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EXTENT, TIMING, AND PALEOGEOGRAPHIC SIGNIFICANCE OF MULTIPLE PLEISTOCENE GLACIATIONS IN THE BERING STRAIT REGION

A

THESIS

Presented to the Faculty
of the University of Alaska Fairbanks

in Partial Fulfillment of the Requirements
for the Degree of

DOCTOR OF PHILOSOPHY

By

Patricia Anne Heiser B.S.

Fairbanks, Alaska
December 1997

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EXTENT, TIMING, AND PALEOGEOGRAPHIC SIGNIFICANCE OF MULTIPLE PLEISTOCENE GLACIATIONS IN THE BERING STRAIT REGION

by

Patricia Anne Heiser

RECOMMENDED:

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ABSTRACT

This study utilizes a multidisciplinary approach to the investigation of the extent, timing, and potential effects of repeated Pleistocene glaciation in Bering Strait region. A major focus of this study was directed toward testing the hypothesis that a continental-scale ice sheet existed in Beringia during the Late Wisconsin glacial period. Satellite synthetic aperture radar (SAR) imagery was used to compile a map of glacial moraines in Chukotka, Russia, and to attempt preliminary correlations with the glacial record in Alaska. Geophysical modelling of the solid-earth response to postulated glacial loading, and the reconstruction of regional snowline were combined with the results of the SAR investigation to test the ice sheet hypothesis. Finally, a detailed study of the Quaternary stratigraphy and surficial geology of St. Lawrence Island was used to correlate the glacial and sea level histories of western Alaska and Chukotka, Russia.

The sequences of moraines in Chukotka, mapped from SAR imagery, are similar in morphology and position to moraine sequences described in Alaska, recording a succession of glacial events that most likely began in the middle Pleistocene and ended with the Late Wisconsin. The record of repeated mountain glaciation, characterized by radial flow out of high topographic areas provides strong evidence against the existence of a southward-flowing, continental-scale ice sheet in Beringia at any time in the latter part of the Pleistocene.

Geophysical modelling of the solid-earth response to glacial loading predicted relative sea level changes on the scale of meters to tens of meters (rising or falling

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depending on forebulge effect) around the shores of present-day Bering Strait if a large ice sheet had, indeed, occupied the Beringia during Late Wisconsin time. There is no evidence of these predicted sea level changes anywhere in the region. The reconstruction of Late Wisconsin snowlines in Russian and Alaska show that the paleoclimatic conditions needed to 'grow' the hypothesized ice sheet did not exist.

Field mapping and stratigraphic work on St. Lawrence Island revealed that ice advanced onto the island twice in the late Pleistocene, once in the Middle Pleistocene and once after the Last Interglacial, probably during the Early Wisconsin. The record of glaciers advancing from Chukotka onto the island provides an important 'Rosetta Stone' for correlating the glacial histories of northeast Siberia and Alaska.

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The completion of this dissertation would have been impossible without the help, encouragement, and contributions of many colleagues, faculty members, family and friends. My advisor David Hopkins shared with me his enthusiasm, insight, vast store of knowledge, and especially his love for Beringian natural history. Thank you to the many other members of my committee, Craig Lingle, Mary Edwards, Ken Dean, Will Harrison, and Jim Beget for their careful reviews and comments on my research. Craig Lingle must be distinguished for his patience and diligence in helping me understand the complex principles of crustal deformation and Greens Functions. Jim Clark of Calvin College kindly provided his calculated Greens Functions that I used in my isostatic model. Thanks also go to Julie Brigham-Grette of Univ. Massachusetts, whose collaboration and enthusiasm made this project a reality.

I was supported financially from many sources through the course of this research. A grant form the National Science Foundation (OPP-15234) supported me for three years. The National Park Service provided support on this project and several others. The Graduate School of the University of Alaska also supported me with a Dissertation Year Fellowship, without which, this thesis may never have been completed. Research and logistical support was also provided by GSA Student Research Grant, National Park Service, the Alaska Quaternary Center and the Geist Fund of the UAF Museum.

This work was greatly enhanced by the company and field assistance of Robert Tungiyan and other residents of St. Lawrence Island. Robert, his brother Rubin, Dexter

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CHAPTER 1

INTRODUCTION

A major aim of Quaternary geology is to describe the extent and timing of Pleistocene climate change on a global basis. This aim is motivated by the necessity of understanding the history of climate change before we can decipher its causes and predict its future courses. The temporal and spatial patterns of past climate changes are often critical in deciphering causation.

The prehistoric fluctuations of glaciers are striking manifestations of climate change. Glaciers periodically covered large areas of the northern hemisphere leaving abundant geological evidence of their former distributions. In North America and Europe, glacial stratigraphy and landforms have been extensively studied, and the timing and extent of Pleistocene glacial events there are relatively well understood. However information on glacial history is quite limited in other parts of the world due to geographical and political isolation. For instance, northeastern Russia and Alaska share similar environments today and probably experienced similar histories of climate change through the Pleistocene. However, correlations of glacial history across the Bering Strait have been hampered by political and logistical obstacles, as well as by differences in scientific methodologies and technology between two nations. Scant and conflicting information from the Russian side of Bering Strait has led to widely divergent hypotheses regarding glacial extent in

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Beringia, (those areas of the Yukon, Alaska, and northeastern Russia which were largely unglaciated during the last ice age) (Hopkins 1966).

Here I report on a multidisciplinary investigation of glacial history in the Bering Strait region, particularly in the Chukotka region of northeastern Siberia. I collected stratigraphic and other field data, used satellite remote sensing to map glacial landforms in Russia, and applied a geophysical model to predict the solid earth response to glacial loading. The results of this study contribute to the continuing effort to understand the glacial and climate history of the Bering Strait region.

Background

Beringia is a broad biogeographic region bounded on the west by the Kolyma River in Russia and on the east by the MacKenzie River in Canada. The region is presently divided by the Bering Strait, which connects the Pacific and Arctic oceans. The appearance of North Atlantic molluscs in the stratigraphy of Karaginsky Island off the coast of Kamchatka Peninsula, records the first opening of Bering Strait about 4 million years ago (Gladenkov *et al.*. 1991). Since then the region has experienced a number of marine transgressions and regressions that fluctuated in concert with changes in global sea level. The Bering Strait region has thus served as an important episodic link for marine faunal dispersal between the Atlantic, Arctic and Pacific Oceans, and alternately, as a link for terrestrial flora and fauna between Asia and North America.

Marine and glacial sediments exposed along the shores of Chukotka, western Alaska, and the islands of Bering Strait record the changes in sea level and glacial extent

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that occurred in Beringia through the Late Cenozoic. The development of a comprehensive framework for the history of the marine transgressions in Beringia has been the lifework of David M. Hopkins (1959, 1965, 1967, 1973, 1982), and the timing of these events has been continuously refined with the development of new geochronologic methods and technology (Brigham-Grette and Hopkins 1995, Kaufman and Brigham-Grette 1993, Kaufman *et al.*. 1991). Because changes in global sea level are induced by the growth and retreat of continental ice sheets, the sediments deposited during interglacial high sea level events are often interbedded with glacial deposits, which allows the glacial events to be dated as well (Kaufman *et al.*. 1991). The stratigraphic study of these deposits, along the shores of western Alaska, has formed the basis for of our knowledge about the glacial and sea level history of the Bering Strait region.

The record of these events is probably similar on the Russian side. However, until recently, political differences prevented collaborative work with Russian geologists, and thus has hindered attempts to correlate the stratigraphic records across Bering Strait. The recent and fortuitous raising of the 'Iron Curtain' has allowed the development of cooperative research projects between Russian and American scientists. This work is part of a collaborative project funded jointly by U.S. National Science Foundation (#DPP-9015234), the Russian Academy of Sciences, and the U.S. National Park Service.

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The Glacial Record in Northwest Alaska

The lowering of sea level that accompanied glacial periods caused the shallow continental shelf of the Bering and Chukchi Seas to be exposed and form what is called the Bering Land Bridge (Figure 1.1). Because this land bridge was extant only during glacial periods, the timing, extent, and isostatic effect of glacial advances in this region were important controls on the dispersals of plants, animals, and early humans between Asia and North America (Hopkins 1982).

In the mountainous areas of Beringia, Pleistocene glacial events are evidenced by sequences of moraines deposited at and beyond valley mouths. These moraines are radiocarbon dated when possible. However, most of the age determinations are done using relative age techniques such as moraine morphology (Kaufman and Calkin 1988, Hamilton *et al.* 1986, Kaufman and Hopkins 1986).

In northwest Alaska, the Pleistocene glacial record begins with the Sinuk River glaciation that advanced from the Kigluaik Mountains and reached the present coast of the Seward Peninsula, and the Anaktuvuk glaciation known from the northern front of the Brooks Range (Hamilton 1994). In both places, this early event is the most extensive glaciation evidenced on the landscape. The glacial limit is marked by very low relief moraines, drift, and erratic boulders in both regions. Early Pleistocene drift is also exposed in mine shafts, placer pits and stream banks in several places near Nome (Hamilton 1994, Kaufman *et al.* 1991, Kaufman and Hopkins 1986). The presence of tors and altiplanation terraces in areas glaciated during this interval, in the Brooks Range and on the Seward Peninsula, suggest that they are correlative and much older than subsequent glacial

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advances (Hamilton 1994). This glacial event is assumed to be early Pleistocene in age (Hamilton 1994).

About 400,000 years ago, an interglacial marine transgression inundated the coastal plain at Nome. Named the Anvillian Transgression, it left fossiliferous marine sand and gravel at altitudes up to 22 meters above present sea level (Kaufman and Brigham-Grette 1993, Kaufman *et al.* 1991). Amino acid geochronologic studies and radiometric dating of Nome River glacial deposits (see below) which overlie Anvillian sediments on the Nome coastal plain, have placed this transgression in Middle Pleistocene time (Kaufman and Brigham-Grette 1993, Kaufman *et al.* 1991). Comparison with the marine oxygen isotope record led Kaufman and colleagues to further constrain the age to this high-sea level event to isotope stage 11, which is thought to be the last time sea level was as high as the shorelines of the Anvillian Transgression (Kaufman *et al.* 1991). Marine sediments of this age have also been described from Kotzebue Sound, the Arctic Coastal Plain, and St. Lawrence Island (Hopkins *et al.* 1972, Huston 1990, Kaufman *et al.* 1991, Kaufman and Brigham Grette 1993)

The Nome River glaciation followed the Anvillian transgression and, at Nome, extended more than 50 kilometers from the Kigluaik Mountains and terminated beyond the present shoreline leaving well preserved moraines and drift on the coastal plain at Nome (Kaufman and Hopkins 1986). The Nome River glaciation is firmly placed in the Middle Pleistocene by radiometric dates on basalts that directly overlie moraines from this glaciation (Kaufman *et al.* 1991). At Minnie Creek, a basalt flow above the drift was analyzed using $^{40}Ar/^{39}Ar$ laser fusion and dates to 470,000 $\pm$ 190,000. The well estabished

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age of Anvillian sediments, indicates that the Nome River glaciation occurred after 400,000 years BP (Kaufman *et al.* 1991).

The extent of the Nome River glaciation was nearly an order of magnitude greater than any later Pleistocene glacial advance (Kaufman *et al.* 1991). In the Brooks Range, Sagavinirktok moraines mark a similar advance of ice less extensive than the preceding Anaktukvuk, but more extensive than any subsequent glaciation (Hamilton 1994). On the Baldwin Peninsula (which forms the eastern shore of Kotzebue Sound) marine, glaciomarine, and fluvial sediments of Anvillian age were deformed and overridden by glacial lobes that advanced out of the Noatkak, Selawik, and Kobuk valleys of the southwestern Brooks Range (Huston *et al.* 1990). Hopkins described marine deposits of middle Pleistocene age on St. Lawrence Island that were deformed by glaciers advancing from the Chukotka Peninsula of Russia (Benson 1994, Hopkins *et al.* 1972). This "Nome River" event is the oldest, well dated, and widely recognized Pleistocene glacial advance in Beringia. The discovery of Nome River age deposits on St. Lawrence Island provided the first distinct evidence that the glacial histories of Russia and Alaska could be correlative in timing and extent (Hopkins *et al.* 1972).

The next climatically driven geomorphic event to dramatically affect the Beringian landscape was the Pelukian marine transgression that accompanied the Last Interglacial warm period. Pelukian shorelines are widespread along the present shores of Beringia at an average of 7-10 meters a.s.l.. Mollusk fauna in these sediments suggest that the oceans were warmer than at present (Kaufman and Brigham-Grette 1993, Brigham-Grette and Hopkins 1995). Amino acid geochronologic studies and morphostratigraphic relationships

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correlate this event with the last Interglacial period which occurred 120,000-130,000 years before present. Pelukian sediments and wave cut shorelines are found on the north slope of Alaska, Russia, and on St.Lawrence Island (Kaufman and Brigham-Grette 1993, Brigham-Grette and Hopkins 1995).

The glacial advances that followed the Middle Pleistocene event were more than an order of magnitude less extensive. On the Seward Peninsula, post-Nome River advances are identified in high mountain valleys. The Stewart River glaciation is the older of these, and advanced 5-10 km beyond Salmon Lake on the south side of the Kigluaik Mountains (Kaufman and Hopkins 1986). The Stewart River is represented by poorly dated, and poorly understood, drift masses and end moraines. The Salmon Lake glaciation formed lobes that extended beyond the mountain front of the Kigluaiks, and is recognized on both sides of the range (Kaufman and Hopkins 1986). The Stewart River and Salmon Lake glaciations are correlated with the Sagavinrktok and Itkilik I advances in the Brooks Range and probably date from Isotope stage 6, and the early Wisconsin, respectively (Hamilton 1994, Hopkins pers. comm. 1997). Some time after the Last Interglacial, during the Early Wisconsin or possibly oxygen isotope Stage 5b or 5d, glaciers again advanced from Chukotka and onto St. Lawrence Island (Benson 1994, Brigham-Grette *et al.* 1992, Heiser *et al.* 1992).

The Late Wisconsin glaciation, which is characterized by the encroachment of ice sheets over large parts of North America and Europe, was quite limited in extent in northwestern Alaska (Mann and Hamilton 1995, Hamilton 1994, Kaufman and Hopkins 1986, Hopkins 1972). The global lowering of sea level that accompanied the late

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Wisconsin glacial period caused the shallow continental shelf of the Bering and Chukchi Seas to become emergent and form a land bridge between Asia and the North America. It is believed that the cold and relatively dry climate of the full glacial period (Hopkins 1972) and the removal of a nearby coastal moisture source restricted the growth of glaciers on both sides of Bering Strait, and left the land bridge open as a route for ice age dispersal of plants, animals and humans between Asia and the New World.

The glacial record for the Late Wisconsin (24,000-12,000 BP) in Northwest Alaska corresponds well with the paleoclimatic data that suggests a cold and dry environment. In the Brooks Range, Late Wisconsin age glaciers (Itkillik II) extended only 5-25 kilometers beyond the mountain front (Hamilton 1986). The Kigluak Mountains on the Seward Peninsula support a similar record of Late Wisconsin glaciation. Here glaciers of The Mt. Osborn glaciation advanced only a few kilometers from their cirque wall and formed under an equilibrium line lowering of only 200 meters (Kaufman and Hopkins 1986).

In summary, there are four major glacial episodes (probably 5) recognized in northwestern and other regions of Alaska. These glaciations are probably correlative (Table 1.1) and date from the Middle Pleistocene, possibly isotope stage 6, then late isotope stage 5 or stage 4 (80,000-120,000 yBP), and finally the Late Wisconsin (12,000-24,000 BP). Figure 1.2 is a map showing the approximate extent of these glaciations.

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The Glacial Record in the Russia Far East

The glacial record in the Russian Far East has been based primarily on stratigraphic evidence found in Quaternary deposits on Chukotka Peninsula. Oleg. M. Petrov was one of the first researchers to put together a stratigraphic scheme for Chukotka (Petrov 1966). Until now, it has been difficult to correlate glacial and marine events across Bering Strait because nomenclature for stratigraphic schemes has changed frequently as new workers and new methods have been applied to the problem.

One of the oldest glaciations recognized in Chukotka is called the Kresta (Petrov 1966, Ivanov 1982) and is probably Middle Pleistocene in age. This is probably the most extensive glacial advance recorded in Chukotka. Coalescent valley glaciers, or piedmont lobes, advanced from the mountains of Chukotka and extended offshore in many places (Ivanov 1982). A small ice cap growing over the Chukotka mountains spread beyond the southeast coast of the peninsula and eventually encroached onto St. Lawrence, deforming sediments deposited by the Anvillian marine transgression (Benson 1994, Hopkins 1972). Farther west, glaciers of this same age (Middle Pleistocene) coalesced to form piedmont lobes, and spread onto the broad plains of the Anadyr Valley (Archipov *et al.* 1986). However, Olga Glushkov has assigned a Zyryankan (early Wisconsin equivalent) age to these moraines and does not recognize any glaciation older than this in the Anadyr lowland (Glushkova 1989, Glushkova 1994).

The Zyrankan glaciation (probably early Wisconsin) in northeast Siberia was characterized by coalescing alpine glaciers that, like the earlier glaciers, advanced off shore

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in some places. This advance was less extensive and left moraines in many of the lowlands of the region (Archipov *et al.* 1986, Glushkova 1989, Glushkova 1994).

The Sartan (Late Wisconsin) glaciation, according to most Russian geologists, was limited in extent and constrained to high mountain valleys (Bespally 1984, Archipov *et al.* 1986, Glushkova 1989, Ivanov 1986, Glushkova 1994). In the Tenyanni Mountains, south of Lavrentiya Bay, Sedov (1992) describes LGM aged moraines confined to cirques. In the Pekulney Mountains, near Anadyr lowland, a series of moraines, mapped from satellite imagery and airphotos, show the late Wisconsin advance to have been limited to high mountain valleys (Glushkova 1989, Glushkova 1994). This pattern was similar in the Verkhoyanks mountains to the west and the Koryaks to the south (Kind 1975, Archipov *et al.* 1986)

Continental Scale Ice Sheets in Beringia

The lack of consistent and easily correlatable information regarding the timing and extent of Pleistocene glaciations in the Russian Far East has led, recently, to a controversy over the thickness and extent of Late Pleistocene glaciers in Beringia. A few workers recently postulated that a series of large marine-based ice sheets. of continental scale, advanced southward from the Arctic continental shelf over much of Chukotka and through Bering Strait as far south as St. Lawrence Island (Grossvald 1988, Hughes and Hughes 1994, Hughes and Grossvald 1995). This model is in conflict with the field data collected from northwestern Alaska and, particularly, from Russia.

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The idea of a Bering Strait ice sheet appears to originate with the American naturalist John Muir, who visited both sides of the Bering Strait aboard the *Corwin* in 1881. Muir described glacial landforms and sediments in the Bering Strait region and interpreted them as evidence for a large ice sheet advancing from the north (Muir 1917). More recently, M.G. Grossvald proposed that the continental shelf of the Arctic Ocean fringing northern Russia supported marine-based ice sheets during the Late Wisconsin (Grossvald 1988). His ideas were inspired by the ice shelves of East Antarctica, the marine based ice domes of the West Antarctic ice sheet, and by a marine-based ice sheet known to occupy the Barents Sea during the Late Wisconsin (Boulton and Rhodes, 1974). An earlier version of Grossvald's vision of previous glaciation in Beringia included ice domes along the southern edge of the Bering Sea's continental shelf with fringing, Antarctic-style ice shelves (Grossvald and Vozovik, 1983).

In 1994, T. Hughes and B. Hughes presented the Marine Ice Transgression Hypothesis (MITH), which asserts that marine-based ice sheets developed on the continental shelves of the Arctic Ocean during Pleistocene glacial maxima and proposes mechanisms for their development. According to MITH, marine ice sheets form as perennial sea ice thickens through bottom freezing and eventually becomes grounded on the sea floor. Meteoric accumulation and run-off from rivers dammed along the landward margins of the growing ice mass contribute to the growth of the ice domes. Eventually the ice domes intersect the snow line and advance landward as ice sheets and seaward as floating ice shelves (Hughes and Hughes, 1994; Hughes, 1994). Invoking this dramatic

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formative mechanism, Grossvald and T. Hughes hypothesized the presence of several distinct ice sheets in the Russian Arctic and the Bering Strait region. For the LGM, they propose the existence of an ice sheet reaching 1500 meters thick over Wrangel Island and flowing south over Chukotka and through Bering Strait (Figure 1.3).

Objectives and Outline of this Dissertation

The present controversy, and the contributing lack of information on the Russian side, demand the development of a framework for the study of glacial history in northeast Siberia and development of a better correlation scheme with northwest Alaska. A considerable amount of work has been done in Chukotka by our Russian colleagues. However, conflicting conclusions among Russian investigators (e.g. Petrov 1966 and Ivanov 1986, Grossvald 1990 and Glushkova 1986) prevent an easy evaluation of glacial histories in the region using the existing literature. Furthermore, methods, materials, and style vary to such a degree that comparison with the work of American scientists is often difficult if not impossible. This is especially true when comparisons are made based on literature rather than joint field studies.

This study attempts to test directly the existing paradigms and hypotheses regarding late Wisconsin, as well as older, glacial extents in Beringia, and develops a framework for the correlation of glacial and sea level history across Bering Strait. This dissertation is written as a series of manuscripts joined by synthesizing Introduction and Conclusion. The manuscripts discuss 1) the use of satellite synthetic aperture radar (SAR) as a tool for mapping glacial moraines in Russia, 2) tests of the MITH using SAR,

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isostatic modelling, and snowline reconstuctions, and 3) the surficial geology and stratigraphy on St. Lawrence Island.

Satellite synthetic aperture radar imagery has proven useful for the study of glacial moraines in the Russian Far East. In Chapter 2, James Roush (presently of the National Park Service) and I use SAR imagery to compile a map of glacial landforms of the Chukotka Peninsula in Russia. In areas where we find moraine sequences we have attempted to estimate relative age using moraine morphology and position. In order to quantify morphologic characteristics, we developed a technique which enabled us to measure the surface slope angle of a moraine, a criterion used in the relative age dating of glacial moraines in Alaska and elsewhere (Kaufman and Calkin 1988, Hamilton et. al. 1986). While our age assignments are rough estimates at best, the resulting map of glacial landforms should provide an essential geographic framework for the increasing number joint US and Russian projects and investigations presently being initiated in the Russian Far East.

In Chapter 3, Craig Lingle, Daniel Mann, and I combine the SAR-based glacial map with glacial isotatic modelling and snowline reconstruction in a manuscript that addresses, directly, the Marine Ice Transgression Hypothesis (MITH) of Hughes and Grossvald (1995). We reconstruct the ice sheet postulated by the authors of the MITH, and geophysically model the expected solid earth response to that ice load. We then compare the results of the model with geologic evidence for changes in relative sea level during Late Wisconsin time. As anoter test of the MITH, we use SAR imagery and topographic maps to reconstruct valley glaciers and determine equilibrium line altitudes

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(ELAs) for the most recent glaciers (late Wisconsin) that occupied the high mountains of Chukotka and Alaska.

St. Lawrence Island lies almost directly in the center of the ice age land bridge. Glacial deposits on St. Lawrence indicate that glaciers from the Chukotka Peninsula flowed out of the mountains, across the continental shelf, and encroached upon the island. Since datable interglacial marine deposits are also found there, the island provides an important location for the investigation of the timing and extent of glacial advances from the mountains of Chukotka. The third manuscript (Chapter 4) included in this dissertation is my spatial and stratigraphic study of glacial sediments on northwestern St. Lawrence Island and is presented as a surficial geologic map with an expanded legend describing the map units and other features. An overview of the more significant stratigraphic sections, and geochronologic work done to date, is included.

