frac{\text{Weight of sample}}{0.25}\right)$ ml pH2.

APPENDIX B

Amino Acid Data

APPENDIX B.1

AAL#s OF SAMPLES PREPARED USING 76B PREPARATION METHOD
(Note: Computer printouts are compiled in the following pages in numerical order.)

MODERN

2871

3448

WALAKPA-AMINOZONE 1

2797	2830
2799	2838
2800	2854
2806	2855
2822	3563

KARMUK-AMINOZONE 2

2801	2853
2802	2895
2803	2896
2804	2897
2805	2927
2807	2928
2808	
2809	2957
2810	2988
2833	3193
2834	3197
2835	3412

TUAPAKTUSHAK-AMINOZONE 3

	2990
2811	3194
2812	3195
2831	3201
2836	3202
2837	3437
2893	3438
2894	3439
2900	3441
2926	3444
2956	3564
2989	3565

KILLI CREEK-AMINOZONE 4

2832	3203
3196	3440
3198	3562

KARMUK-AMINOZONE 2

2839	3413
2840	3445
2851	3446
2852	3447
	3566

NUVALIK-AMINOZONE 5

2898	3200
3199	

SIBERIAN SAMPLES

(Note: These samples (except 1450) were reprepared under the 76b method but the old AAL#s from the 81 prep were retained.)

1450	1454
1451	1455
1452	1456
1453	1457
	1458

APPENDIX B.2

AAL#S OF SAMPLES PREPARED USING THE 81 PREPARATION METHOD

1685	1921	1970	2549
1686	1922	1971	2550
1687	1923	2026	2551
1898	1924	2027	2552
1892	1925	2028	2554
1893	1941	2029	2555
1894	1942	2030	2577
1895	1943	2031	2578
1896	1944	2032	2579
1897	1945	2033	
1899	1946	2055	
1900	1947	2056	
1901	1948	2057	
1902	1949	2058	
1903	1950	2059	
1904	1951	2060	
1905	1952	2061	
1906	1953	2062	
1907	1954	2063	
1916	1955	2064	
1917	1966	2093	
1918	1967	2100	
1919	1968	2101	
1920	1969	2548	