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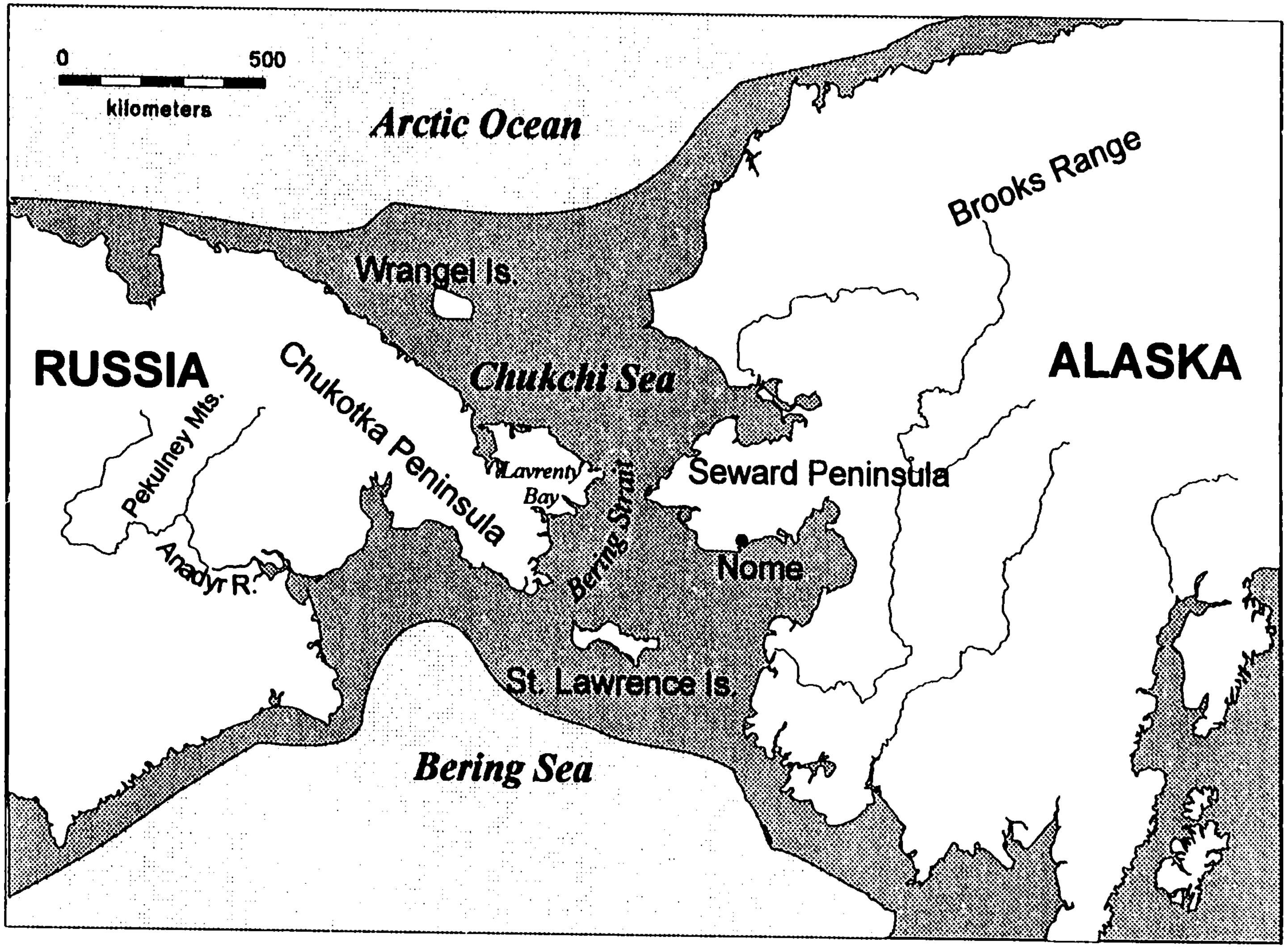

Figure 1.1 Map of Beringia showing locations mentioned in the text. Extent of the Late Wisconsin land connection is shown in dark grey.

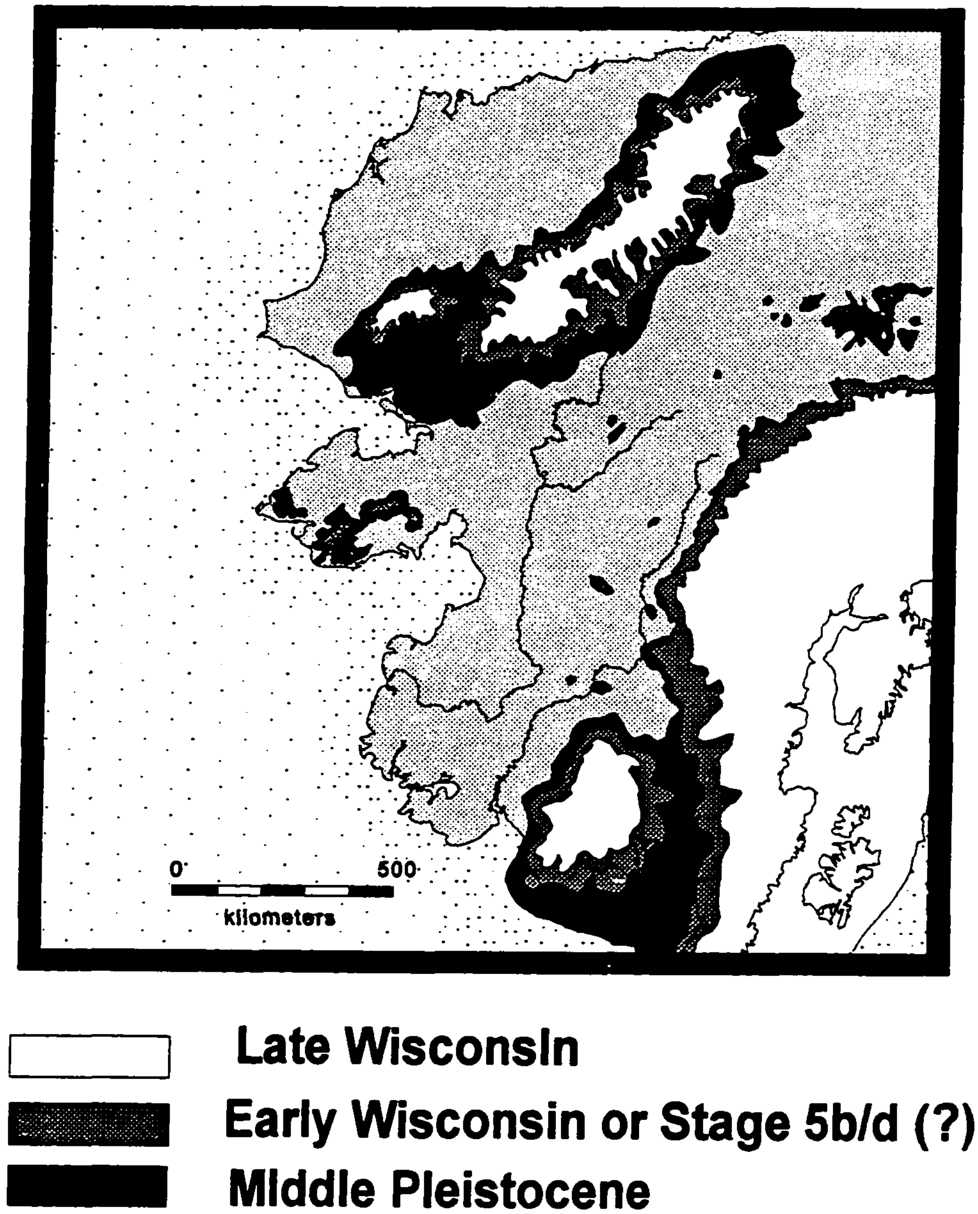

Figure 1.2. Extent of Pleistocene glaciations in Alaska.

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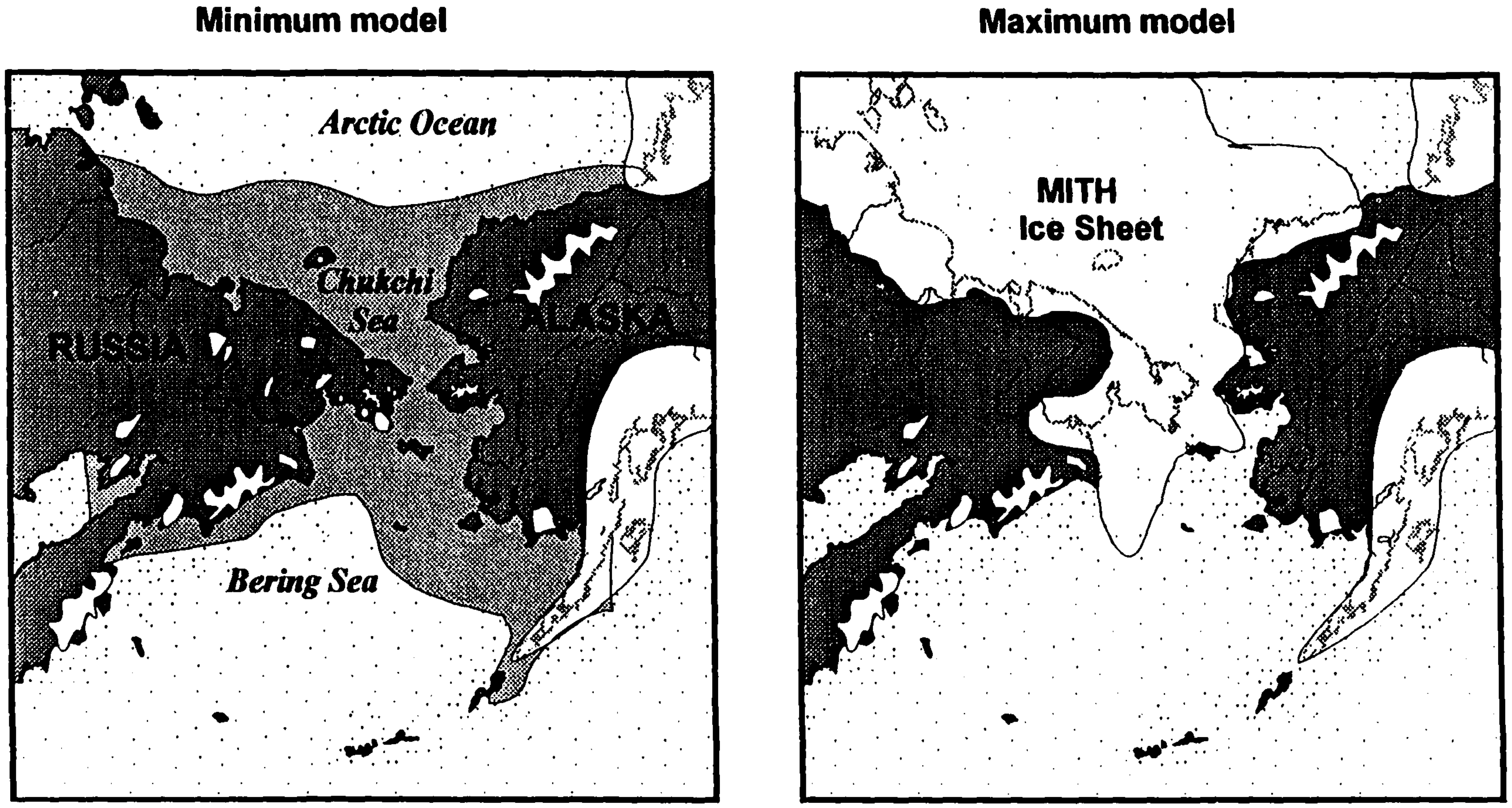

Figure 1.3. Two models of glacial extent during the Late Wisconsin. In the minimum model, the lowering of global sea level and emergence of the Bering Land Bridge, induced a drier continental climate in Beringia. Moisture starved glaciers advanced only slightly beyond mountain fronts. The postulated ice sheet in the maximum model was 'grown' in a numerical model that called for a regional snowline lowering of 1000 meters below present. The MITH calls for ice doming on the Chukchi Sea shelf and subsequent southward flow over Chukotka and the land bridge.

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Table 1.1. Correlation of glacial events in northwest Alaska.

Age/Event	Brooks Range		Seward Peninsula
Late Wisconsin	Itkillik II	Late Phase 13-11.5 ka Main Phase 24-13 ka	Mount Osborn 11.5 ka
Penultimate Glaciation	Itkillik I	> 40 ka Phase B Phase A	> 40 ka Salmon Lake Stewart River
Middle Pleistocene	Sagavinirktok	Late Phase Early Phase	Nome River 0.47 $\pm$ 0.05 Ma
Early Pleistocene	Anuktuvuk River		Sinuk

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CHAPTER 2

PLEISTOCENE GLACIATIONS IN CHUKOTKA, RUSSIA: MORAINE MAPPING USING SATELLITE SYNTHETIC APERTURE RADAR (SAR) IMAGERY

Abstract

Using satellite Synthetic Aperture Radar (SAR) imagery we are able to identify many glacial features on the Russian landscape that were previously unobservable by any scientists outside Russian institutes. Measurements from SAR imagery provide information on glacial extent, moraine morphology, and non-sedimentary glacial features such as valley morphology and flow direction. Moraine sequences in several areas of Chukotka record multiple Pleistocene glacial events. Several morphologic characteristics, such as relative degradation and surface slope can be determined from SAR imagery and used to estimate relative age of glacial features. By conducting comparative studies on moraines of *known* age in Alaska, we are able to estimate the ages of moraines in Russia and develop a framework for the correlation of glacial history across Bering Strait. Our study of these images suggests that the last major glaciation (Late Wisconsin) was much less extensive than older Pleistocene events, whose marks remain on the landscape beyond the younger moraines.

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Introduction

The determination of Pleistocene glacial limits is an important component in the study of ice ages and climate history. Correlations of glacial records on a global scale have assisted ongoing investigations of climatic change and serve as important boundary conditions for climate and sea level models (Manabe and Brocoli 1985, COHMAP 1988, Peltier 1994). In North America and Europe, glacial landforms and stratigraphy have been studied extensively, and the timing and extent of Pleistocene glacial events is relatively well understood (Denton and Hughes 1981). Political and geographic isolation, however, has limited the extent of our knowledge about glacial history in some other parts of the world. The Chukotka Peninsula of northeastern Siberia, for example, inhabits a physiographic and climatic environment similar to parts of Alaska, and may contain similar records of climate change through the Pleistocene (Figure 2.1). Little is known, however, about the extent and timing of Pleistocene glacial events in this region. Correlations of glacial history across the Bering Strait have been hampered by political and logistical obstacles, as well as by fundamental differences in scientific methodology and technology between two nations. An understanding of glacial history in the Bering Strait region is important to our understanding of the paleogeographic conditions of the Bering Land Bridge, which joined Asia and the New World during ice ages. Since the Land Bridge was emergent only during glacial periods, the timing and extent of glaciations in this region were of paramount importance to the dispersal of plants, animals, and early humans between Asia and North America (Hopkins et al. 1982).

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The first step in testing any model of glaciation is to map glacial landforms, such as terminal moraines, that record the extent of ancient glaciers. Typically, glacial maps are produced by first using topographic maps and airphotos to identify and locate glaciogenic features. Preliminary map work is then followed by field investigation to determine stratigraphic and age relationships. Because the Bering Land Bridge is part of both Russia and the United States, there is considerable difference in the form and degree of availability of map and airphoto coverage on the two continents. In Russia, available maps are of a scale too small to be useful to glacial mapping, airphoto coverage is often incomplete and photos are frequently not available outside of the research institutes in Russia. While some glacial geologic work has been done on the Russian side, methods and materials differ to such a degree that comparison with the work of American scientists is difficult. Conflicting conclusions among Russian investigators (e.g. Petrov 1966 and Ivanov 1986, Grossvald 1988 and Glushkova 1989) also hamper any easy evaluation of glacial histories in the region using the existing literature.

For these reasons, it has been necessary to find an alternative method for investigating the glacial history of Chukotka. Satellite synthetic aperture radar (SAR) imagery has proven useful for this purpose. Using SAR imagery we have compiled a map of glacial landforms of the Chukotka Peninsula, Russia. In areas where we find moraine sequences we have attempted to estimate relative age using moraine morphology and position. In order to quantify morphologic characteristics, we utilized a technique which enabled us to measure the surface slope angle of a moraine, a criterion used in the relative age dating of glacial moraines in Alaska and elsewhere (Hamilton et. al. 1986, Kaufman

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and Calkin 1988). While our age assignments are rough estimates at best, the map of glacial landforms should provide an essential geographic framework for the increasing number of US and Russian investigations aimed at interpreting the Pleistocene history of Beringia.

Methods:

Satellite Synthetic Aperture Radar

Synthetic Aperture Radar (SAR) is a high resolution, active microwave sensor that can image the earth's surface regardless of daylight or weather conditions (Curlander and McDonough 1991). SAR coverage is presently available from the First European Earth Resources Satellite (ERS-1) of the European Space Agency. Regional data from ERS-1 is received and processed into imagery at the Alaska SAR Facility at the Geophysical Institute of the University of Alaska Fairbanks.

The study of landforms in SAR imagery is, to an extent, analogous to the technique of aerial photo analysis. In both types of imagery, qualitative interpretations can be made by visual inspection. In contrast to aerial photography, however, a SAR image is recorded as digital information. SAR data can therefore be analyzed computationally in order to make determinations about the nature of the surface being imaged. Both visual inspection and digital analysis have been employed in this study.

Mapping from SAR imagery

Visual analysis of a glaciated landscape using SAR involves the identification and description of those geomorphic features within an image that appear to have been

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deposited or formed by glaciers. Arcuate, sub-linear features bounded by or located at the mouths of wide mountain valleys are interpreted as terminal moraines. Other easily identifiable glaciogenic features include kettles, lateral moraines, drumlins, and eskers. In order for a feature to be resolved it must be large enough to cross through at least two pixels which has the effect of distinguishing the feature from the surrounding area of otherwise random pixel brightness values (Olmstead 1993). Large scale features can be resolved in SAR imagery which has a pixel size of 12.5 m and a resolution of 25-30m. The dimension of a feature can be estimated by measuring the number of pixels that compose the feature and calculating a distance or an area based on the 12.5 m pixel size. For each moraine identified in this study, we included a general description of the landform, the size and location, the relative contrast with the surrounding area, the degree of subsequent stream dissection and erosion, and any other notable features that distinguish each moraine (Table 2.1).

Estimating Relative Ages of Moraines

The relative degree of geomorphic degradation has been used extensively as an estimate of relative age for glacial landforms, especially moraines (Kaufman and Hopkins 1986, Hamilton 1986, Kaufman and Calkin 1988, Kaufman 1988). The surface relief of a glacial landform is reduced or 'flattened', over time, by geomorphic agents such as gravity, water, wind, and frost. Younger features tend to express 'fresh' morphology while older one are more subdued and degraded. Although many factors may influence the morphologic character of a moraine, most workers in glacial geology find that moraine

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morphology is highly dependent on relative age (Kaufman 1988). An important criterion used in morphology-dependent relative age studies of terminal moraines is slope angle.

The digital data in a SAR image can be analyzed geometrically in order to determine the surface slope angle of landforms in the image. We utilize here a method that allows us to measure slope angles of glacial moraines by measuring the difference in average pixel brightness returned by a sloping surface versus a nearby flat surface (S. Li, pers. comm. 1994). Each pixel in an image has a numerical '*dn*' value, which represents 'brightness' or amount of energy returned to the satellite, and ranges from zero (no return) to 255 (maximum return). This value is a grey scale analog used to represent the *dB* value that expresses the actual signal return to the satellite. The *dB* value is function of density and electrical properties of the ground surface, ground surface roughness, and surface slope angle relative to the incident radar wave. Assuming that the first two parameters remain generally constant over the width of the moraine (see section on Distortions, Potential Errors and Constraints), the brightness of the moraine can be attributed to its surface slope relative to that of the surrounding terrain.

The average brightness value for a moraine slope is measured by defining a sloping target surface that has a slope normal to the radar look direction (parallel to satellite flight path), then recording the brightness values for each pixel inside the defined area of interest, and averaging the values. The same is then done for the surrounding (and presumably level) terrain. The difference in mean intensity of return between the slope and a nearby flat surface (ΔdB) is used to determine the slope angle.

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Local incidence angle (γ) is determined using ΔdB in the following equation and solving it iteratively until a best solution for (γ) is found (S.Li pers comm. 1994):

$$10\log_{10}\left(\frac{\tan^2(\theta)}{\tan^2(\gamma)}\right) = \Delta \mathrm{dB}$$

The surface slope angle is calculated by subtracting the local incidence angle (γ) from the look angle (θ) of the satellite where the latter is always approximately 23° (Figure 2.2).

Distortions, Potential Errors, and Constraints

It is important to point out that SAR images contain geometric distortions which complicate the location and measurement of features. Terrain is severely distorted due to the ERS-1 satellite look angle which is inclined approximately 23 degrees from vertical (Curlander and McDonough 1991). This has the effect of causing extreme topography to appear to 'lean' toward the satellite. Image data within the foreslopes and backslopes of steep mountains, for example, will be lost. This is not a significant problem, however, when observing glacial landforms in wide valleys with relatively flat floors. The topographic relief of moraines is too small to cause significant distortion, but is enough to make the feature easily identifiable. Another distortion is related to the velocity of the satellite and how that differs relative to different points in the image. This velocity difference effect causes the image area to be warped rather than perfectly square

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(Olmstead 1993). The effect is less pronounced than the terrain distortion and is not considered a significant source of error. All images are radiometrically calibrated to attain a uniform range of brightness values.

Additionally, the dielectric properties and the 'roughness' of the surface being imaged, may have an effect on the return of energy from that surface. Dielectric properties are affected by moisture content of soil and vegetation. We try to ameliorate this affect by choosing images taken between late fall and early spring, when the ground is most likely to be frozen and biomass (leaf cover) is at a minimum. Roughness on a small scale (< 1 meter) may affect the amount of radar backscatter. Most of the moraines in Alaska, and those observed in Russia, are covered with soil and vegetation which lessens the effect of small scale roughness. The effect of large scale slope of the target surface is much greater than the combined effects of dielectric properties and small scale roughness, and is the primary signal returned to the satellite.

In order to assess the accuracy of our slope angles estimated from the satellite data, we conducted a study to compare field-measured slope angles with SAR-measured slopes. The slope angles on moraines in the field come partly from our own measurements in Interior Alaska, and partly from published and unpublished data collected by other workers (Figure 2.3).

We also conducted an experiment to see how much variation in return might be due to vegetation type. We measured moraine slope angles on the ground at Donnelly Dome which corresponded very well with the SAR measured slope angles for the same moraine (Figure 2.3). The surface slope measured was vegetated dominantly by willow

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shrubs with several widely scattered spruce trees. In determining the slope angle from SAR imagery of this moraine, we chose three flat surfaces that had distinctly different vegetation cover. One surface was thickly vegetated with a mature, closed canopy, spruce forest (D(f), Figure 2.3). A second surface was vegetated in an 'open spruce parkland' that is, shrub tundra dotted with widely spaced, solitary spruce trees (D(p), Figure 2.3). The third surface was covered in willow shrub tundra with no spruce trees (D(s) Figure 2.3). The largest variation was seen when we used the closed spruce forest as the background flat surface. The difference in slope angle was only 2-3 degrees. The other surfaces varied only by 1-1.5 degrees. This is in contrast, however, with the rather large (6°) difference seen when using differently vegetated background surfaces in Lituya Bay in southeast Alaska. Here the target surface vegetation was broad leaf forest versus closed Sitka spruce forest (L(p) and L(s) in Figure 2.3). These tests show that surface slope angles obtained from SAR data are indeed subject to variations in vegetation type. This problem seems to be minimized, however, in arctic and subarctic regions where the diversity of plant species and community type is greatly reduced. We feel that these measurements are valid enough, when combined with other criteria, for inter-moraine comparisons and relative age estimates.

Results:

Moraines in Chukotka

From the SAR imagery, we were able to identify nearly fifty glacial moraines in Chukotka (Figure 2.4). The various types of data collected from the imagery are

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presented in Table 2.1 and the location of the moraines are shown in the following series of maps (Figs 2.4, 2.5, 2.8-2.10).

The Anadyr region (Figure 2.5) contains twenty glacial moraines differing in apparent age and extent. Nested moraines on the coastal lowland mark glaciers that terminated 200-300 km from their source area. The outermost moraines (*Tn1, Mk1, Ka1*: Figure 2.5) exhibit low slope angles, low contrast, and are highly degraded and eroded. The inner moraines on the coastal plain exhibit slightly higher surface slopes and are clearly more visible on the images (Figure 2.6) (*Tn2, Mk2, Ka2*: Figure 2.5). Similar morphology is seen in another set of paired moraines which are located in the Osinovaya Valley to the northwest (*Os1, Os2*: Figure 2.5). An intermediate set of moraines is located between 75 and 200 km from a probable source area (*Tn3, Mk3, Ka3*: Fig. 2.5). These moraines are less distinct as ridges and tend to exhibit a hummocky and kettled topography. The smallest, and apparently youngest moraines are confined to valley mouths at higher elevations (~500 m) in the Pekulney Mountain range (*Pk1-6*), and exhibit extremely fresh morphology (Figure 2.7). These moraines are located only 6-15 km from cirque headwalls. See Table 1 for detailed information on individual moraines. Linear oriented mounds on the far eastern portion of this map (Figure 2.5) are interpreted as drumlins. The presence of these features suggests an older, more extensive glaciation that may be recorded in sediments along the coast of the Gulf of Anadyr.

The Amguema region to the north of the Anadyr lowland also contains a sequence of moraines characterized by increased morphologic preservation with decreasing extent (Figure 2.8). On the northern coastal plain are located several moraines

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emanating from the Vankarem and Amguema river valleys (*Am1, Am2, Va1, Va2*: Fig. 2.8). These moraines are highly degraded, have low slope angles, and extend 200-300 km from the mountains. In several places along the Vankarem and Velmay rivers, a drift limit of intermediate extent is visible (*Va3, Vl*: Fig. 2.8) Upstream in the Amguema valley are several very distinct moraines at the mouths of the Vulvyveyem and Irvynetyveym valleys (*Vu1, Vu2, Ir*: Fig. 2.8). These moraines are double crested, exhibit high slope angles and appear fresh. These younger moraines extend 50-60 km from valley headwalls (Table 1).

The coastal lowlands surrounding Kresta Bay are marked by the presence of several moraines. These appear to be laterals emanating from the fjord valleys now presently forming Egvekinot and Kengynin Bays at the top of Kresta Bay (Figure 2.9). Two outermost features may be composite moraines or possibly marine terraces (*Ma, Vk*: Fig. 2.9). A more distinct group of moraines parallel the shore of Kresta Bay. These ridges have intermediate slope angles (10-13°) and limited subsequent dissection and degradation (*Kg1, Kg2, Kg3*: Fig. 2.9). The moraines on the coastal plain mark glaciers that extended 50-75 km from their source. One small moraine, possibly a lateral, was observed at the mouth of the Egvekinot valley (*Eg*).

In the Mechigmen region, the moraines seen on SAR imagery are scattered and do not form sequences as well as in the other regions of Chukotka. Two features along the shores of Kaliuchin and Ioniveyem Bays may be remnants of glacial moraines, but a definitive identification from SAR is impossible (*Et, Kh1*:Figure 2.10). The ridges are somewhat arcuate and strongly degraded. Displacing the Kalhuerveem River is a series of

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nested arcuate moraines that all appear to be of similar age (*Kh2*: Fig. 2.9). These moraines are intermediate in extent (75 km) and have slope angles of approximately 13°. To the east, Getleyanen Lake is dammed by a moraine that forms the outer coast, and another moraine crosses the bay itself (*Gt1, Gt2, Gt3*: Fig. 2.9). These moraines extend 75 km from the valleys to the south west. Inland from Mechigmen Bay there are several places where, although moraine crests are absent, the limit of drift is clearly marked by kame and kettle topography (*Mg1, Mg2*: Fig 2.9). Several small cirques in the Tenyani range just south of Lavrenty Bay contain small terminal or lateral moraines (*Lv1, Lv2, Lv3*: Fig 2.9). The outer shore of Lavrenty Bay, on the south side, may be mantled by a lateral moraine (*Lv1*:Fig. 2.9)

Age Classification

Our results indicate that glacial moraines and other geomorphic landforms can be identified, and that certain morphologic characteristics can be observed and measured from SAR images. Using these characteristics (relative position, appearance, slope angle) we are able to classify the Russian moraines into three major categories (Figure 2.11): ***1) extensive oldest glaciation*** characterized by glaciers that formed distinct terminal moraines 200-300 km from their source area, are highly dissected by subsequent stream erosion, and exhibit degraded morphology and low surface slope angles (<7°), ***2.) intermediate glaciation(s)*** that advanced 50-200 km from source areas, sometimes exhibiting several moraine crests or recessional positions, sometimes lacking distinct crests (appearing only as drift limit), are often kettled, and have intermediate surface slopes (7-

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13°), and ***3.) youngest limited glaciation*** limited to valley mouths in mountainous (>500 m altitude) regions, typically < 20 km from cirque headwalls (except in upper Amguema region where they extended 50-60 km), exhibiting "fresh" morphology and high surface slopes (>14°). This classification is summarized in Table 2.2.

Discussion:

By applying our knowledge of the relationship between moraine morphologic characteristics and relative age in similar environments in Alaska (Kaufman and Calkin 1988, Hamilton et. al. 1986) we can make rough age estimates for the moraines in Chukotka. Comparison suggests that the 'fresh', steep (≥15°), proximal moraines at Pekulney, Vulvyveyem, Irvynetyveyem, and Tenyanni are correlative with the Mt. Osborne moraines on the Seward Peninsula and the Itkillik II moraines in the Brooks Range. These Alaskan moraines are considered to be Late Wisconsin in age (Kaufman and Calkin 1988, Hamilton et. al. 1986, Hamilton 1994, Kaufman and Hopkins 1986). Similarly, the low slope angles and large extent of the outermost moraines in the Anadyr and Amguema regions are similar in morphology and extent to the Nome River and Sagavinirktok moraines in the Kigluaik mountains and Brooks Range, respectively. These moraines are assigned to the Middle Pleistocene by Kaufman and Calkin (1988), Kaufman et. al. (1991), Kaufman and Hopkins 1986, and Hamilton (1986). Stratigraphy on St. Lawrence Island records the advance of a glacier, that originated in the mountains of Chukotka and eventually encroached onto the island. This advance is dated with amino acid ratios obtained from mollusk shells and is of Middle Pleistocene age (Benson et. al. 1994, Heiser et. al. 1992) Although there are no well preserved moraines on St. Lawrence

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Island, this glacial advance is comparable in extent to the one marked by the outer moraines in the Anadyr region, and is probably correlative.

The intermediate moraines are more difficult to compare with moraines on the Alaskan side of Bering Strait. Several advances are recorded in Alaska between the Middle Pleistocene and the Late Wisconsin. The most distinctive of these in Alaska are the Salmon Lake moraines of the Kigluaik Mountains (Kaufman and Calkin 1988, Kaufman and Hopkins 1986) and moraines of Itkillik I age in the Brooks range (Hamilton 1986), both of which are presumed to be Early Wisconsin in age. Stratigraphy on St.Lawrence Island also records a second glacial advance that was nearly as extensive as the Middle Pleistocene glaciation. This event occurred sometime after the Last Interglacial period, 125,000 years BP (Heiser et al. 1992, Benson 1994). There is no evidence of any, non-local, Late Wisconsin glaciation on St. Lawrence. The age assignments of Russian workers agree, in general, with our estimates. In the Anadyr lowland, for example, Glushkova (1989) has compiled a map of moraines based on satellite imagery and aerial photography. Her map corresponds very closely with the moraines mapped in this study using SAR. The smaller moraines in the Pekul'ney mountains are considered to be Sartan (Late Wisconsin) age (Glushkova 1994, Glushkova et al 1989), and the inner moraines in the Anadyr lowland are considered Zyryan (Early Wisconsin) by Glushkova (1994). Glushkova does not assign older ages to any moraines north of the Anadyr River; all non-Sartan moraines are considered to be terminal or recession moraines of Zyryan age.

Somewhere in or near the Osinovaya or Belaya River valley (Figure 2.5), Svitoch

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(1967) describes a section that contains a gravel unit that may be glacio-fluvial and is dated by thermoluminescence to 366 $\pm$42 thousand years old. Ice rich-silts higher in the section date to 31,000 $\pm$850 ka (MGU-311). Both of these units contain cold climate shrub-tundra pollen, and are separated by a unit that contains a warmer pollen spectra (Svitoch 1967). This stratigraphy and geochronology suggest that Sartan (Late Wisconsin) aged glaciers never reached the lower valleys of Chukotka.

Petrov (1966) describes a section somewhere "on the right shore of the lower Vankarem River (Figure 2.8) that contains a unit of glacial till between alluvial gravels. He assigns the glacial till to the Vankarem suite of sediments which are considered to be of Zyryankan or Early Wisconsin age. This assignment agrees with our age estimate based on the moraine morphology and position. Again, there is no evidence that Sartan aged glaciers extended out of the mountains.

In the Tenyanni mountains, Sartan moraines are mapped and described by Sedov 1992. These glaciers were extremely limited in extent and indicate snowline lowering of only ~150 m for late Wisconsin glaciers. The extent and elevation of these moraines is consistent with what is known about Late Wisconsin age glaciers in Alaska (Kaufman and Hopkins 1986).

The model of extensive ice sheet growth in Beringia during the Late Wisconsin (Grossvald 1988, Hughes and Hughes 1994, Grossvald and Hughes 1995) is thus not supported by the moraines observed on the Chukotkan landscape. If *all* the moraines described had been deposited during the recession of extensive Late Wisconsin age ice sheets as suggested by Hughes, (pers comm. 1994) they should all have very similar

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morphologic characteristics (steep slope, high contrast, fresh appearance, etc.). Instead, the various sets of moraines observed in SAR images exhibit distinct differences in degree of weathering and degradation that correspond well with position (fresher upstream, subdued downstream).

Conclusions:

Satellite SAR imagery shows great promise as a reconnaissance tool for glacial mapping and moraine morphological investigations in regions of the world that have little or no aerial photo or large-scale map coverage. Study of this imagery is especially useful during the preliminary stage of any research effort in these areas, as it provides otherwise unavailable visual and digital information that can aid preliminary morphologic characterizations and age estimates.

Our work, so far, suggests that at least two, and probably three, advances of alpine glaciers and mountain ice caps in Chukotka are correlative with glacial advances recognized in Alaska. A glacial event of limited extent, characterized by 'fresh' moraine morphologies with steep slope angles, is comparable in appearance and morphology to Late Wisconsin aged moraines in Alaska. A more extensive glacial limit, marked by low-angle degraded moraines, is probably correlative with the middle Pleistocene age moraines described from the Seward Peninsula and the Brooks Range. An intermediate series of moraines probably marks one or more separate advances that may be correlative with the Salmon Lake glaciation on the Seward Peninsula and the post Last Interglacial glacial advance that encroached on St. Lawrence Island. The sequences of moraines in Chukotka are similar, in morphology and position, to moraine sequences described in Alaska, and

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therefore evidence a sequence of glacial events that most likely began in the middle Pleistocene and ended with the Late Wisconsin. The large ice sheet proposed by Grossvald (1988) and Hughes and Hughes (1994) is not consistent with the glacial moraines mapped in this study.

Reconnaissance mapping using remote sensing is an important step in developing a geographic framework for the timing and extent of Pleistocene glaciation on the Russian side of Bering Strait. This work is intended as a guide to future investigators and must be followed by extensive field investigations of the stratigraphy and geochronology of the glacial landforms described here.

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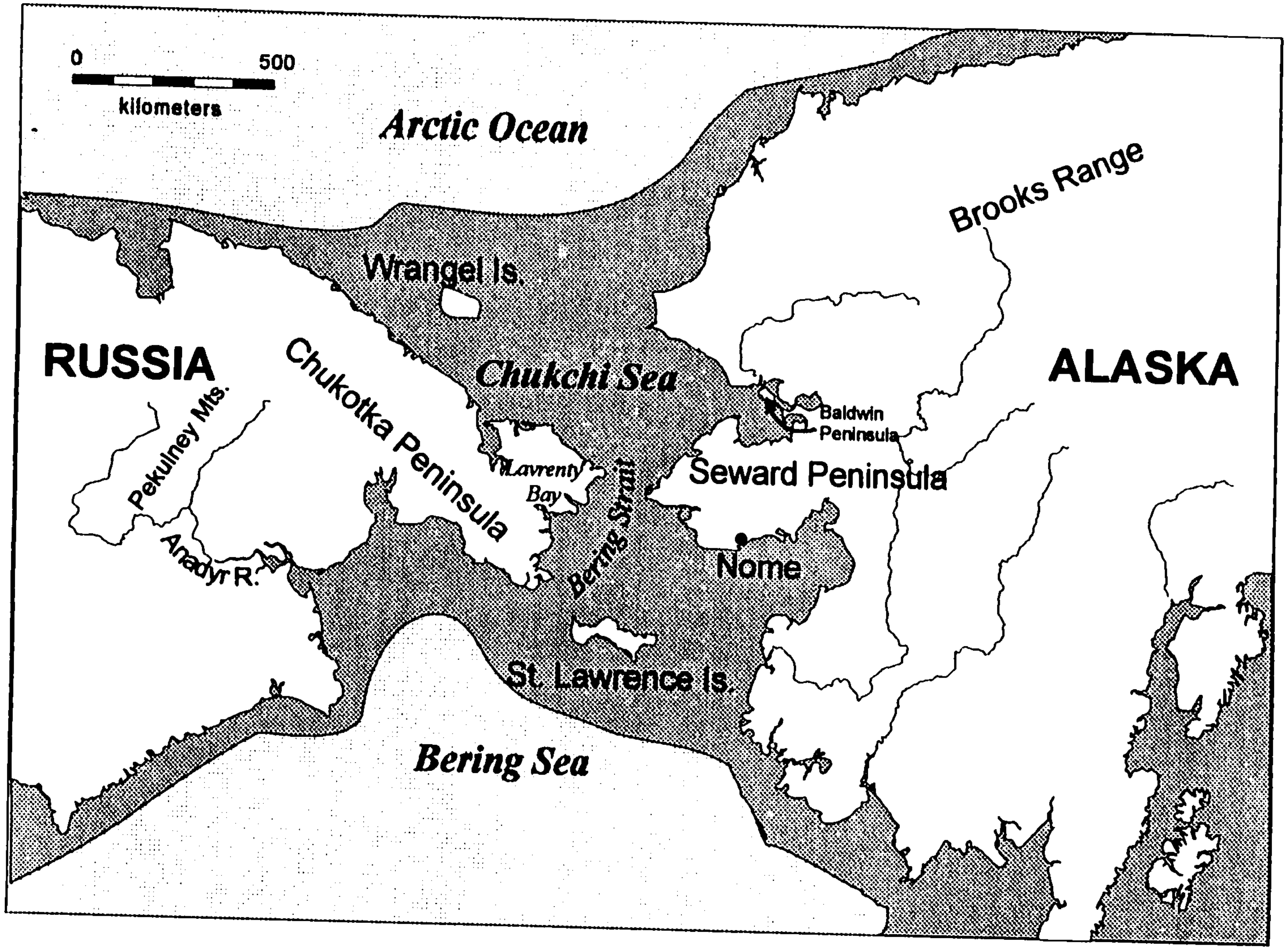

Figure 2.1 Map of Beringia showing locations mentioned in the text. Extent of the Late Wisconsin land connection is shown in dark grey.

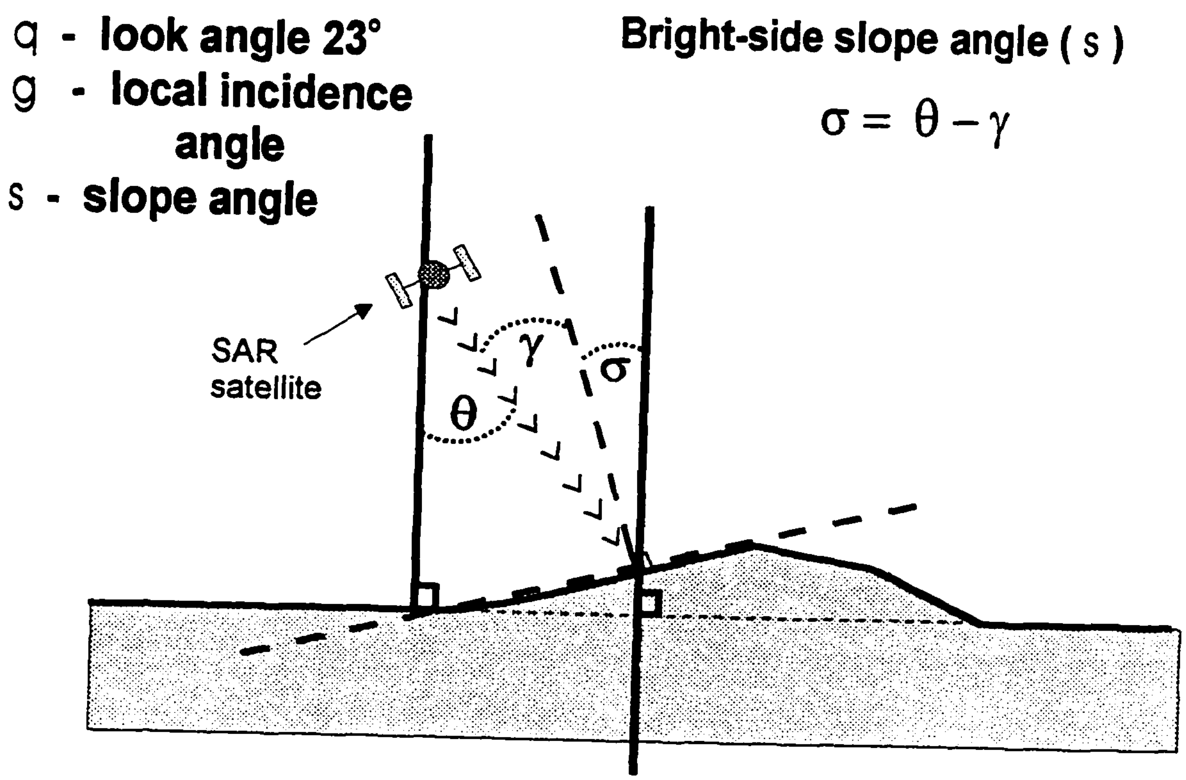

Figure 2.2. Geometric relationship between target surface and ERS-1 satellite.

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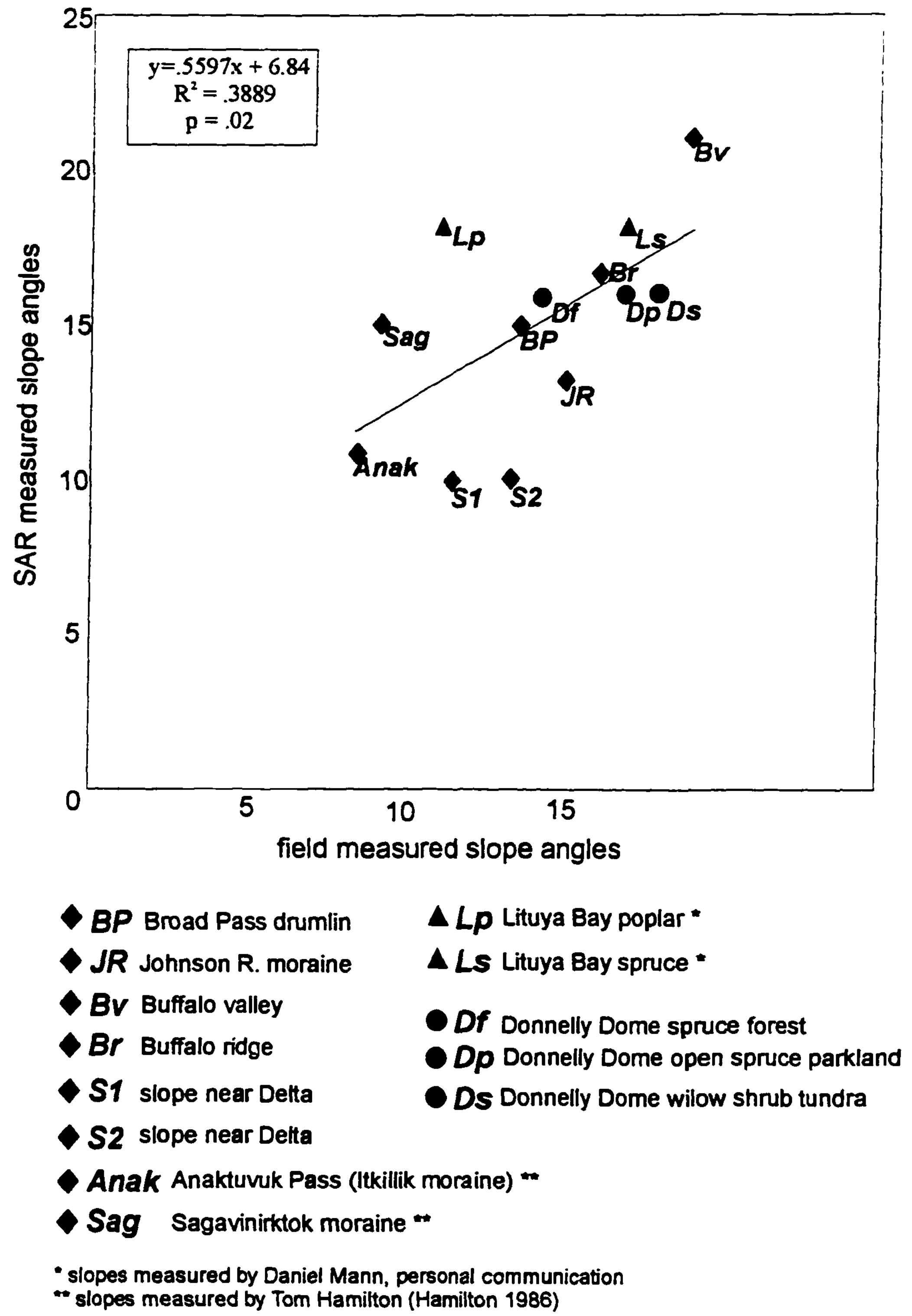

Figure 2.3. Relationship between surface slope angles measured from SAR imagery and slope angles measured from the field.

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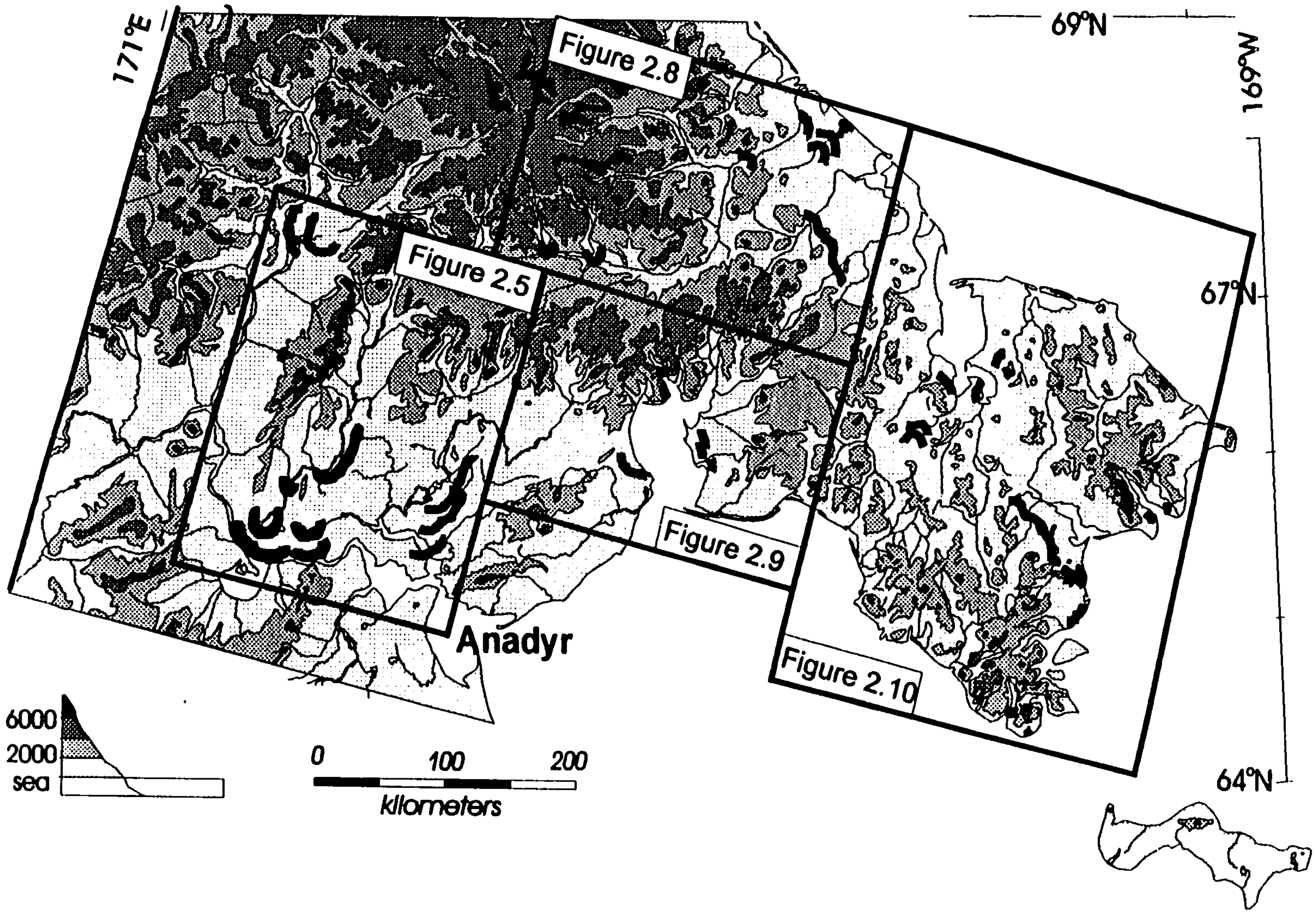

Figure 2.4. Glacial moraines of Chukotka Peninsula mapped from satellite SAR. Boxed areas refer to maps shown in greater detail in other figures.

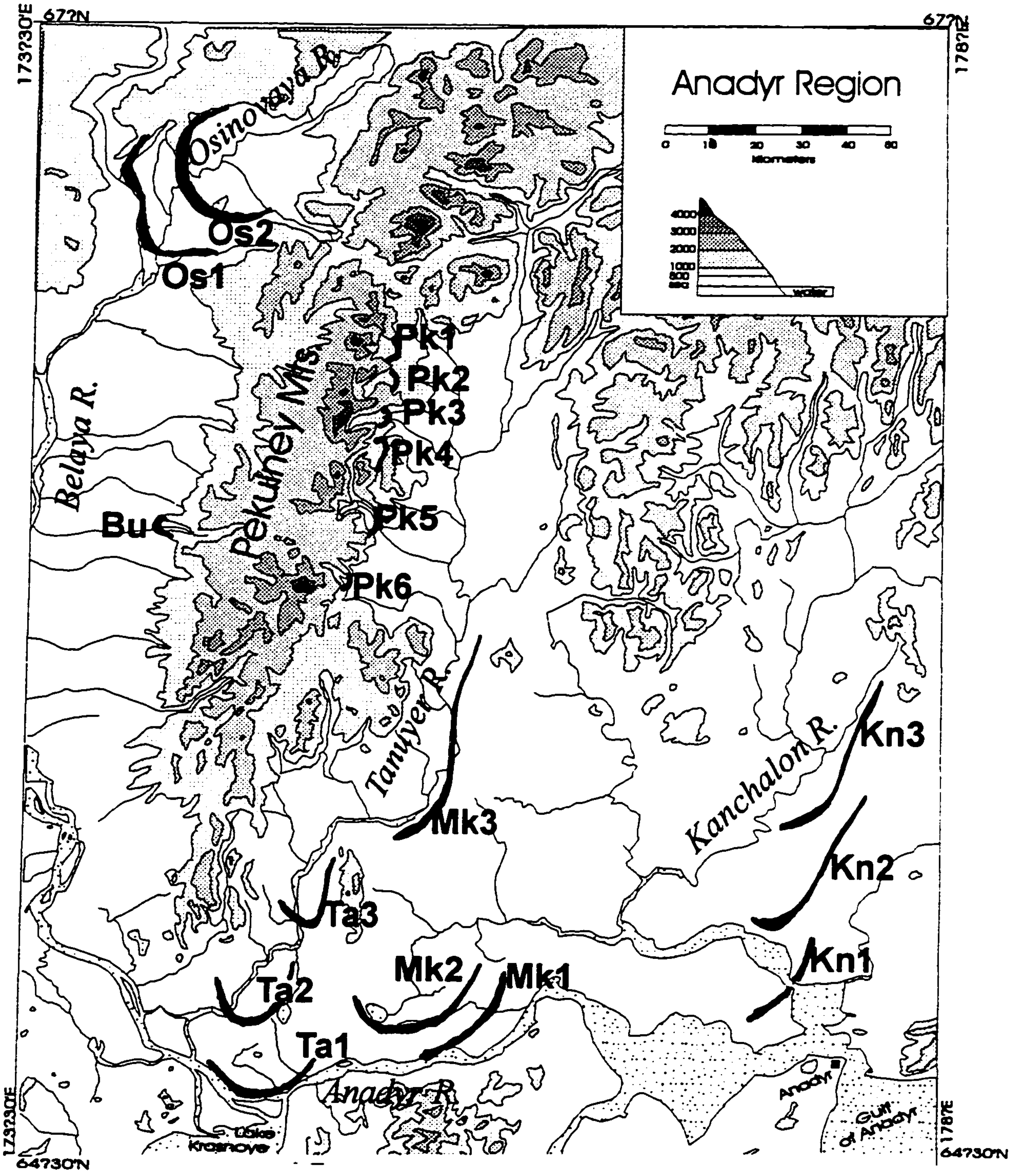

Figure 2.5 Glacial moraines observed in the Anadyr region.

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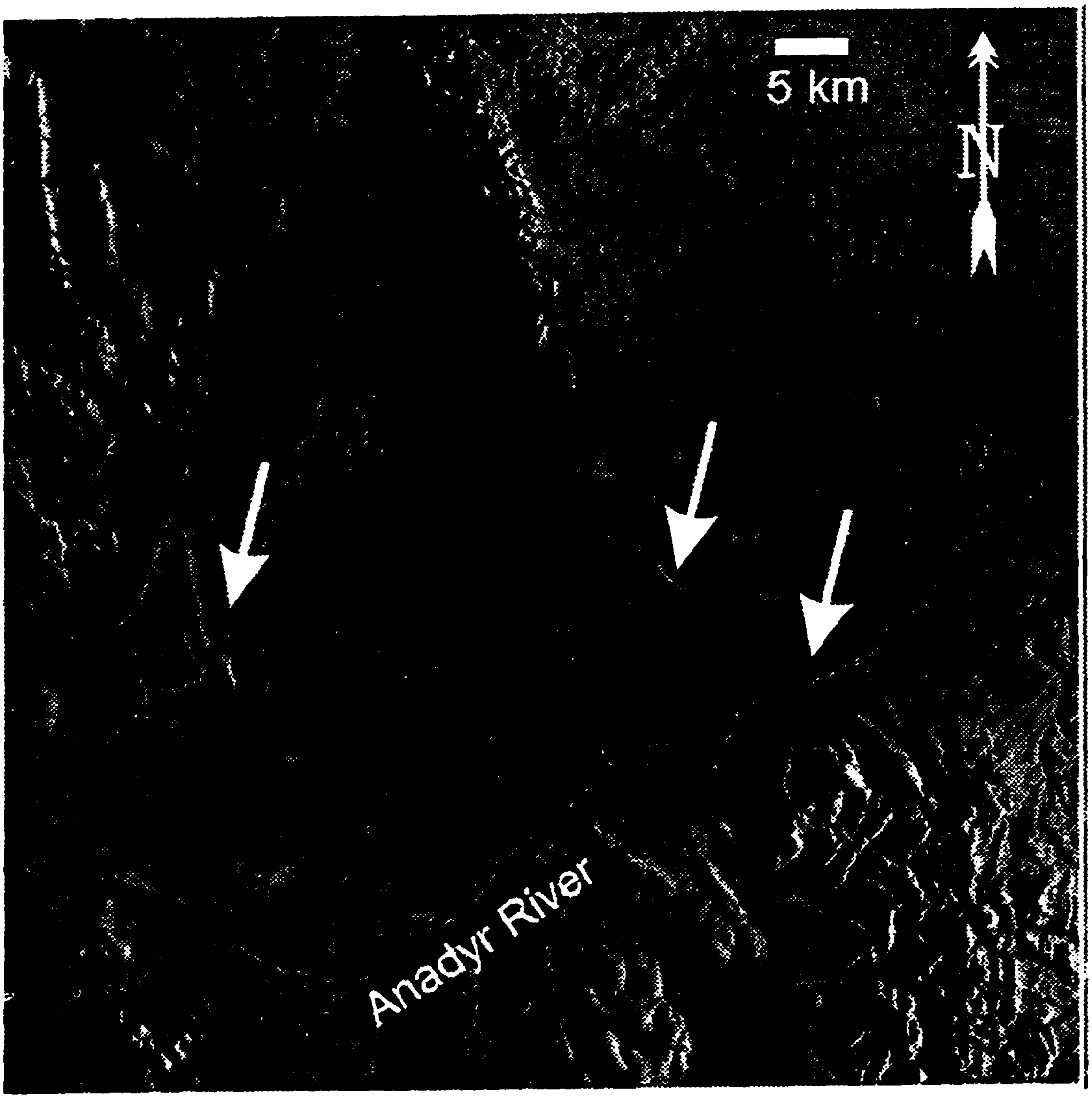

Figure 2.6. SAR images of Tanuyer and Melkoy moraines, just north of the Anadyr River.

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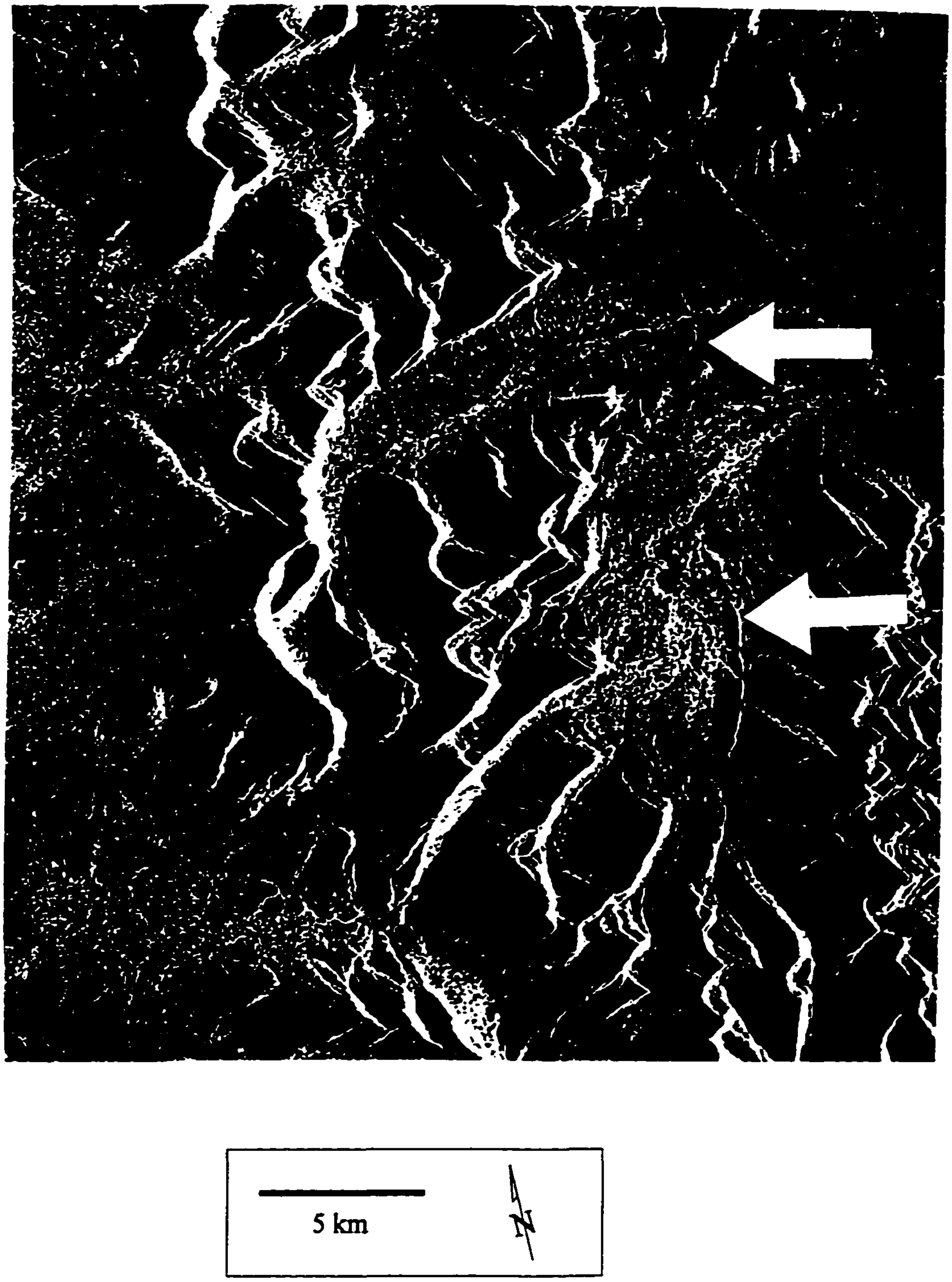

Figure 2.7. SAR image of Pekul'ney moraines Pk3 and Pk4.

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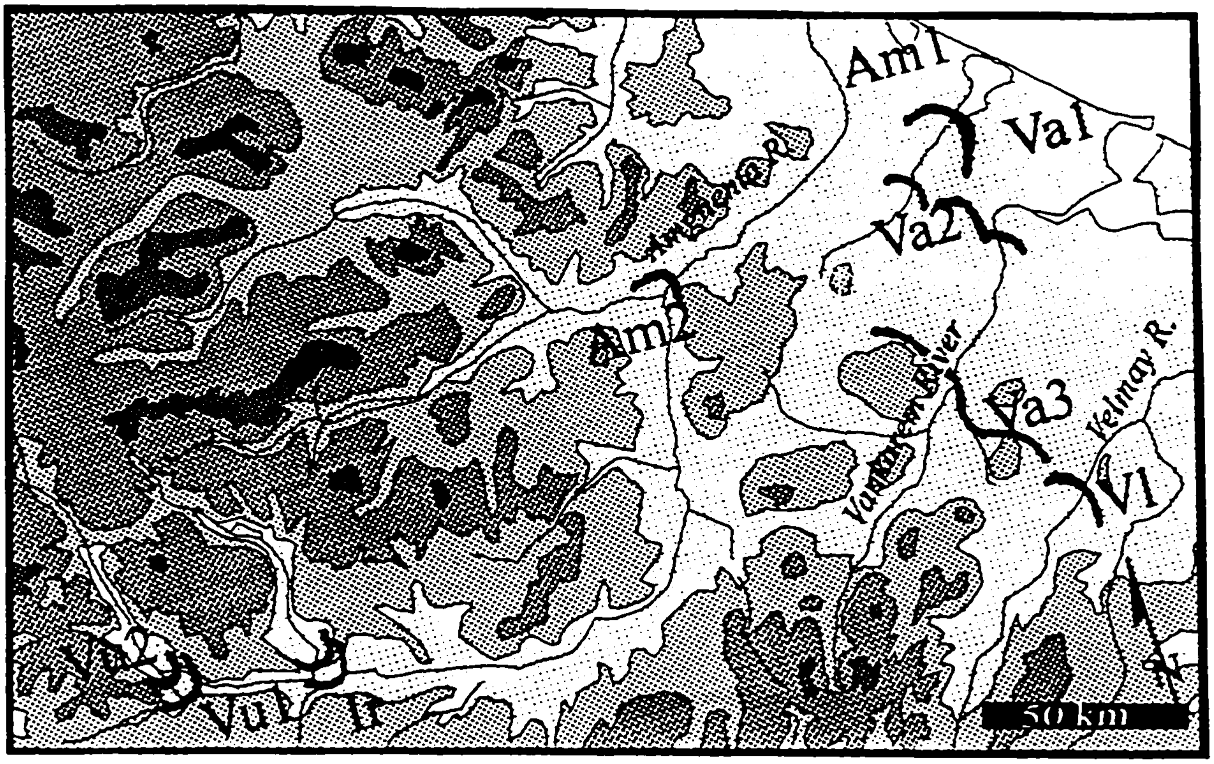

Figure 2.8. Glacial moraines observed in the Amguema region.

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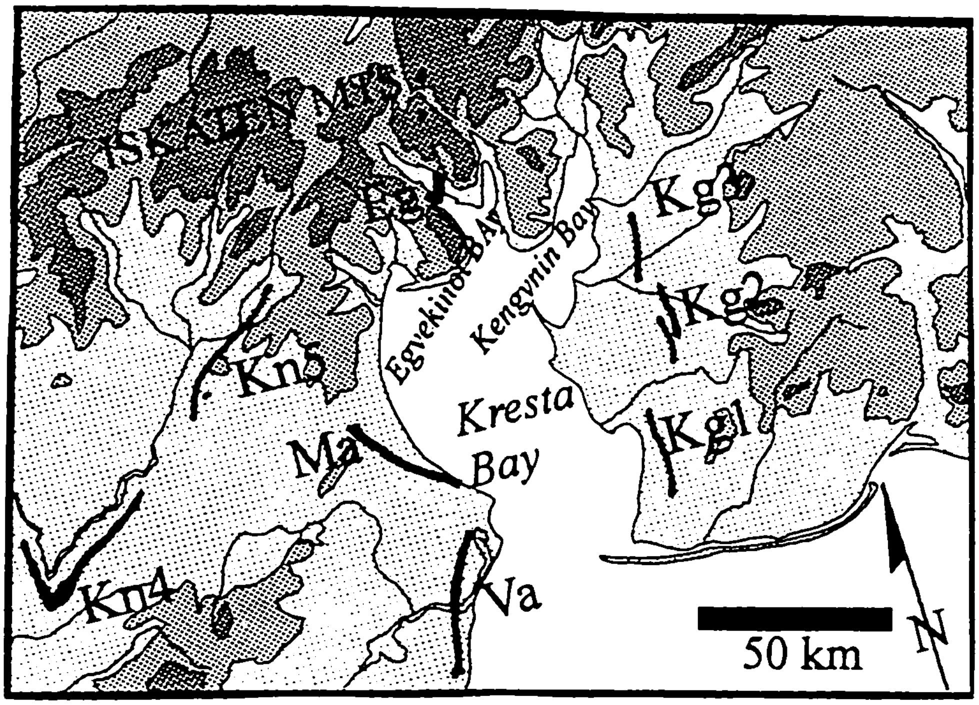

Figure 2.9. Glacial moraines observed in the Kresta region.

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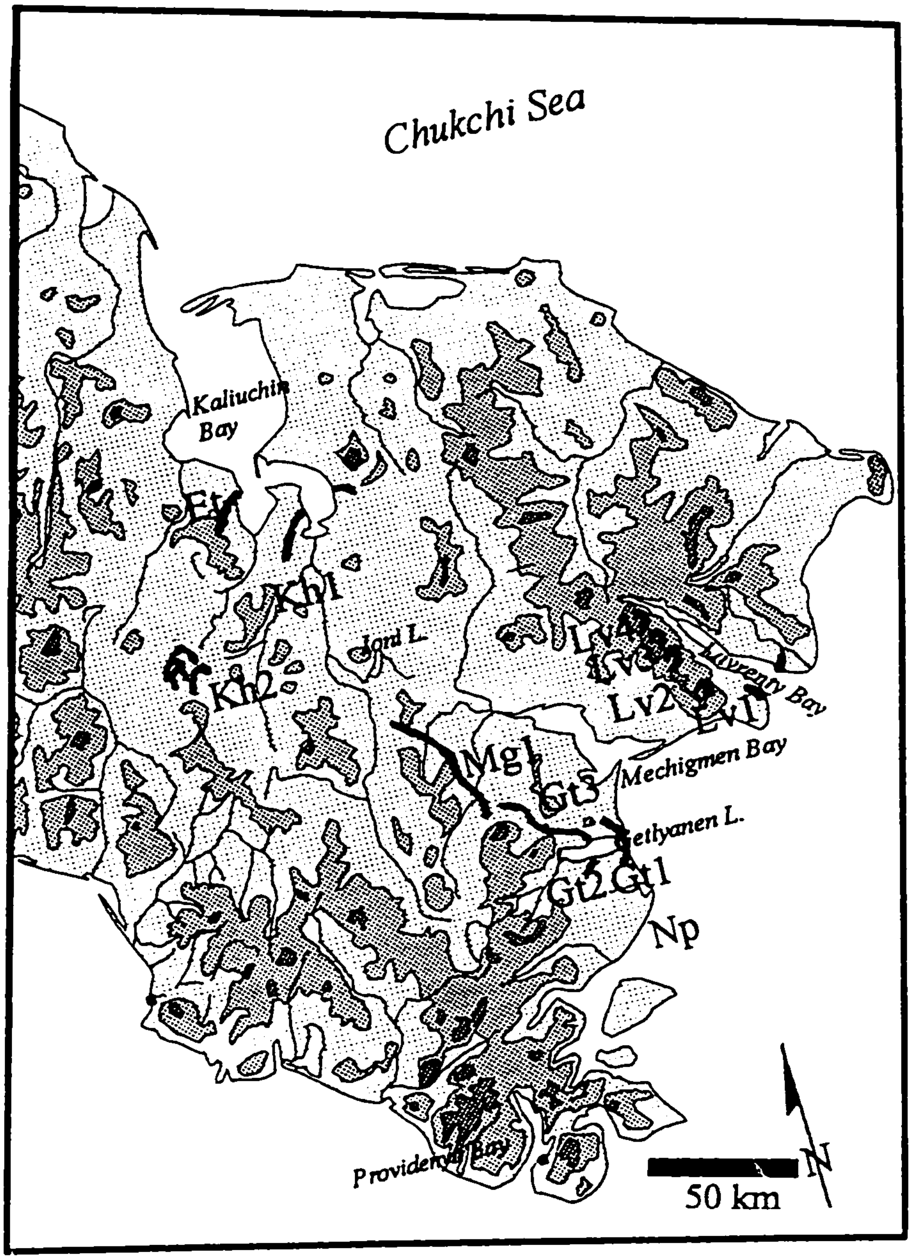

Figure 2.10. Glacial moraines and drift limit mapped in the Mechigmen region.

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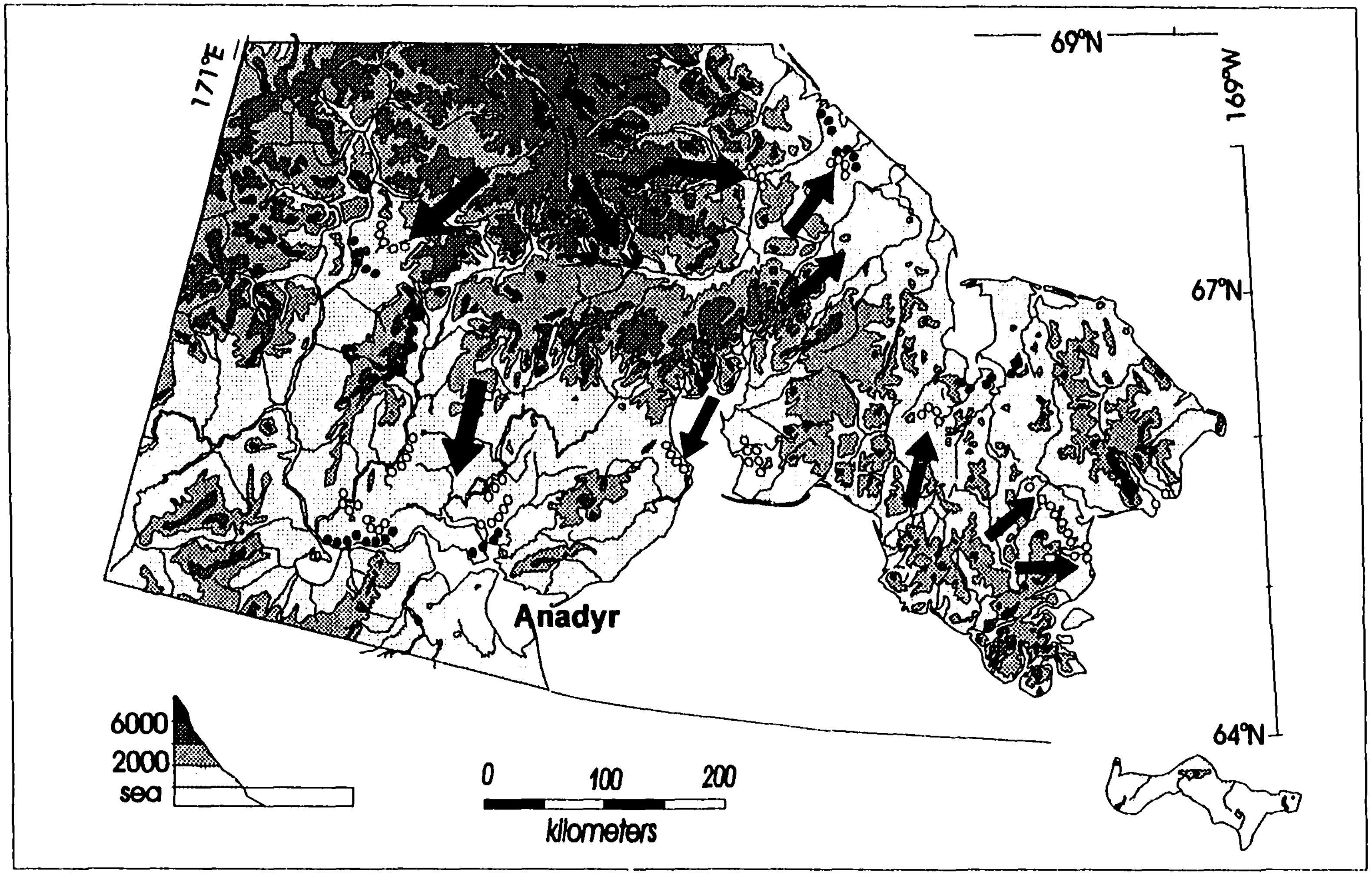

Figure 2.11. Glacial moraines on the Chukotka Peninsula mapped from SAR imagery. Black arrows show radial flow out of mountians. Estimated relative age of moraines indicated by symbols as follows: Youngest Glaciation ( ), Penultimate Glaciation (oooooo) and Oldest Glaciation (••••••). See Table 2.2 for summary of criteria used to estimate relative age.

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Table 2.1. Data collected from moraines using satellite SAR imagery.

Moraine name	**location** (lat / long, geography, map #)	**appearance on SAR imagery**	**size**
Anadyr Region	***(Figure 2.5)***		
Osinovaya 1 (Os1)	*173°55'E, 66°48'N displaces Yurumkuveem R.*	*discontinuous arc low contrast*	*~30 km long 4-7 km wide*
Osinovaya 2 (Os2)	*174°15'E, 66°40'N 15 km upstream from Os1*	*mostly continuous arc moderate contrast*	*~40 km long 3-5 km wide*
Buchia (Bu)	*174°5'E, 65°58'N, west side of Pekulney range*	*continuous arc moderate contrast*	*12 km long 2 km wide*
Pekul'ney 1 (Pk1)	*175°15'E, 65°20'N east side of Pekul'ney Mountains*	*continuous arc very high contrast*	*2 km long*
Pekul'ney 2 (Pk2)	*175°12'E, 66°17'N, east side of Pekul'ney Mountains*	*discontinuous arc high contrast*	*3 km long*
Pekul'ney 3 (Pk3)	*175°12'E, 66°12'N, east side of Pekulney Mountains*	*continuous arc very high contrast*	*~3.5km long*
Pekul'ney 4 (Pk4)	*175°11'E, 65°58'N, Tret'ya Creek east side Pekulney Mts*	*continuous arc very high contrast*	*7 km long*

dissection and stream development	distance from source	slope (°)	other features
strong, 1st & 2nd order stream development	*150-200*	-	
some consequent drainage on flanks	*150-200*	12	*crosses Osinovaya R.*
minimal dissection and stream development	*~35*	-	
minimal dissection and stream development	*6-8*		
minimal dissection and stream development	*8-10*		*smaller moraine crest visible ~2 km upstream*
minimal dissection and stream development	*8-10*	18	*smaller moraine crest visible ~2 km upstream*
minimal dissection and stream development	*6-8*	18	*smaller moraine crest visible ~1 km upstream*

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Pekul'ney 5 (Pk5)	175°12'E, 66°08'N, east side of Pekulney Mountains	mostly continuous arc very high contrast
Pekul'ney 6 (Pk6)	175°12'E, 66°08'N, Pervaya Creek east side of Pekulney Mountains	continuous arc very high contrast
Tanuyer 1 (Tn1)	174°30'E, 64°42'N, terminates on floodplain of Anadyr River	discontinuous arc very low contrast
Tanuyer 2 (Tn2)	174°35'E, 64°56'N, cut by Tanyuer east of Peschanoe Lk.	continuous arc mod. - high contrast
Tanuyer 3 (Tn3)	174°50'E, 65°05'N, cut by Tanyuer R., east side of "Dosmunaya Mt"	mostly continuous arc mod. - high contrast
Melkoy 1 (Mk1)	175°35'E, 64°40'N, terminates at Anadyr R.	mostly continuous arc moderate contrast
Melkoy 2 (Mk2)	175°20E', 64°50'N, dams Melkoy Lake	continuous arm mod.- high contrast

4 km long	minimal dissection and stream development	12-15	17	smaller moraine crest visible ~1.5 km upstream
2 km long	minimal dissection and stream development	10-12		
20 km long 4 km wide	strong, 1st order stream development	250 -300		moraine rises to ~17 m above surrounding surface
30 km 1 km	dissected by Tanuyer, some consequent streams	~225	10	moraine rises to ~36m above surrounding surface
18 km long 2-3 km wide	dissected by Tanyuer R., limited consequent streams	~200		double crested moraine abundant kame and kettle 25 m above surroundings
25 km 2-3 km wide	strong, 1st and 2nd order stream development	250-300	<5	only ½ remains uncut by Anadyr, 37 m above surface
34 km	moderate, some consequent stream development	240-280	9	exactly parallels Melkoy 1, rises 43 m above surrounding surface

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Melkoy 3 (Mk3)	*175°20'E, 65°09'N, east side of "Dosmunia" Mt*	*wide, discontinuous ridge, moderate- high contrast*
Kanchalon 1 (Kn1)	*177°05'E,64°55'N, cut by Kanchalon Estuary*	*wide, discontinuous arc low contrast*
Kanchalon 2 (Kn2)	*177°05'E,65°07'N, terminates 10 km north of estuary*	*wide arc, no visible crest moderate-low contrast*
Kanchalon 3 (Kn3)	*177°20'E,65°15'N, 20 km upstream of Ka2*	*discontinuous arc low contrast*
Kanchalon 4 (Kn4)	*178°15'E,65°39'N, diverts upper Kanchalon River*	*v. wide, discontinuous arc, low contrast*
Kanchalon 5 (kn5)	*179°05'E,65°49 N, 40 km south of mountain front*	*mostly continuous ridge mod.-high contrast*
Amguema Region	***(Figure 2.8)***	
Vulvyveyem 1	*178°40'E, 66°56'N, crosses*	*continuous arc*

50-70 km 10-12 km wide	*moderate, some consequent stream development*	*~200*		*double crested moraine abundant kame and kettle 32 m above surroundings*
30-40 long 8 km wide	*strong dissection and stream development*	*150-200*		*highly degraded and kettled, dams Kanchalon estuary*
65 km long 3-6 km wide	*strong dissection, 1^{st} & 2^{nd} order stream development*	*150-200*	*6*	*heavily kettled, long lateral on east side is less kettled*
50 km long 2-5 km wide	*moderate - strong dissection some drainage*	*120-160*		*parallel to Kn2, less distinct at arc than on lateral*
50-60 lm 8-10 km	*moderate-strong dissection some drainage development*	*75-100*		*no visible crest, kame and kettle topography*
20 km long 2-4 km wide	*some dissection and draoinage development*	*50-75*	*5.5*	*well preserved crest, could be lateral to Kn4 (?)*
15 km long	*minimal dissection and*	*50-60*	*16*	*outer moraine,*

(Vu)	upper Amguema R.	mod. -high contrast
Vulvyveyem 2 (Vu2)	178°40'E, 68°56'N, crosses upper Amguema R.	continuous arc high contrast
Irvynetyveyem (Ir)	179°35'E, 68°58'N, north side of upper Amguema R.	continuous, kettled arc moderate contrast
Amguema 1 (Am1)	177°27'W, 67°55'N, eroded by braided Amguema River	discontinuous ½ arc low contrast
Amguema 2 (Am2)	178°25'W, 67°44'N, associated with large meander in valley	small ridge, moderate contrast,
Vankarem 1 (Va1)	176°35'W, 67°53'N, terminates at Nutauge Lagoon	mostly continuous ½ arc, low contrast
Vankarem 2 (Va2)	176°50'W, 67°52'N, west side of Vankarem River	discontinuous, wide arc moderate - low contrast
Vankarem 3 (Va3)	177°13'W, 67°15'N, mouths of Yaranaykool and Mayel'ma R.	mostly continuous arc moderate - high contrast
Velmay (Vl)	178°45'W, 67°05'N, crossed Velmay River	non-continuos ridge, moderate contrast

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1.5 wide	stream development			slightly wider and 'smoother'
12 km long ~1km wide	minimal dissection and stream development	50-60	17	inner moraine, appears slightly 'rougher'
40 km long 5 km wide	moderate dissection little stream development	50-60	17	wide moraine with visible double crest, kettled
~15 km long 3 km wide	strong, consequent stream development,	200-300	7	non-distinct and heavily kettled
3 km long 1 km wide	limited dissection	100-200		may be river terrace ??
17 km long ~2 km wide	strong dissection and degradation	200-300	7	not very distinct, parallels inner moraine 8 km away
40 km long 5 km wide	strong, 1st and 2nd order stream development	200-300	9	quite diffuse, numerous kettles or thaw lakes (?)
35 km long 2-4 km wide	moderate dissection, some consequent drainage	100-200	8	good crests on east side, diffuse and kettled on west
~25 km long 3-4 km wide	dissected with limited drainage	100-200		no apparent crest, marks limit of drift, rough & kettled

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Kresta Region	***(Figure 2.9)***	
Egvekinot (Eg)	*179°10'W,66°26'N, 7-8 km upstream from Egvekinot Bay*	*mostly continuous arc high contrast*
Kengynin 1 (Kg1)	*178°20'W,66°08'N, parallels eastern shore of Kengynin Bay*	*mostly continuous ridge moderate - high contrast*
Kengynin 2 (Kg2)	*178°22'W,66°58'N, parallels shore of bay, 10 km east of Kg1)*	*mostly continuous ridge moderate - high contrast*
Kengynin 3 (Kg3)	*178°28'W,65°51'N, 5 km east of Kg2*	*two sections of ridge moderate contrast*
Mamchergyrgyn (Ma)	*179°55'W,65°50'N, cut by western shore of Krest Bay*	*wide discontinuous ridge low - moderate contrast*
Vel'kally (Va)	*179°37'W,65°38'N, ~10 km inland, west side Kresta Bay*	*wide ridge or terrace low contrast*

4-5 km long	*limited dissection and stream development*	*10-12*		*hard to distinguish in steep valley, may have double crest*
17 km long	*limited dissection and stream development*	*50-75*	*13*	*lateral moraine from Egvekinot Bay (?)*
14 km long	*limited dissection and stream development*	*50-75*	*13*	*lateral moraine from Egvekinot Bay (?)*
5 km each	*limited dissection and stream development*	*50-75*	*11*	*lateral moraine from Egvekinot Bay (?)*
~20 km long 4-5 km wide	*some dissection and stream development*	*100*		*may be composite lateral from upper Kresta Bays*
18 km long 3-4 km wide	*strongly dissected, 1^{st} & 2^{nd} order stream development*	*100*		*may be marine terrace (?)*

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Mechigmen Region	***(Figure 2.10)***	
Etureveyem (Et)	*174°35'W,66°32'N, parallel to shore of Kaliuchin Bay*	*wide arc moderate - low contrast*
Kalhuerveem 1 (Kh1)	*174°W,66°23'N, forms west shore of lioniveyem Bay*	*low contrast, discontinuious arc*
Kalhuerveem 2 (Kh2)	*175°05'W,66°05'N*	*several continuous arcs moderate - high contrast*
Getleyanen 1 (Gt1)	*172°11'W,65°15'N, dams bay at outer coast*	*distinct, mostly continuous arc, low contrast*
Getleyanen 2 (Gt2)	*172°20'W,65°12'N, 5 km inland from coast, cut by bay*	*mostly continuous arc low contrast*
Getleyanen 3 (Gt3)	*172°32'W,65°18'N, westward continuation of Gt2*	*mostly continuous arc moderate- low contrast*
Lavrenty Bay 1 (Lv1)	*171°W,65°35'N, southwest side of mouth of Lavrenty Bay*	*continuous ridge moderate contrast*
Lavrenty Bay 2 (Lv2)	*171°20'W,65°42'N, 1 km from south shore Lavrenty Bay, in cirque named "Vulcanaya"*	*short ridge high contrast*

15 km long 2-3 km wide	*strongly dissected, 1st & 2nd order stream development*	*100-150*		*non-distinct, may be bedrock (?)*
	highly dissected	*?*		
5-8 km each	*moderate dissection and stream development*	*75*	*11*	*series of nested, arcuate moraine crests, some kettles*
10-15 km 2-3 km wide	*strongly dissected, 1st & 2nd order stream development*	*80*		*forms eastern shore of end of bay, cut by coast*
15 km 2-3 km wide	*strongly dissected, 1st & 2nd order stream development*	*75*		*center of morainal arc is covered by Getleyanen Bay*
12 km 2-3 km wide	*strongly dissected, 1st & 2nd order stream development*	*75*		*loops around hill north of Getleyanen Bay*
12 km long 2 km wide	*strongly dissected, 1st & 2nd order stream development*	*50+*		*probably lateral moraine terminating off shore*
1-2 km long	*limited dissection and stream development*	*8-10*		*small terminal moraine at mouth of "fresh" cirque*

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Lavrenty Bay 3 (Lv3)	*171°28'W,65°15'N, in small valley west of Vulcanaya*	*ridge along east valley wall, high contrast*	*3 km*	*limited dissection and stream development*	*4-5*	*lateral moraine in "fresh" cirque*
Lavrenty Bay 4 (Lv4)	*171°30'W,65°17'N, next small valley west of Lv 3*	*ridge along eastern valley wall, high contrast*	*2-3 km*	*limited dissection and stream development*	*4-5*	*lateral moraine in "fresh" cirque*
Mechigmem 1 (Mg1)	*172°54'W,65°32'N, diverts Iyeniveyem River*	*appears as drift limit moderate contrast*	*35 km long 10+ km wide*	*some dissection and drainage,*	*40 -50*	*kame field, no morainal crest visible, drift limit*
Mechigmen 2 (Mg2)	*173°45'W,65°49'N, 5 km south of Ioni lake*	*appears as drift limit moderate contrast*	*20 km long 15+ km wide*	*some dissection and drainage,*	*50-60*	*kame field, no morainal crest visible, drift limit*
Nychigen Point (Np)	*172°07'W,65°04'N, forms outer coast of Senyavina St.*	*wide ridge along shore moderate contrast*	*20 km long 2 km wide*	*strongly dissected, 1st & 2nd order stream development*	*70 +*	*probably lateral moraine from Senyavina Strait*

Table 2.2. Summary of criteria used to estimate relative age of moraines.

	Extent (km from cirque)	Surface Slope	Morphology
Youngest Glaciation	12-60	> 16°	very high contrast, high relief exhibit 'fresh' or 'rough' surface morphology, sometimes have laterals, sometimes double crested
Penultimate Glaciation	50-200	7-13°	moderate contrast and relief sometimes lacking distinct and continuous crests, often kettled, moderate surface relief (roughness), some have several 'crests' or recessional positions
Oldest Glaciation	200-300	<7°	low relief, low contrast highly degraded and dissected by subsequent stream erosion

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CHAPTER 3

LATE WISCONSIN GLACIAL EXTENT IN BERINGIA: ADDRESSING THE CONTROVERSY

Abstract

The extent of glaciers in Beringia during the Late Wisconsin (23,000-13,000 BP) is controversial. The current paradigm calls for limited mountain-glacier complexes in Alaska and northeastern Russia that were considerably more restricted in extent than glaciers in other regions of the northern hemisphere. The limited extent of glaciers left the Bering Shelf exposed and available as an important route for the dispersal of ice age mammals and humans between Asia and North America (Figure 3.1). This limited model is generally accepted by most workers in Beringia and is supported by numerous field investigations on both sides of Bering Strait. An alternative model of extensive late Wisconsin glaciation in Beringia, based on ice sheet modeling and selected field data, features an ice cap over Chukotka that was confluent with an ice sheet over the Arctic Ocean.

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prepared for submission to Quaternary Research

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Glacial mapping from satellite Synthetic Aperture Radar (SAR) and other remote sensing data, support minimum Late Wisconsin glaciation in Beringia. Imagery of the Chukotkan landscape shows moraine sequences that record numerous glacial advances that have decreased in maximum extent since the Middle Pleistocene. Age estimates are based on moraine position, on the relative geomorphic preservation of moraines in Chukotka compared with dated moraines in Alaska, and on age determinations made by Russian researchers. The innermost moraines in Chukotka, presumably Late Wisconsin in age, are limited to mountain valleys and comparable in extent and geomorphic preservation with the limited glacial advances of that age in Alaska. All moraines on the northern coastal plain were formed by glaciers flowing *north* from the mountains. There is no evidence of glaciers encroaching southward from the postulated marine based ice domes. The lack of isostatically uplifted, post-Late Wisconsin, shorelines in the Bering Strait region provides additional evidence against the presence of a large ice sheet over Chukotka during the last ice age. Snowline reconstructions for the Late Wisconsin in northeast Siberia rise from the southeast to the northwest, and do not support the contention that snowline may have been lowered by 1000 meters during that time (Grossvald and Hughes 1995). Paleoclimatic data from around Bering Strait, including radiocarbon dates on floral and faunal remains, preclude the existence of an ice sheet in Beringia during full glacial time.

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I. INTRODUCTION

The extent of glaciers in the Bering Strait region during the Last Glacial Maximum (LGM) has become controversial. The established paradigm calls for limited mountain-glacier complexes in Alaska and northeastern Russian that were considerably more restricted in extent than glaciers at slightly lower latitudes (40° - 60° N) in the northern hemisphere (Coulter *et al.*, 1965; Hopkins, 1982; Hamilton, 1994). Challenging this view is a model suggesting that a massive ice sheet centered over Chukotka and its northern continental shelf in the Chukchi Sea (Figure 3.2) enveloped most of the north-central part of Beringia, including the Bering Strait, during the LGM around 19,000 years ago (Hughes and Hughes, 1994; Grossvald and Hughes, 1995). Resolution of this controversy is important because of the potential impacts that glacier extent could have on the functioning of the Bering Land Bridge as a corridor for biogeographical exchange between Asia and North America at the end of the last ice age. A massive ice sheet extant during the LGM would have restricted greatly the times when the land bridge was open for crossings. Isostatic adjustments associated with such an ice sheet would have led to a sea-level history much different from the one now inferred for the interval 20 ka to 0 BP (Elias *et al.*, 1993; Mann and Hamilton, 1995). The extent and timing of Beringian glaciations are also important inputs for global climate models (Kutzbach *et al.*, 1993) and in oceanographic models of climate change (Shaffer and Bendtsen, 1994).

The restriction of Late Wisconsin glaciation to local mountain massifs is supported by most field studies in central and northern Alaska and northeastern Russia (Hopkins,

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1972; Hamilton et al 1986; Glushkova, 1994; Ivanov, 1986; Arkipov, 1986). As discussed in detail below, considerable paleoecological data from Beringia also supports the limited glaciation model. Areally restricted glaciation in northern Alaska and Chukotka during the LGM is thought to be an indirect result of lowered sea level and colder global climate. Eustatic sea-level fall during Isotope Stage 2 (ca. 25-10 ka BP) exposed the shallow continental shelf of Bering Sea and isolated Beringia from moisture originating in the North Pacific Ocean and the remnant Bering Sea (Hopkins 1972, Mann and Hamilton 1995). Distant moisture sources and cold ice-age temperatures restricted the growth of glaciers on both sides of Bering Strait and left the land bridge open as a potential route for the intercontinental dispersal of plants and animals (including humans) between Asia and North America.

The idea of a Bering Strait ice sheet appears to have originated with the American naturalist John Muir, who visited both sides of the Bering Strait aboard the *Corwin* in 1881. Muir described glacial landforms and sediments in the Bering Strait region and interpreted them as evidence for a large ice sheet advancing from the north (Muir, 1917). More recently, M.G. Grossvald, a geographer at the Russian Institute of Geography in Moscow, proposed that the continental shelf of the Arctic Ocean fringing northern Siberia supported marine-based ice sheets during the Late Wisconsin (Grossvald 1988). His ideas were inspired by his acquaintance with the ice shelves of East Antarctica, by the marine based ice domes over West Antarctic, and by a marine-based ice sheet known to occupy the Barents Sea during the Late Wisconsin (Boulton and Rhodes, 1974). An earlier

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version of Grossvald's vision of previous glaciation in Beringia included ice domes along the southern edge of the Bering Sea continental shelf with fringing, Antarctic-style ice shelves extending over much of the shrunken remnant of the Bering Sea (Grossvald and Vozovik, 1983).

In 1994, T. and B. Hughes presented the Marine Ice Transgression Hypothesis (MITH), which asserts that marine-based ice sheets developed on the continental shelves of the Arctic Ocean during Pleistocene glacial maxima and proposes mechanisms for their growth and expansion. According to MITH, marine ice sheets form when perennial sea ice thickens through bottom freezing and eventually becomes grounded on the sea floor. Meteoric accumulation and run-off from rivers dammed along the landward margins of the growing ice mass contribute to the growth of the ice domes. Eventually the ice domes grow enough to intersect the perennial snow line and advance landward as ice sheets and seaward as floating ice shelves (Hughes and Hughes, 1994; Hughes, 1994). Invoking this dramatic formative mechanism, Grossvald and T. Hughes hypothesized the presence of several distinct ice sheets in the Russian Arctic and the Bering Strait region. For the LGM, they propose the existence of an ice sheet that reached 1500 meters thick over Wrangel Island and flowed southward over Chukotka and through Bering Strait as far as St. Lawrence Island, as well as northward as a thick ice shelf over the Arctic Ocean (Figure 3.2).

Here we test the alternative hypotheses concerning the extent of glaciers in Beringia during the LGM using remote sensing data, geophysical models of lithospheric

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response to an ice load, regional snowline reconstructions, and field data from Bering Strait and Chukotka. Topographic maps and remote sensing data are difficult to obtain from Russian sources. Consequently, we use satellite imagery to map glacial extent, to estimate the relative ages of glacial landforms, and to determine paleo-ice flow directions. We use geophysical modeling to estimate the surface displace (including rebound) that would have been induced if the ice sheet predicted by Hughes and Grosswald did indeed exist during the LGM. Finally, we use regional paleoclimatic indicators including snowline depression to determine whether the environmental conditions required to support the MITH ice sheet were actually present during the LGM.

II. METHODS

A. LOCATING AND ESTIMATING AGE OF MORAINES IN CHUKOTKA

We used false color, infrared LANDSAT images (image enlarged to approximately 1:250,000) and a number of grey-scale synthetic aperture radar (SAR) images (scale 1:500,000) to map glacier extent and to evaluate geomorphic parameters related to the relative age of moraines in northeastern. These same techniques were applied to moraines in Alaska where other remote sensing data (false color infrared, stereo U2 photography, scale 1:60,000) were available and/or where detailed ground-based studies had been conducted by other workers (see Chapter 2). The geomorphic parameters we used to judge the relative ages of moraines were slope angle, the relative degree of moraine dissection through erosion by stream and thermokarst, and the planform

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supraposition of different moraines. These parameters have been widely used by American workers to evaluate the relative ages of moraines on the Alaskan side of Bering Strait (Kaufman and Hopkins, 1986; Hamilton, 1986; Kaufman and Calkin, 1988; Kaufman, 1988; Mann and Peteet, 1994).

B. MODELING ICE SHEETS, SOLID-EARTH DEFORMATION, AND RELATIVE SEA LEVEL CHANGE

We modified a numerical model developed by J. A. Clark and colleagues to study glacial isostasy in the Great lakes region (Clark *et al.*, 1990; Clark, 1994). If the history of surface loading, and structure and viscosity of the earth is known, then displacement of the lithosphere in response to a given load can be determined (Clark, 1990). We used Greens functions, (which were kindly supplied by J. A. Clark), calculated for a spherical, self-gravitating, viscoelastic, E1 Earth model (Clark, 1990; Clark, 1994) to model the deformation of the solid surface of the Earth relative to the geoid for a load history that varied spatially and temporally. Time step Greens functions, when convoluted with a surface point load, describe the solid earth response for a given time at any given distance from that load. Greens functions include terms for the elasticity of the lithosphere, lithospheric thickness, flexural rigidity, and mantle viscosity (Peltier, 1974; Lingle and Clark, 1985). Clark *et al.* (1994) show that an E-1 Guttenberg-Bullen Earth model with a 112 km-thick elastic lithosphere overlying a uniform mantle with a viscosity of 10^{21} Pa and with an inviscid core, most accurately represents the deformation seen around the Great Lakes region. We have used the same E-1 Earth model here because it is likely the

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closest approximation of lithospheric thickness in the Bering Strait region. In actuality, the lithosphere is probably thinner in the Bering Strait than in the center of the north American continent, and our model results probably underestimate the amount of deformation that may be experienced under a given load in this region.

In order to numerically model the solid earth response to an ice sheet, the size and thickness of the load must be known and represented in a way that allows easy division into elemental loads. We use a series of discreet disc loads of varying thickness to represent the ice sheet proposed by Grossvald and Hughes (1995). Disc loads are not the optimum way to represent an ice sheet because the gaps between them result in an underestimation of the ice load (Clark 1994). However, since disc loads are easier to handle numerically, and since we are seeking a minimum estimate of isostatic depression rather than a maximum, underestimating ice load does not compromise this test of the MITH model.

Each disc was divided into discreet elemental loads of ~10 km^2 (in plan view) and one meter thick. Over the surface of the Earth an elemental load of this size can be considered a point load. The mass of each element was determined by multiplying the volume of each elemental load by the density of ice. The total point load was then determined by multiplying the mass of the element by the ice sheet thickness (number of elements stacked over point). The problem is linear so the solution is obtained through summation of the responses, at a given distance and time, to the elemental loads which make up a given disc. Disc loads of varying thickness were then arranged to represent the ice sheet postulated by Grossvald and Hughes (Grossval and Hughes 1995, Figure 5).

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The total solid-earth response at any given location and was determined by summing the response, at that location, to all of the disc loads present in Beringia at that time and prior to that time. The locations we chose to predict the solid-earth response are all at present sea level, which serves as a reference point for past sea-level changes.

The load history is important because the viscoelastic earth 'remembers' past load changes. Ice sheets are dynamic and it is unlikely that any ice sheet remains in its maximum configuration long enough to achieve isostatic equilibrium (Clark, 1994). Since the ice sheet we are modelling grew and decayed within a finite interval of time, we represent the load history as a simple three stage process (Figure 3.3). Stage I represents the nucleus of the ice sheet, which according to the MITH, began to accumulate on the continental shelf of the Chukchi Sea (Hughes and Hughes 1994, Grossvald and Hughes 1995). For Stage I, we assumed growth to be approximately half of the maximum ice thickness. Stage II is intermediate in extent and thickness between Stages I and III. Stage III is the ice sheet at its maximum and is based on ice-surface contours provided by Grossvald and Hughes (1995) (Figure 3.3).

There is no firm constraint on the timing of ice-sheet initiation, time of maximum extent, or deglacial history for this hypothetical ice sheet. The creators of the MITH suggest it took 20,000 years to grow the ice sheet to its maximum position. Consequently, we sought limiting cases of ice-growth history by choosing four different load histories to represent possible glacial advance and retreat scenarios (Figure 3.4). The timing of initiation, duration of maximum position, and timing and rate of retreat in all four scenarios was based roughly on glacial chronologies from surrounding areas of the North

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Pacific (Mann and Peteet, 1994; Mann and Hamilton, 1995) and in the Verkhoyansk Mountains of Siberia (Kind 1967). In devising load histories (Figure 3.4), we chose a minimum model characterized by a catastrophic retreat at 15 ka BP (LH 1), a maximum model with rapid deglaciation at 10 ka BP (LH 3), an intermediate model with rapid retreat (LH 2), and an intermediate model with symmetrical (with regard to time) growth and decay (LH 4).

The solid-earth response obtained for each ice-load history includes terms for geoid perturbation; that is, the distortion of the gravitational equipotential surface corresponding to mean sea level caused by changing ice and water loads. The resulting curves represent changes through time in the separation between the earth surface (ocean bottom) and the geoid (ocean surface). This change in separation, occurring at a particular location during a particular time interval, is defined as the change in relative sea level (RSL). The predicted relative sea-level curves were adjusted so that present sea level is zero. The rise in eustatic sea level at the end of the last ice age (Fairbanks, 1989) was then subtracted from the modeled RSL curve in order to predict changes in sea level, relative to today, for the last 20,000 years. These results tell us the magnitude and direction (rising or falling) of sea-level changes following the deglaciation of the hypothesized MITH ice sheet

C. RECONSTRUCTING LATE WISCONSIN SNOWLINES IN BERINGIA

We reconstructed the configurations of ancient valley glaciers in Chukotka using moraines that were previously assigned an LGM age according to the geomorphic criteria

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discussed in Chapter 2. The best maps of Chukotka that are available to us are 1:500,000 Tactical Pilotage Charts (TPC) with a contour interval of 1500 feet (150 m). We used several methods to estimate average paleo-equilibrium line altitudes (ELAs) including cirque floor altitude, median glacial altitude, and toe-to-headwall ratio (THAR) (methods discussed in detail in Mierding, 1982, Waythomas, 1990). In this study we obtained a maximum and minumum estimate of the headwall altitude and the toe (terminal moraine) altitude (Table 3.2). These 'max' and 'min' altitudes were then converted to meters and used in the THAR method (with a THAR value of 0.6; see Meirding, 1982) and median glacial altitude (MEG) method. In both methods and the maximum and minimum value were averaged. The approximate averaged value of the THAR was used to create the snowline map in Figure 3.7.

III. RESULTS

A. LOCATION, INFERRED FLOW DIRECTIONS, AND RELATIVE AGE OF PLEISTOCENE GLACIERS IN CHUKOTKA

Numerous moraine systems are visible between the Belaya River and the eastern coast of Chukotka on SAR imagery (Figure 3.5). These moraines record the repeated termination of glaciers in the same general area. Some of these moraine sets differ markedly in geomorphic preservation, suggesting significant age differences. Along the northern side of the lower Anadyr River, for example, three moraine systems are distinguishable on the basis of slope angles, relative position, and degree of dissection by

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stream erosion. Comparisons with Alaska moraine sequences suggest these groups of moraines could span 10^5 years.

SAR-derived relative age criteria for Russian moraines, when compared with moraines in Alaskan that have been assigned an LGM age by previous workers, suggest that Chukotka glaciers were restricted to mountain valleys during the LGM. In Alaska, LGM moraines preserve sharp crests, exhibit little alteration by stream erosion or thermokarst, and retain steep slope angles ranging from 15-20° (see Chapter 2). Similarly characterized moraines in Chukotka were deposited by glaciers which originated in mountain ranges with altitudes above 1200-1500 meters.

The position and orientation of the concave side of terminal moraines, relative to mountain valleys and upland source areas, indicate a general radial flow pattern in Chukotka and give no indication of predominate north-south flow (Figure 3.5). For instance, in the Pegtymelskiya Mountains north of Anadyr, glaciers flowed south into the Anadyr basin, southeast into Kresta Bay, northeast down the Anguema valley, and southwest down the Belaya valley. Moraine relative ages suggest that this flow pattern was repeated during several successive glacier advances. Similarly, moraines record multiple advances flowing north out of the Iskaten Mountains in southern Chukotka (Figure 3.5). Other notable end moraines recording one or two episodes of northward advance towards the Chukchi Sea coast occur in the Vankarem, Velmay, and Anguema valleys. There is no evidence of glaciers flowing southwards and onshore onto the northern coast of Chukotka.

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B. EXPECTED CRUSTAL RESPONSES TO THE *MITH* ICE SHEET

The ice sheet hypothesized by T. Hughes and M. Grosswald (Hughes and Hughes, 1994, Grossvald and Hughes, 1995) would have caused sizable lithospheric adjustments in the region, our geophysical modeling predicts relative sea-level histories during the last 10,000 years in different parts of the region (Figure 3.6, locations shown in figure 3.1) that differ quite distinctly from geologic records of relative sea level during the same interval of time. Our results show that Bering Strait coastlines should have experienced RSL changes analogous to those experienced in the regions under, and adjacent to, the Laurentide and Fennoscandian ice sheets during the late Pleistocene and Holocene, (i.e. vertical movements on a scale of tens of meters). The predicted direction and magnitude of displacement varies markedly with location (Table 3.1). We present the model results for the late Holocene to assure complete flooding of the Bering and Chukchi Seas' continental shelves and to assure the presence of ocean waters capable of forming shoreline features. Near the center of the hypothesized MITH ice sheet, relative sea level is predicted to have been dropped 15-20 meters in the last 5000 years. In contrast, at Cape Espenberg (Figure 3.1) near the margin of the ice sheet, sea level is predicted to have risen continuously 6 to 13 m over the last 5000 years (Table 3.1) as a consequence of a collapsing forebulge effect (Clark 1990).

Relative sea level responses to the crustal loading under the MITH ice sheet if it existed, should be observable today along the coasts of the Bering and Chukchi Seas. In particular, raised marine terraces are predicted by our model on Wrangel Island and along the northern coast of Chukotka (e.g., Uelen) that are similar to those formed in

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Spitsbergen after 13 ka BP (Salvigsen 1981, Foreman *et al.* 1987) and in response to the collapse of the marine-based Barents Sea ice sheet between 13 and 11 ka BP (Foreman *et al.* 1995). However no raised shorelines are detectable on SAR images of the north and south coasts of Wrangel Island, nor have they been observed in the field (P. Anderson, personal communication). Uelen, a village on the north coast of Chukotka, is located on a spit that enclosed a shallow lagoon. Low hills along the coast would be mantled with raised beaches if crustal rebound (i.e. falling RSL) had occurred, but, no trace of raised shorelines exist there.

Rising relative sea level causes barrier beaches to transgress over shorelines and drowns river mouths. All our load histories predict rising RSL at Cape Espenberg (Figure 3.6; Table 3.1). Extensive radiocarbon dating of the foredune complex and its associated archaeological sites shows, however, that there has been no significant change in sea level at Cape Espenberg for the last 4500 years (Mason and Jordan, 1994). The only evidence for RSL higher than today on the northern Seward is the presence of ancient barrier islands that formed at a sea level of 7-10 meters higher than today (Pelukian Transgression) and date to the last interglacial (Brigham-Grette and Hopkins, 1995). On the Russian side, Mechigmen Bay is enclosed by an aggrading spit backed by a beach ridge plain similar to that at Cape Espenberg. Ridge altitudes vary little and do not suggest any recent relative sea level change.

The altitudes of shoreline deposits along the Alaska coast dating to Stage 5 and older interglacials also do not support the existence of a MITH ice sheet with its consequent solid-earth displacements and RSL changes in the Bering Strait region. The

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altitude of the Pelukian Transgression (Last Interglacial) is concordant throughout northwestern Alaska, typically standing 7-8 m above modern sea level (Brigham-Grette and Hopkins, 1995). The Anvillian Transgression left sediments and shoreline features at approximately 20 m above present sea level during the middle Pleistocene. The preservation of these older high stand features precludes the possibility of relative sea level rise greater than 8-10 meters along the western side of the Bering Strait during the last 125,000 years, and > 20 meters during the last 500,000 years.

C. EQUILIBRIUM LINE ALTITUDES (ELA) DURING THE LGM

A compilation of ELA estimates for the LGM in Alaska shows a general rise towards the east and north away from the Bering Sea (Figure 3.7). These estimates derive from toe-to headwall ratios, accumulation-ablation area ratios, median glacier altitude, and cirque floor altitude. In general, the LGM snowline surface parallels the modern one documented by Péwé (1975) and Detterman (1986). Snowline depression (from present) during the LGM ranged from a maximum of 650 meters along the Gulf of Alaska coast in southwestern Alaska (Mann and Peteet, 1994) to as little as 200 meters on the Seward Peninsula in northwest Alaska (Kaufman and Hopkins, 1986).

A compilation of data describing ELAs on valley glaciers of probable LGM age in Chukotka suggests a mirror image of the Alaskan side of the Bering Strait with snowline rising steadily towards the northwest (Figure 3.7). In Russia, our methods of estimating former ELA included the THAR and median glacier altitude as well as cirque-floor altitude. We estimated ELAs in several glacial valleys previously studied by Glushkova

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(1994) and obtained values 100-200 m lower than hers. One of her methods, "Gephers formula" appears similar to THAR but uses the altitude of the drainage divide instead of headwall altitude, which may account for her higher ELAs. While our values are consistently lower, they agree with Glushkova's reconstruction in that they show a distinct regional rise to the northeast. (Figure 3.7).

IV. DISCUSSION

A. GEOLOGICAL EVIDENCE AGAINST THE *MITH* ICE SHEET

Unpublished observations by Grossvald "make clear that Wrangel Island is characterized by strikingly glacial geomorphology" (Grossvald and Hughes, 1995). This glacial geomorphology is described as Lappland-type geomorphology which results in the preservation of geomorphic features under the frozen bed of an advancing ice sheet (Kleman 1994). While this phenomena seems to have actually occurred in Sweden, it was also accompanied by evidence of the glacier that actually over-ran, yet preserved, the features beneath it (Kleman 1994). The glacial geomorphology on Wrangel Island would necessarily have to be created by glaciers that left no trace. Researchers who have actually been to the island report that the northern lowlands are blanketed with marine sediments, the plains are covered in fluvial gravel and silts and there is no evidence of extensive glaciation on the island (Vertanyan et al 1993). Again, satellite imagery shows little evidence of recently glaciated valleys, or any other glacial features such as moraines, drumlins, or lineations. The lowlands are dotted with thaw lakes and are characterized by several generations of large alluvial fans. In fact they have a distinctly similar appearance

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on SAR imagery to the unglaciated coastal plain of the Seward Peninsula.

LANDSAT and SAR imagery failed to reveal moraines or ice-marginal fluvial deposits on the Chukotka coastal plain. With its > 1000 m thickness, the MITH ice sheet would have been warm based north of Chukotka, making the formation of drumlin fields and eskers possible on the coastal plain. No evidence of these features was found. Similarly, the large through valleys hypothesized by MITH to have directed ice flow southward across Chukotka do not exist. While it is true that there are a number of north-south trending valleys in Chukotka, it is also true that the regional geology has a north-south trend and many, if not most, of these valleys are structurally controlled. Examination of satellite imagery shows that most of these valleys are actually v-shaped and narrow. Only several of them, (at the mouth of which we can also see moraines recording northward flowing glaciers), exhibit the characteristic broad, U-shape of glaciated valleys.

Ice sheets flowing across Chukotka would surely have overprinted and probably erased evidence of pre-LGM glaciation. In contrast, relative age estimates for moraine sequences in Chukotka indicate the region has been glaciated on multiple occasions during the Pleistocene, and the predominant ice-flow pattern has been radial from local mountain massifs (Figure 3.5). Even the oldest moraines (outermost and most eroded) which evidence more extensive glaciation in the form of piedmont lobes in coastal lowlands (e.g., Anadyr basin and Kresta Bay) show a similar radial flow pattern. Radial ice flow on multiple occasions during the Pleistocene is not consistent with the presence of a MITH ice sheet during the LGM.

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SAR imagery suggests that the youngest and best preserved moraines are restricted to mountain valleys. Comparisons of Alaskan moraines of known LGM age with Chukotkan deposits suggests that these youngest moraines in the mountain valleys also date to the LGM. This conclusion is consistent with the field-based work of our Russian colleagues (Glushkova 1989, Glushkova 1990, Ivanov 1986, Sedov 1992), and recent studies by American workers (Gualtieri pers. comm. 1997). The pattern of more extensive, older glaciers and quite restricted LGM ones is consistent with glacier sequences from St. Lawrence Island (Heiser *et al.* 1994, Brigham-Grette and Hopkins 1995) and is similar to the glacial record on the Seward Peninsula and in the Brooks Range of Alaska (Kaufman and Hopkins, 1986; Hamilton, 1986; Hamilton, 1994). In Alaska, extensive glaciation occurred during the Middle Pleistocene, followed by several less extensive advances after the last Interglacial, and a relatively minor advance during the LGM (Kaufman and Hopkins 1986, Hamilton, 1994). The MITH predicts a much different pattern with widespread, young glaciation dating to the LGM.

Large glaciers and ice sheets induce a significant gravitational load on the surface of the earth and can displace the lithosphere. In other areas influenced by continental-scale ice sheets during the LGM, evidence of isostatic recovery is abundant in the form of raised beaches often reaching 10-100 m above modern sea level (Foreman *et al.*, 1995). Results of numerical modelling using a range of ice-load histories together with the ice thicknesses suggested by Hughes and Hughes (1994) result in predicted Holocene sea-level changes at four locations in the study region. On Wrangel Island, large-scale post-

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glacial uplift was predicted but no evidence of raised shorelines were found through remote sensing. In similar settings in the Barents Sea, sites near the center of the former ice sheet have experienced 100+ m of isostatic rebound (Foreman *et al.*, 1995). Similar discrepancies exist at the three other sites. At Cape Espenberg, a site on the east side of Bering Strait with a well-studied record of Holocene sea level, the MITH ice sheet should have caused 6-13 m of lithospheric depression since 5 ka B.P. due to a shifting forebulge (Figure 3.6, Table 3.1). However, sea level has remained constant since ca. 5 ka B.P. (Mason and Jordan, 1994). At Mechigmen and Uelen, there is a similar lack of evidence for changing relative sea level in the last 5000 years. Isostatic modelling of the MITH ice sheet fails to predict observed Holocene sea level trends in the Bering Strait region.

Field data from Alaska and northeastern Russia that describes the magnitudes and regional slopes of snowline during the LGM (Figure 3.7) are inconsistent with the growth of the MITH ice sheet. On the Alaskan side of Bering Strait, LGM snowline rises from the southwest to the northeast, and generally parallels present day ELAs. This pattern suggests that the moisture gradient that occurs today, increasing aridity with distance from the Bering Sea and the Gulf of Alaska, was present also during the LGM. This suggests strongly that Bering Sea has been the major source of moisture to central Beringia in glacial as well as interglacial times. In northeastern Russia, ice age snowline rises to the northeast, mirroring the pattern in Alaska and indicating the Bering Sea was the dominant moisture source for this region as well. More extensive sea ice over the Arctic Ocean during the LGM may have accentuated this moisture gradient. Near-coastal highlands bordering the Bering Sea seem like much more likely places for an ice sheet to have grown

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than over the continental shelf in the precipitation shadow north of the Chukotka Peninsula. Note also that the MITH ice sheet was "grown" over a period of 20 ka using modern precipitation rates - a dangerous assumption given the drier climate suggested by GCMs (Kutzbach, 1993) and by paleoecological data (Guthrie, 1990) for Beringia during the LGM.

To generate the MITH ice sheet, Hughes and Grosswald needed to lower regional snowline by 1000 meters in order to get ice to thicken on the continental shelf of the Chukchi Sea. However, snowline depression has been shown to be much less than this in parts of Alaska distant from the Gulf of Alaska during the LGM (Mann and Hamilton, 1995). For instance on the Seward Peninsula, ELAs were as little as 200 m below modern in Late Wisconsin times and were restricted to altitudes 400 to 500 m above present sea level (Kaufman and Hopkins, 1986). Our data (Figure 3.7) and Glushkova's (1994) suggest that LGM snowlines dipped no lower than altitudes of 700-1000 m above present sea level in northern Chukotka. Evidence from the Tenyanni Range, near Lavrenty Bay (Figure 3.5), indicates a snowline depression of only 150 meters during the LGM (Sedov 1992). Even these glaciers nearer to the modern southern coast of Chukotka and closer to maritime airmasses had LGM snowline altitudes 400-450 m above present sea level (Sedov 1992).

B. PALEOECOLOGICAL EVIDENCE AGAINST THE *MITH* ICE SHEET

For many years researchers on both sides of Bering Strait have been investigating the ice age environment of the Bering Land Bridge. These studies range in focus from

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vegetation history and climatic change, to paleoecology of Pleistocene mega fauna, to sea level history and even insect paleoecology (Hopkins, 1982, Colinveaux, 1967, Elias, 1993, Guthrie, 1990, Vartanyan *et al.* 1993). Without exception, these studies all indicate that the ice age environment of Beringia was colder and probably drier than present, with the possible exception of a mesic environment on the exposed Bering shelf (Elias, 1993). All of these investagions are based on terrestrial fossils that indicate life was present during the full glacial and that Beringia was not glaciated nor did it experience the extreme periglacial climate expected at the margins of a continental scale ice sheet.

Perhaps the most convincing paleoecological evidence against the existence of the MITH ice sheet is the long tenure of mammoths on Wrangel Island. Radiocarbon dated mammoth bones indicate these animals continuously occupied Wrangel Island from approximatesly 20,000 to 3500 years ago (Vartanyan *et al.* 1993, Vartanyan et a.l 1995). The only hiatus in fossil ages occurs between 12,000 and 7000 y BP Considering that it took 20,000 years to grow the postulated ice sheet, it seems very unlikely that its nucleus was in the vicinity of Wrangel Island.

V. CONCLUSIONS

Our studies of glacial landforms indicate that glaciers advanced radially from the mountainous areas of Chukotka. There is no evidence for ice encroaching from the north

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or overriding the peninsula. Similarly there is no evidence, anywhere in the region for an isostatic response to a MITH ice sheet. Paleoclimate records and ELA reconstrutions show that the climate necessary to form the MITH ice sheet probably did not exist in Beringia during the last ice age. The former presence of plants and animals in the hypothesized location of the MITH ice sheet provide what we feel is conclusive evidence against this bold hypnthesis.

Could the MITH ice sheet have existed during earlier glacial periods? Our SAR results suggest that at least the last three glacial advances in Chukotka originated in local mountains and flowed radially from the uplands. Certainly the Alaskan records suggest that no MITH - style ice sheet existed there since at least the early Pleistocene (Hamilton 1994, Hamilton *et al.* 1986).

Although we find no evidence for the existence of a MITH-type ice sheet in LGM Beringia, the hypothesis is by no means mythical. It is interesting to consider the > 100 year debate surrounding the Barents Sea ice sheet (Solheim *et al.*, 1990; Polyak *et al.*, 1995), an argument settled recently in decisive favor of a glacial form previously only dimly imaginable by most terrestrially based geologists.

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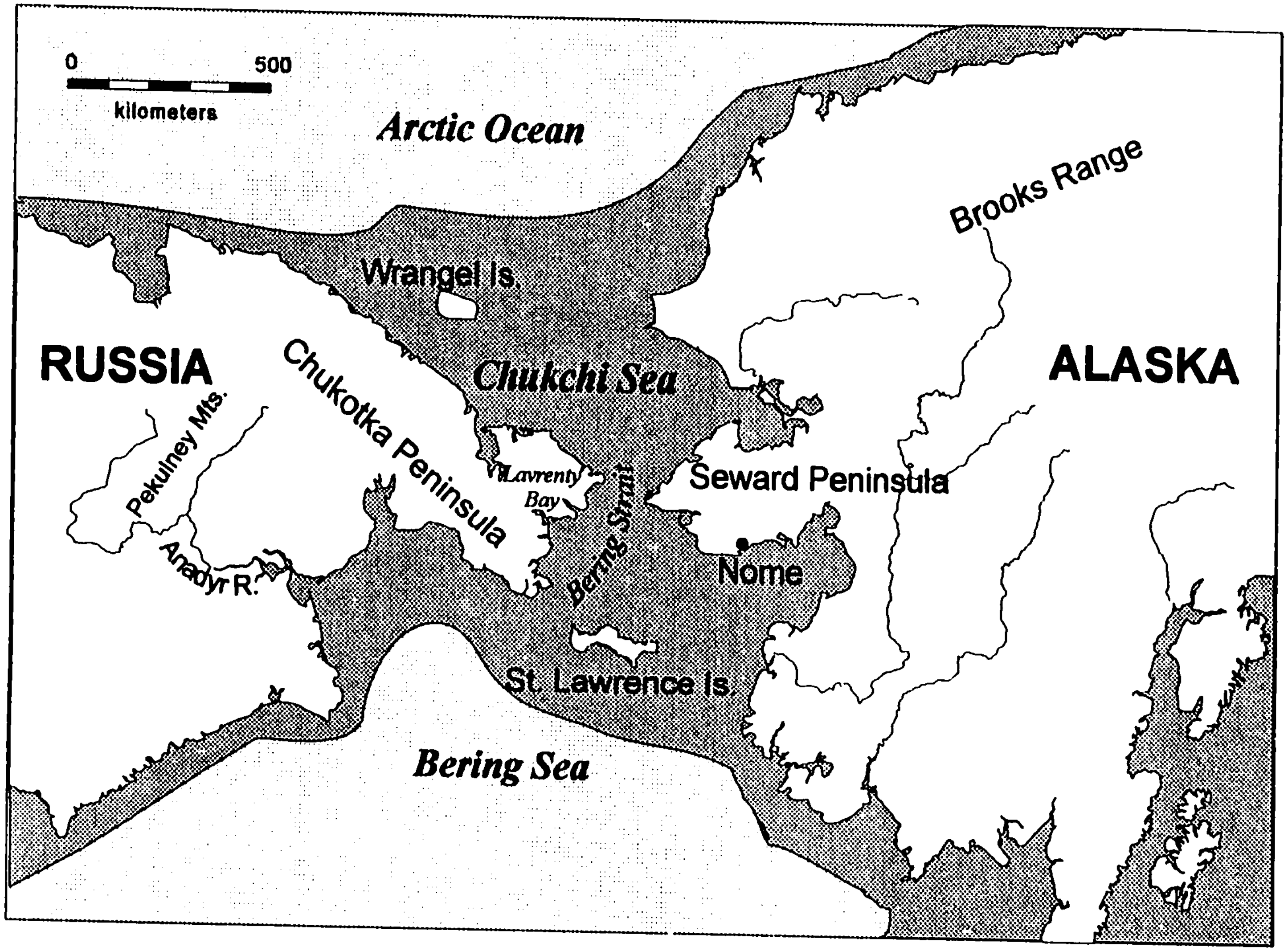

Figure 3.1 Map of Beringia showing locations mentioned in the text. Extent of the Late Wisconsin land connection is shown in dark grey.

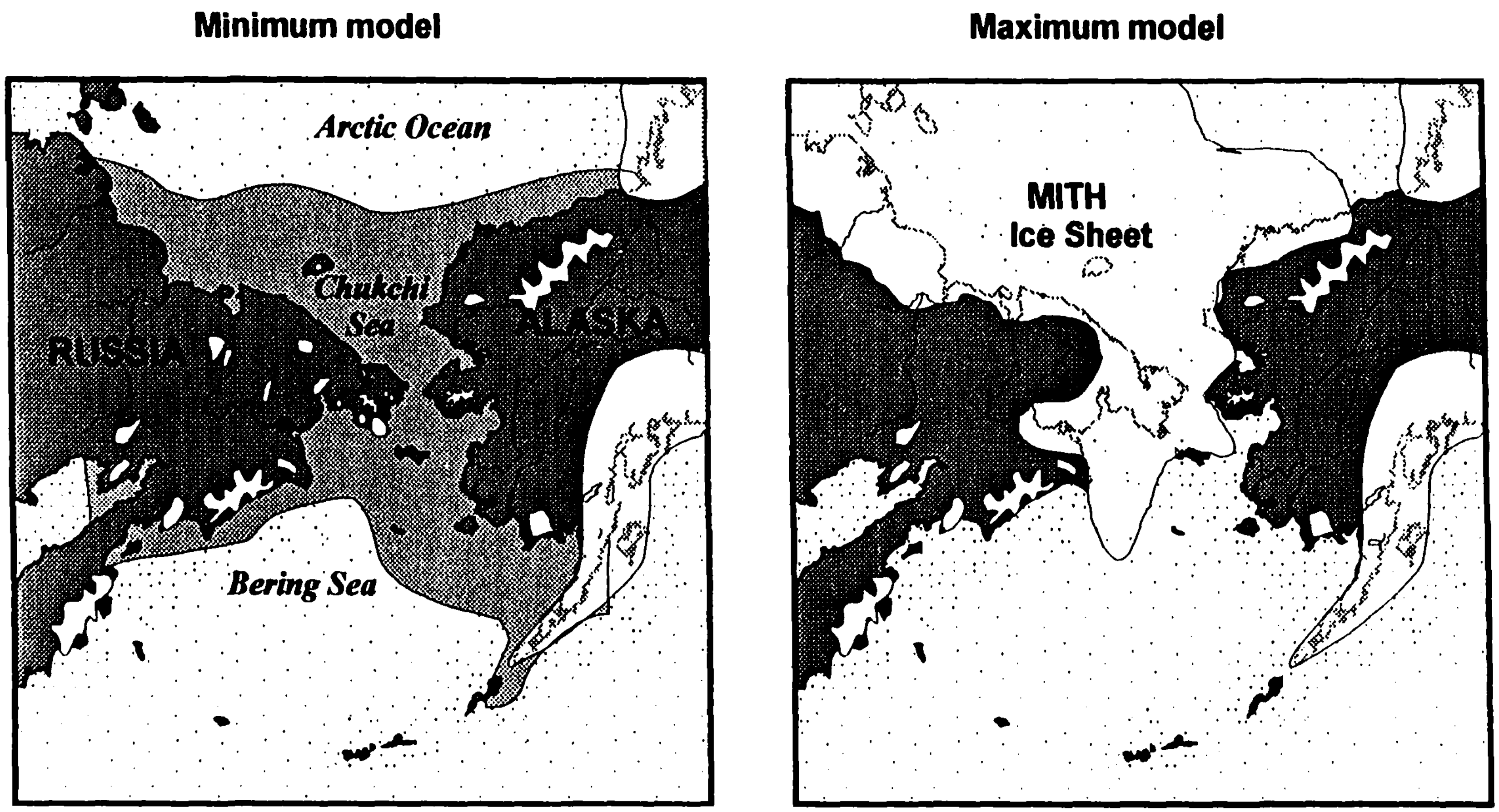

Figure 3.2. Two models of glacial extent during the Late Wisconsin. In the minimum model, the lowering of global sea level and emergence of the Bering Land Bridge, induced a drier continental climate in Beringia. Moisture starved glaciers advanced only slightly beyond mountain fronts. The postulated ice sheet in the maximum model was 'grown' in a numerical model that called for a regional snowline lowering of 1000 meters below present. The MITH calls for ice doming on the Chukchi Sea shelf and subsequent southward flow over Chukotka and the land bridge.

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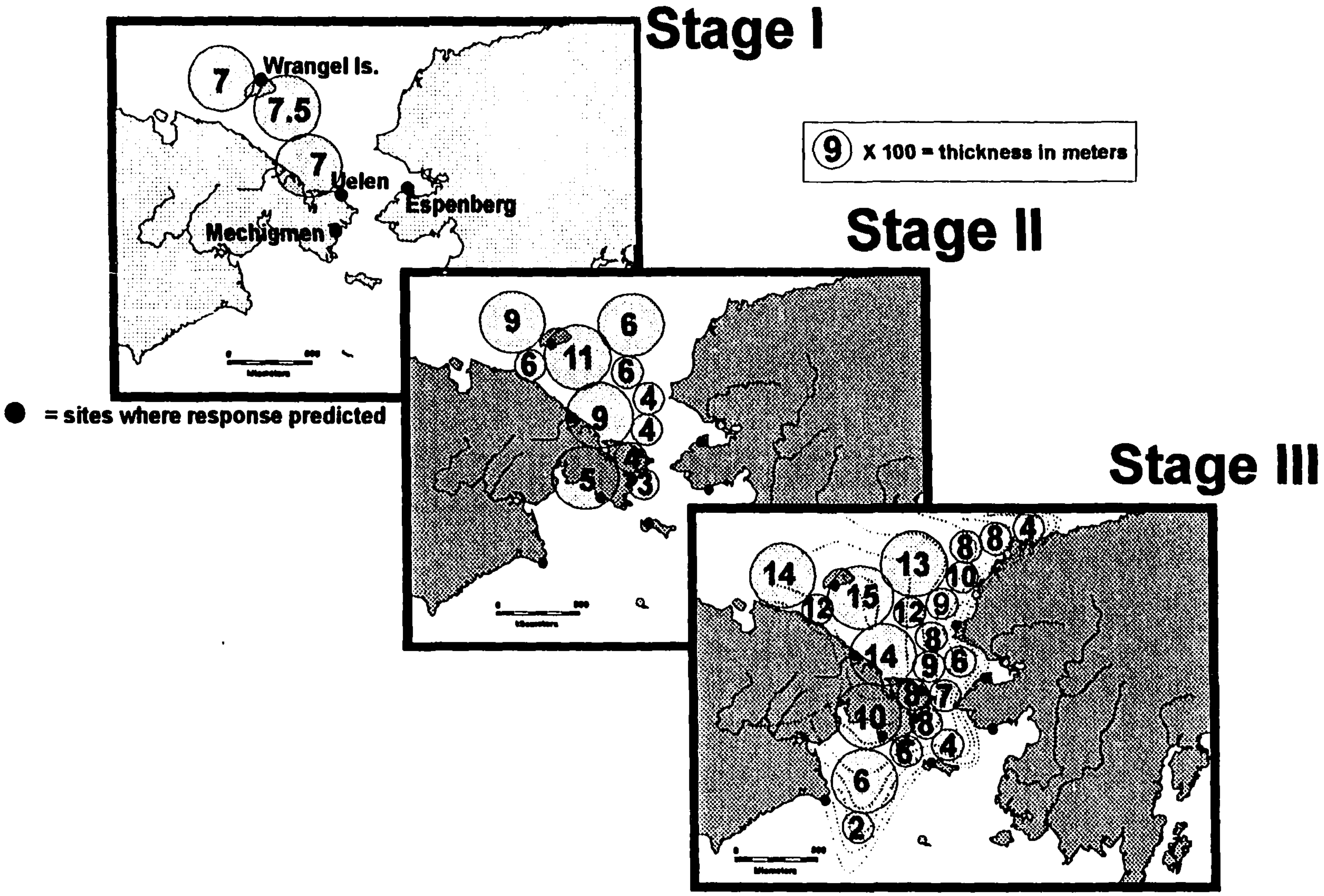

Figure 3.3 . Three stages of glacier growth define load history of MITH ice sheet. The ice sheet is represented as a series of disc loads of changing thickness.

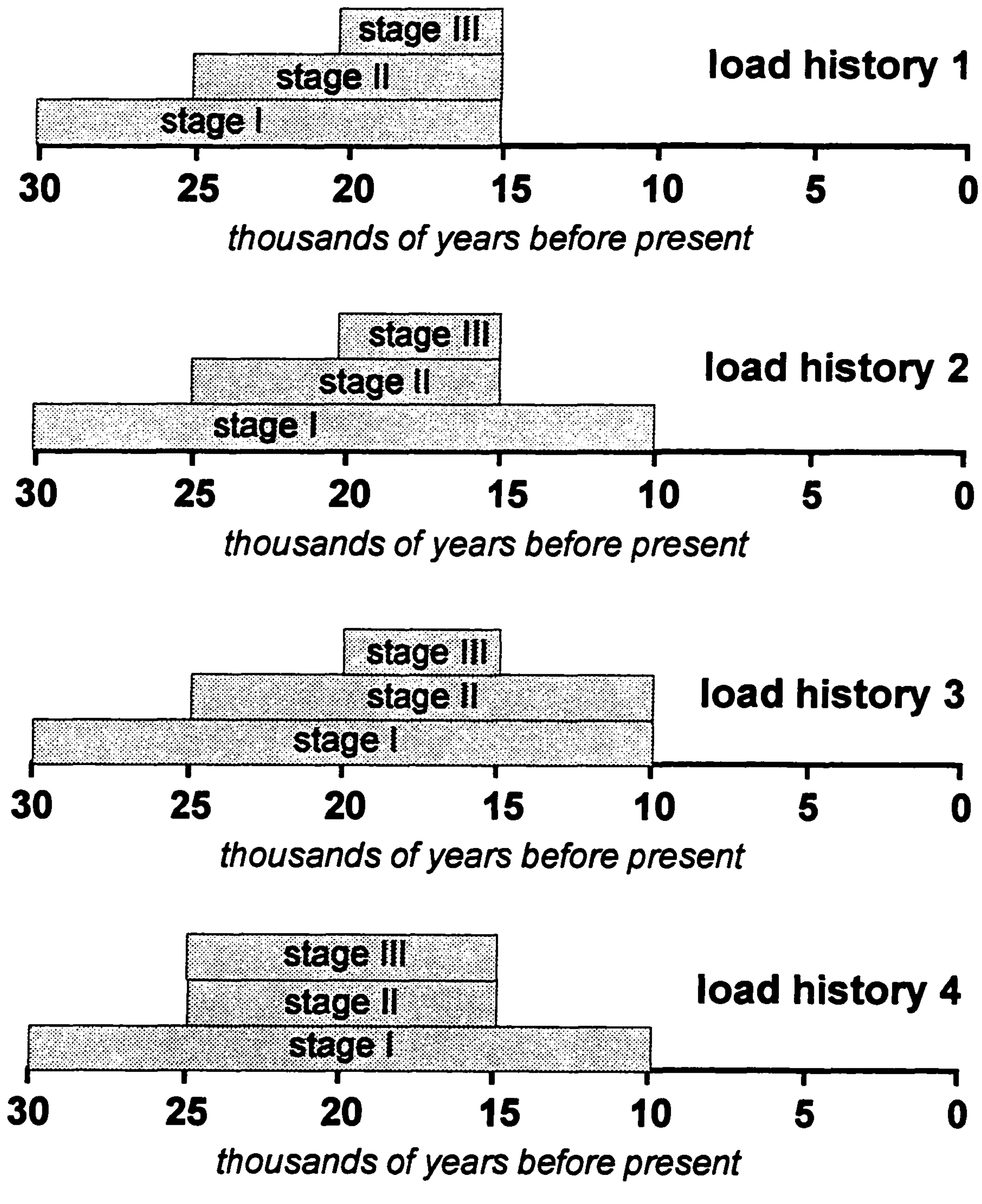

Figure 3.4. Schematic diagram of four load histories used in this study.

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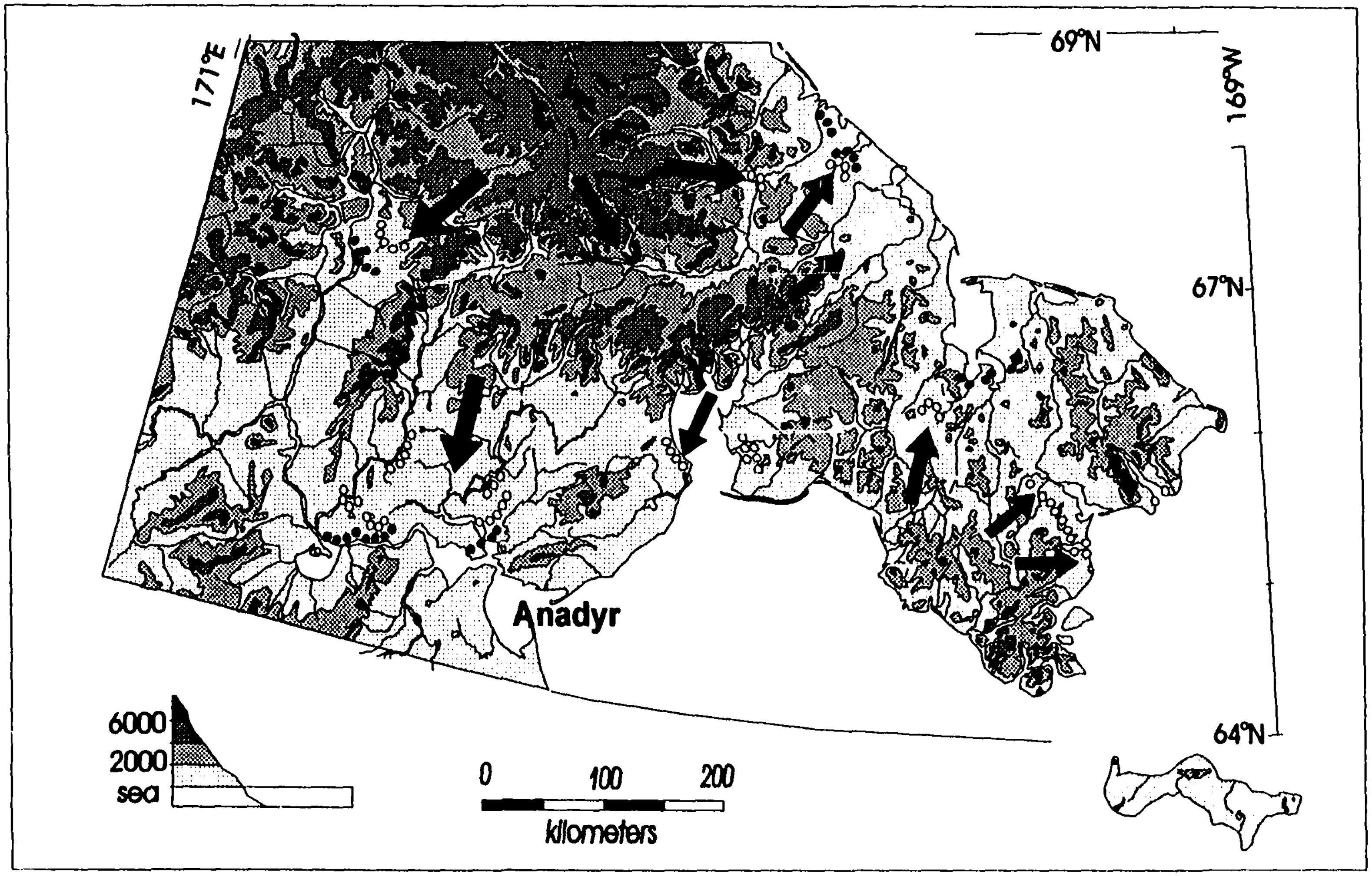

Figure 3.5. Glacial moraines on the Chukotka Peninsula mapped from SAR imagery. Black arrows show radial flow out of mountians. Estimated relative age of moraines indicated by symbols as follows: Youngest Glaciation ( ), Penultimate Glaciation (oooooo) and Oldest Glaciation (••••••). See Table 2.2 for summary of criteria used to estimate relative age.

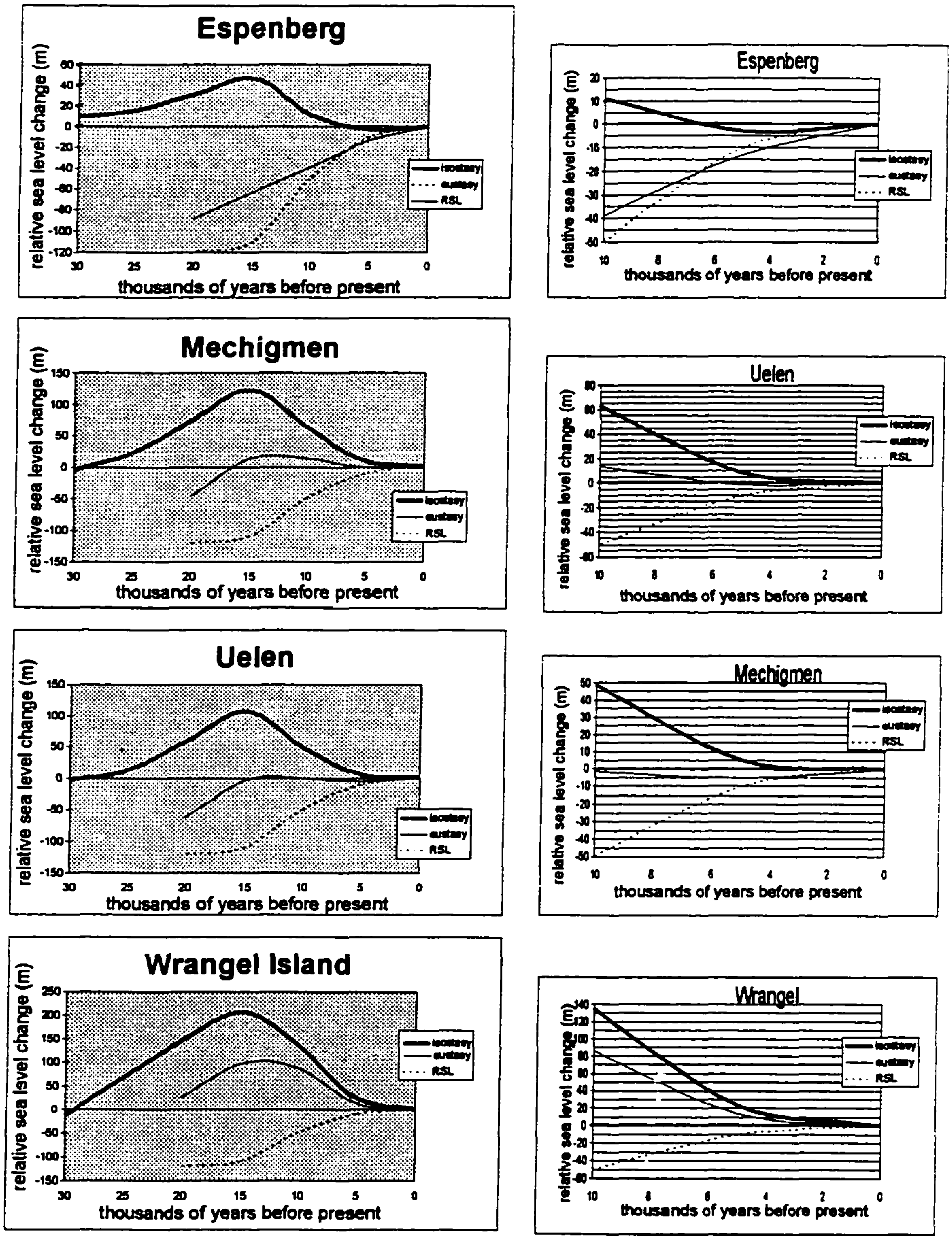

Figure 3.6. Relative sea level changes predicted by the modeled solid earth deformation in response to the MITH ice sheet.

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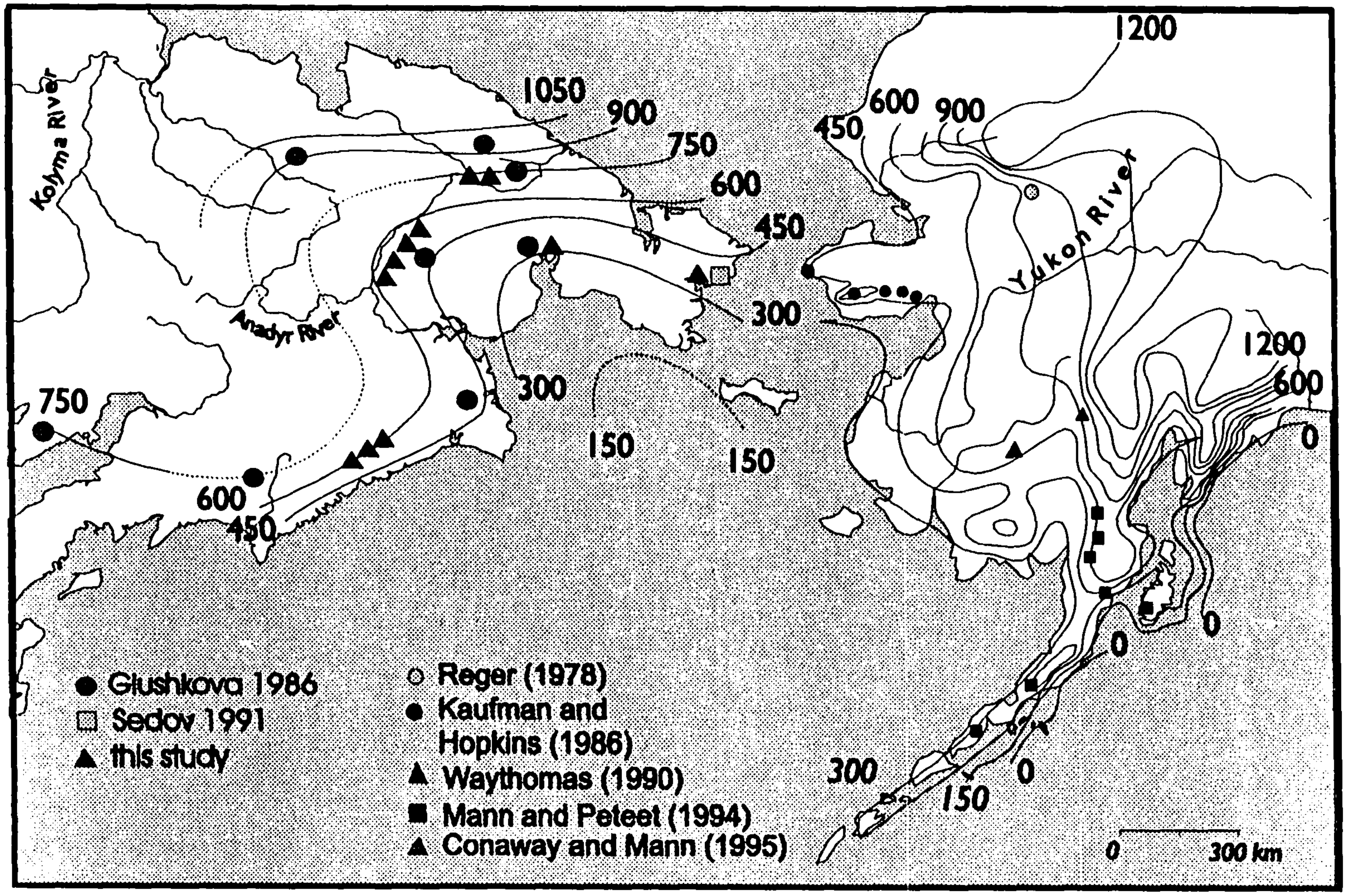

Figure 3.7. Paleosnowline reconstructions forthe LGM in Beringia Beringia. Alaska reconstructions based on previous work of others. Snowline reconstruction in Russia based on work of Glushkova (1992) and Sedov (1990) and from the results of this study. LGM snowline in Russia appears to mirror the trend in Alaska, rising to the northwest from ~ 300 m near Bering Sea to 1000 m on the north shore of Chukotka. (Alaska contours based on Pewe 1975.)

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Table 3.1 Relative sea level change (in meters) expected to have occurred over the last 5000 years at several locations in Bering Strait. The data is shown for four sea level curves obtained using four different load histories (see Figure 3.3)

	LH 1	LH 2	LH 3	LH 4
Wrangel Island	-20	-15	-14	-18
Uelen	+4	+5	+1	+3
Mechigmen	+5	+9	+5	+6
Cape Espenberg	+6	+12	+13	+13

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Table 3.2. Equilibrium line altitude reconstructions in Chukotka for the Last Glacial Maximum.

Valley	cirque floor		head wall * (feet)		toe (feet)		head wall ** (m)		toe (m)		THAR value 0.6			MEG (m)		
	feet	(m)	max	min	max	min	max	min	max	min	max	min	ave (m)	max	min	ave (m)
Pekul'ney																
1	1500	**457**	2500	2000	800	700	762	610	244	213	555	451	**503**	503	411	457
2	2000	**610**	3000	2500	700	600	915	762	213	183	634	530	**582**	564	472	**518**
3	1500	**457**	3500	3000	600	500	1067	915	183	152	713	610	**661**	625	533	**579**
4	1500	**457**	3000	2500	700	600	915	762	213	183	634	530	**582**	564	472	**518**
5	1500	**457**	2500	2000	500	400	762	610	152	122	518	415	**466**	457	366	**411**
6	1250	**381**	2500	2000	400	350	762	610	122	107	506	408	**457**	442	358	**400**
Amguema																
1	2000	**610**	4000	3500	1100	1000	1220	1067	335	305	866	762	**814**	777	686	**732**
2	2000	**610**	3500	3000	1000	900	1067	915	305	274	762	658	**710**	686	594	**640**
Tenyanni																
Vulcanaya	1000	**305**	2000	1500	250	0	610	457	76	0	396	274	**335**	343	229	**286**
2	1000	**305**	2000	1500	250	0	610	457	76	0	396	274	**335**	343	229	**286**
3	1000	**305**	2000	1500	250	0	610	457	76	0	396	274	**335**	343	229	**286**
Egvekinot																
1	1000	**305**	2000	1800	200	50	610	549	61	15	390	335	**363**	335	282	**309**
Koryak																
1	2000	**610**	3000	2500	1200	1000	915	762	366	305	695	579	**637**	640	533	**587**
2	1500	457	3000	2500	800	500	915	762	244	152	646	518	**582**	579	457	**518**
3	2500	762	3500	3000	1000	800	1067	915	305	244	762	646	**704**	686	579	632

* obtained from TPC charts in feet

**feet converted to meters

THAR: in meters

MEG: in meters

CHAPTER 4

SURFICIAL GEOLOGY AND PLEISTOCENE HISTORY OF NORTHWESTERN ST. LAWRENCE ISLAND

Introduction

St. Lawrence Island is located just south of present day Bering Strait and near the southern shore of the land bridge that joined Asia and North America during ice ages (Figure 4.1). The Quaternary sediments found on the island provide an important stratigraphic record, a Beringian 'Rosetta Stone', for linking the glacial and sea level records of western Alaska and the Chukotka Peninsula. Our field efforts in 1991, 1992, and 1993, combined with the earlier work of D. M. Hopkins and others has enabled us to compile a surficial geologic map and stratigraphic scheme for the Quaternary deposits on St. Lawrence Island.

Geographic and Geologic Setting

St. Lawrence Island is located approximately 75 kilometers southeast of the southern tip of the Chukotka Peninsula and 150-200 kilometers from the Alaskan mainland (Figure 4.1). The island experiences a strong maritime climate with a mean annual air temperature of -3.9° C and annual precipitation of 39cm. The island is located within the limit of seasonal sea ice and has shore-fast ice for much of the winter.

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The dominant bedrock features of western St. Lawrence Island are quartz monzonite plutons of Cretaceous age (Patton and Csejtey 1980). These rocks are part of the Okhotsk-Chukotsk volcanic belt that runs along eastern Chukotka and into Bering Strait. Tertiary claystone, conglomeritic sandstone, mudstone and lignite are exposed along the western shore of Nivrukpuk Lagoon (Figure 4.2). These rocks are poorly consolidated and glaciotectonically deformed (Hopkins *et al.* 1972, Patton and Csejtey 1980).

Previous Work

W.W. Patton and D.S. McCulloch first discovered fossiliferous interglacial marine deposits around Nivrukpuk Lagoon in 1966 (Hopkins *et al.* 1972). D.M. Hopkins visited the island two years later to look for evidence of an hypothesized ice cap that might have originated in the mountains of Chukotka and encroached on the island (Hopkins *et al.* 1972, Benson 1994).

In their 1972 paper, Hopkins and colleagues correlated the interglacial deposits with the Middle Pleistocene Kotzebuan Transgression (Hopkins *et al.* 1972). However, the type section for the Kotzebuan Transgression has now been determined to be correlative with the Anvillian Transgression recorded at Nome, and the term 'Kotzebuan' has been dropped. The Anvillian marine transgression has been dated to ~ 410 ka BP, or oxygen isotope stage 11 (Brigham-Grette and Hopkins 1988, Kaufman *et al.* 1991). Hopkins *et al.* (1972) also correlated these interglacial marine sediments at Nivrukpuk Lagoon with Petrov's (1966) Kresta Suite in Chukotka, Russia.

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Glaciotectonic deformation structures, the presence of till over the marine sediments, and the discovery of glacial erratics foreign to St. Lawrence Island led Hopkins and his colleagues to conclude that the interglacial marine sediments had been overridden by glacial ice that originated in the mountains of Chukotka and that this glacial advance probably occurred in the Middle Pleistocene (Hopkins *et al.* 1972).

In 1988, Hopkins and J. Brigham-Grette collected mollusk shells from the interglacial sediments found along the western shore of Nivrukpuk Lagoon, Penguqusiq Spit and the mouth of Aghnaghak Lagoon (Figure 4.2) (spelling of Yupik place names here are based on spellings provided by residents of the island rather than those found on USGS maps). These shells yielded amino acid aIle/Ile ratios of *two* distinct age groups. The ratios were correlative with ratios from the same genera collected at Nome and suggest correlation with the Anvillian and Pelukian marine transgressions recorded there (Benson 1994, Brigham-Grette and Hopkins 1995). The observation that the interglacial marine sediments of Pelukian, or Last Interglacial, age (125,000 BP) were glaciotectonically deformed like the Anvillian sediments, suggested that Early Wisconsin glaciation in Chukotka may have been more extensive than previously thought (Petrov 1966, Hopkins *et al.* 1972).

Map Units and Boundaries

The main Pleistocene stratigraphic units mapped on northwestern St. Lawrence Island are 1) Anvillian marine sand, 2) Nome River age drift, 3) Pelukian marine gravels,

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and 4) 'Sivuqaq' (probably Early Wisconsin age) drift. In very few locations were definitive contacts found, so the boundaries of these units (Figure 4.3 and Plate 1: Figure 1) are estimated in most places and designated by dashed lines. The rationale for determining specific boundaries is discussed below.

The southernmost limit of the Nome River drift is estimated only by the presence of rare erratics found south of Surprise Lake along the rivers draining Aghnaghak Lagoon and Nivrukpuk Lagoon. The low hills in the southern part of the island are mantled with colluvium, but erratics have not been observed there. It is possible that glaciers advanced west of Poovookpook Mountains, but at present no convincing evidence of glaciation has been found in this area.

The assumed limit of (drift-covered) Anvillian marine transgression is estimated from the presence of these sediments at Surprise Lake and on altitude above sea level. Because the distribution of Anvillian sedimats is largely obscured by overlying Nome River Drift, we have arbitrarily assumed an inland limit at about 25 m above sea level (asl). Anvillian beaches near Nome lie at 22 m asl, and Kaufman and Brigham-Grette (1993) suggest that this is a good approximation of the amount of sea-level rise during the Anvillian transgression. Mollusk shells with Anvillian aIle/Ile racemization ratios are present in marine gravels at along the northern shore of the island. However the presence of abundant shells with Pelukian (Last Interglacial) aIle/Ile ratios in those same units suggest that the Anvillian shells were reworked during the subsequent transgression. The existence of intact Anvillian strata near the northern coast of the island is unlikely.

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The extent of the Pelukian Transgression is clearly seen in stratigraphy exposed along the eastern shore of Aghnaghak lagoon where the marine gravels (representing beach deposition) pinch out and swash debris is preserved. Along the western coast of the island and at Mugum peak, a cinder cone east of Taphook point, rocky headlands are notched with wave cut terraces at approximately 8 meters above present sea level. This altitude is consistent with the height of Pelukian shorelines around Bering Strait (Brigham-Grette and Hopkins 1995). The limit of the Pelukian transgression elsewhere is arbitrarily placed at the 8 meter contour.

Avruwaq Bluff at Penguqusiq Spit

Penguqusiq Spit separates Nivrukpuk Lagoon from the Bering Sea (Figure 4.2). Coastal erosion on the north side of Penguqusiq has formed Avruwaq Bluff exposing approximately 1.5 kilometers of stratigraphy along the spit. Up to 13 vertical meters of section is exposed in the bluffs (Figure 4.4 and Plate 1:Figure 2). The sediments represent a spit-like complex of beach sand and gravel, with occasional clay-silt units, channel forms, and low-angle cross-sets (Benson 1994). In several places, plugs of glacial diamicton are intruded, into the marine sands and gravel. These plugs include large granitic boulders indigenous to the island as well as glacially sculpted and striated schist, limestone, and quartzite thought to be from the Chukotka mainland.

Sonja Benson's thesis work included a detailed study of the glaciotectonic structures at Avruwaq. She showed that the sediments there were deformed by a glacial

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advance from the northwest (Benson 1994). Most of the fault planes, regardless of direction and degree of dislocation, strike north to northeast and a major thrust fault dips to the northwest (Benson 1994). Benson interpreted horst and graben features and normal faults as evidence of subglacial extension. However, these extensional features could also result from proglacial deformation associated with the gentle folding of the sediments that occurs throughout the section.

The maximum position of the Sivuqaq glacial advance estimated using glacial tectonic structure and stratigraphy. The glaciotectonics are consistent with proglacial sediment deformation, and over-riding in places (Benson 1994). In only a few locations is glacial till observed and it is typically intruded into, rather than superposed over, the deformed marine sediments. In 1972, however, Hopkins reported that much of the section at Penguqusiq was capped with till and his photos from that visit show a distinct layer of till (~ 1 meter thick) on top of the section. Coastal erosion at this site is rapid that most of the section preserving glacial till has been eroded. Evidence of both pro-glacial deformation and over-riding suggest that present day Penguqusiq was near the terminal position of most recent glacial advance recorded on norhtwester St.Lawrence Island. The rare occurrence of erratics on the back side of Nivrukpuk Lagoon suggests that tongues of ice may have reached further inland. However, it is impossible to distinguish erratics left by the Nome River advance from those left by the Sivuqaq advance.

The glacial advance that deformed the marine sediments and deposited the now largely vanished till at Avruwak is clearly post-Pelukian, isotope stage 5, but older than

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the late Wisconsin cold period, oxygen isotope stage 2. Benson (1994) found that Mya and Hiatella shells collected at Avruwak have aIle/Ile ratios indistinguishable from amino acid ratios in shells collected from Pelukian (last Interglacial) depsits near Nome, reported by Kaufman (1992). Several radiocarbon dated shells collected at Avruwak, including some of those with the lowest aIle/Ile ratios, are all infinite, that is, their age lies beyond the limits of radiocarbon dating. Had the Sivuqaq glacial advance crossed the strait between St. Lawrence Island and the Chukotka Peninsula, as recently as the last cold period 9isotope stage 2)then we wild expect to find at least some younger radiocarbon and aIle/Ile ratios from shells transported to the island. Further, the presence of ice wedges, ice wedge pseudomorphs, and silt-infiltrated and cryoturbated surfaces on the marine and glacial deposits near Avruwak suggest that the shelly lmarine grabel there must have been extant theough at least one cold period. It is clear that the glacial advance that deformed and partly over-rode the hills at Avruwak musthave taken place no later and possibly slightly earlier than the cold period that marked oxygen-isotope stage 4.

Aghnaghak Lagoon

Numerous exposures along both shores of Aghnaghak Lagoon show the stratigraphic and spatial relationship between the main sedimentary units found on the northwest part of the island. An idealized cross section (Figure 4.5, Plate 1:Figure 3) along the transect A-A' (location shown in Figure 4.2 and Plate 1: Figure 1) illustrates these relationships.

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At Surprise Lake the Anvillian transgression, and possibly an older one, are recorded by shelly marine sands exposed in the eroding shore of the lake. Shells obtained from a lower unit characterized by grey, fine sand yield amino acid ratios too high (mean ratios from *Mya* 0.303 $\pm$ 0.014) to correlate with the Anvillian Transgression but too low to correlate with ratios from the Beringian III transgression described from Nome (mean total aIle/Ile 0.445 + 0.03) (Benson 1994, Kaufman 1992). *Astarte* shells obtained from an upper unit composed of tan medium sand yield quite low ratios that could be grouped with Pelukian aged fossils (Benson 1994). It seems unlikely, however, that Pelukian shells could occur at Surprise Lake, since it is 9-10 meters above present sea level (several meters above the nearby Pelukian wave cut terraces) and well beyond the clearly marked limit of the Pelukian transgression (discussed below). Kaufman (1992) showed that *Astarte* shells may epimerize 50 to 85% the rate of *Mya* or *Hiatella* so these ratios on *Astarte* shells are likely of Anvillian age (Benson 1994).

All the marine sediments at Surprise Lake are slightly deformed by glacial tectonism. A thin layer of till is exposed in places and the surrounding area is mantled by glacially sculpted and striated erratic cobbles and boulders. Further north, along the shore of the lagoon, these marine sands are exposed and show evidence of deformation and glacial over-riding. Although good exposures of till are infrequent, the deformation of the sediments and abundant erratics indicate that the area was glaciated.

The marine limit of the Pelukian transgression is clearly indicated where the shelly beach gravel and sand pinch out and organic deposits indicating beach (storm?) deposition

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mark the limit of the transgression. Organics from this material (detrital wood and sea weed) yielded an infinite radiocarbon date. North of here, erratic cobbles and boulders from the Nome River drift show signs of being 'washed' and redeposited. The Nome River drift and Anvillian sands are truncated by the wedge of Pelukian beach gravels.

The Pelukian beach sediments are, in turn, deformed by glacial tectonism (see description from Penguqusiq Spit). At the mouth of Aghnaghak Lagoon there is a small plug of glacial till, indicating the encroachment of ice into, and possibly over, the interglacial gravels. Thick units of Holocene peat have formed in depressions around the mouth of the lagoon.

Konguk Basin

In the southern part of the study area, there is evidence of cirque glaciation in the hills surrounding Konguk Basin (Figure 4.3 and Plate 1:Figure 1). At Ivekan Mountain, a deep valley with steep headwalls suggests repeated glaciation. A small cirque truncated by the present coast shows relatively fresh headwall erosion and possibly lateral moraine features along the sidewalls.

Determining the age of these features and the glacial cirques themselves is extremely difficult. There are several very subdued, low angle, degraded moraine-like features at the mouth of the Ivekan cirque. There is not sufficient data (not enough ciques or moraines) to make relative age estimates comparable to those made in studies of

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moraines on the Seward Peninsula (Kaufman 1986, Kaufman 1988). Age estimates here are based on our present understanding of the regional glacial history and the relative position and geomorphic degradation of the moraines. The most extensive moraine left at the mouth of the long valley cut into Ivekan mountain is tentatively correlated with the Nome River glaciation. The intermediate moraines are similarly correlated with the Sivuqaq glaciation, and the fresher material in the Ivekan valley and the cirque near the coast are assigned a possible Late Wisconsin age.

Reconstruction of the youngest glacier to occupy Ivekan mountain showed that snowline was lowered to 150 meters above present sea level. This snowline depression is consistent with the regional trend, which shows maximum lowering near the center of the Bering Strait region during Late Wisconsin time (see Chapter 3, Figure 3.7). It is also possible that the fresh material in these cirques was deposited by rock glacier activity during the Late Wisconsin. There is no indication that rock glaciers are present today. Further studies, possibly including exposure age dating of the headwalls, using cosmogenic isotopes, is needed before definitive age estimates can be made.

Post Glacial Features

Most of the Pleistocene deposits found on the island have been at least slightly reworked by cryotubation, frost jacking, and the formation or melting of ground ice. The upper portion of the stratigraphic units exposed along the shores of Aghnaghak and

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Nivrukpuk Lagoons, and along Avruwaq Bluff all show signs of ground ice activity and silt infiltration.

Thaw lakes cover a large part of the lowland ground surface and the shape of Aghnaghak Lagoon suggests it may have been formed by the flooding and subsequent coalescence of thaw lakes. Although there is evidence of recent drainage of some thaw lakes, they appear to be relatively stable compared to lakes on the coastal plain of the Seward Peninsula and on the northern coastal plain of Alaska. This may be due to the lack of deep syngenetic ice wedges, which may not be as extensive on St. Lawrence as in locations that have been subjected to abundant and long term aeolian deposition. The lakes on St. Lawrence seem to be controlled partly by topography, which also suggests only a thin aeolian mantle and relatively lower ground ice content.

Holocene peat has formed in topographic depressions around the mouth of Aghnaghak Lagoon and Nivrukpuk Lagoon and is also found along the shores of some lakes. At the mouth of Aghnaghak, a thin (3 cm) layer of tephra was found in the peat and dated to 3290 BP. This age suggests that the tephra is from the Aniakchak eruption and that this locality is an extension of the area known to be covered by the tephra plume from that eruption.

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Conclusions

The sedimentary record on western St. Lawrence Island evidences the encroachment of glaciers originating in the mountains Chukotka at least twice during the Pleistocene. The first event probably occurred shortly after 410,000 years BP and is correlated here with the Nome River glaciation at Nome. The second advance occurred sometime after the Last Interglacial, probably during Early Wisconsin time (oxygen isotope stage 4 or late stage 5, 70,000 -90,000 years BP). The interglacial marine sediments on the island are correlated, using stratigraphic relationships and amino acid relative age dating, with the Middle Pleistocene aged Anvillian transgression and the Last Interglacial Pelukian Transgression recorded on the Alaskan mainland.

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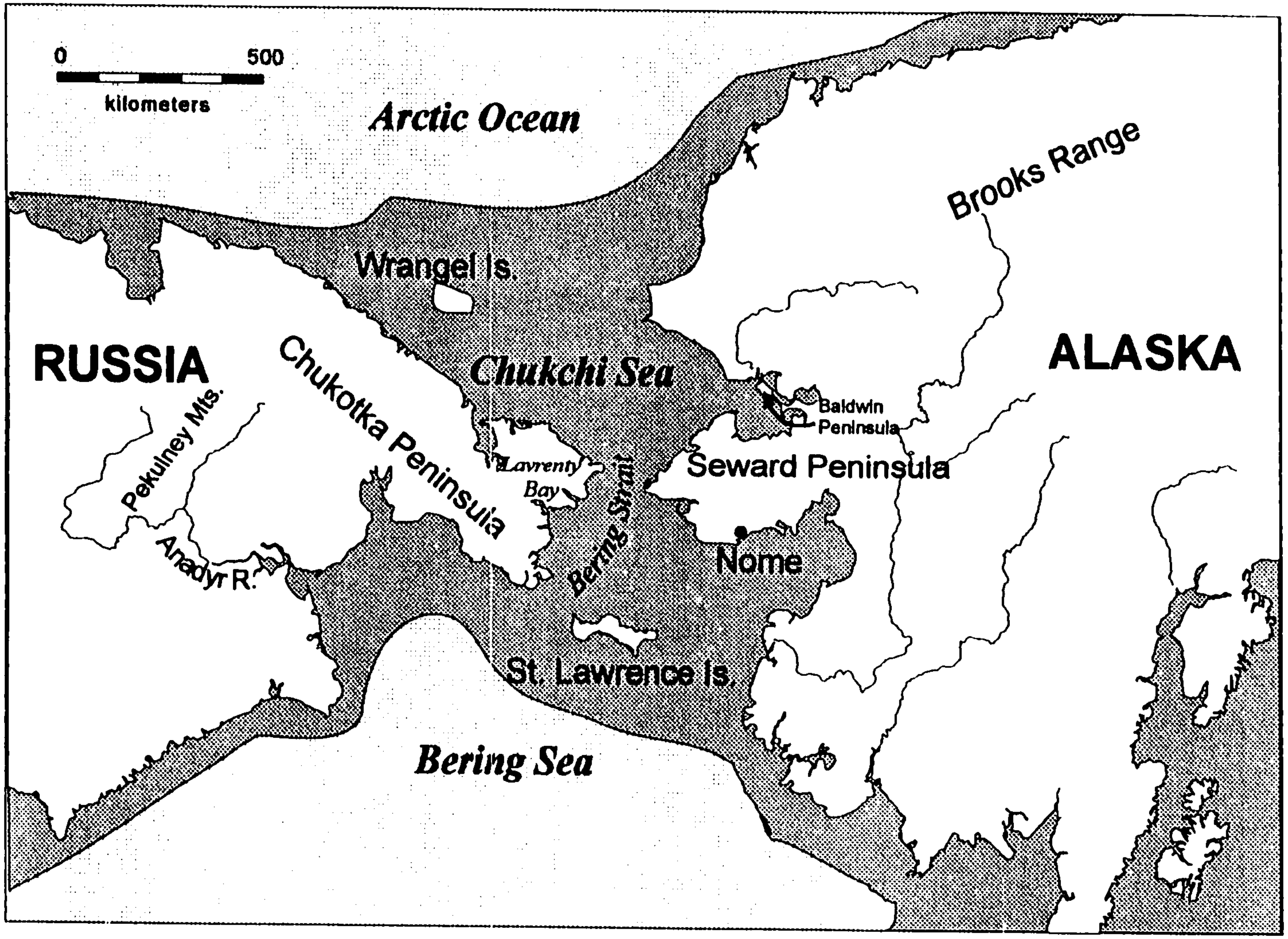

Figure 4.1 Map of Beringia showing locations mentioned in the text. Extent of the Late Wisconsin land connection is shown in dark grey.

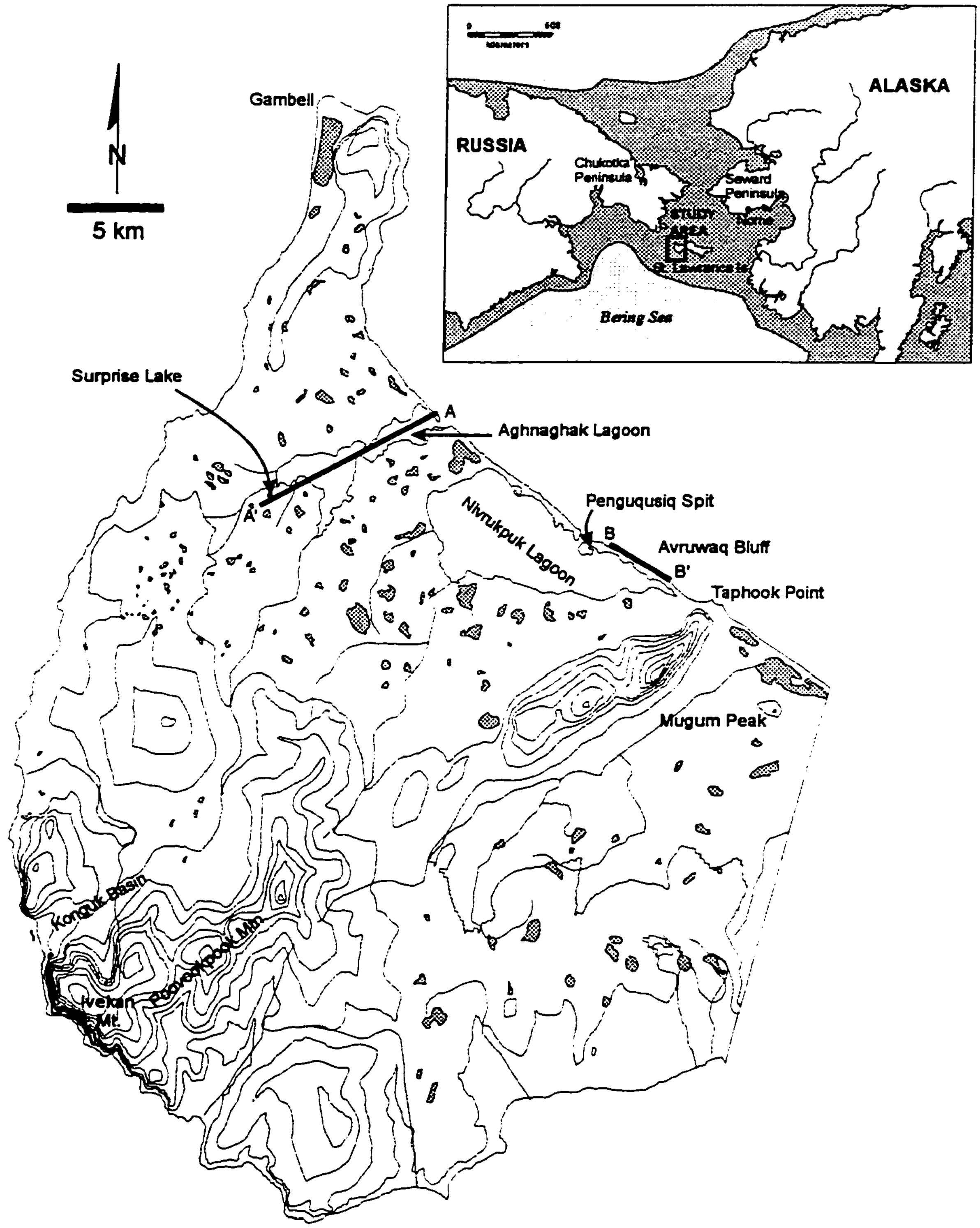

Figure 4.2. Map of northwestern St. Lawrence Island showing locations mentioned in the text. Inset shows location of the island in Bering Strait.

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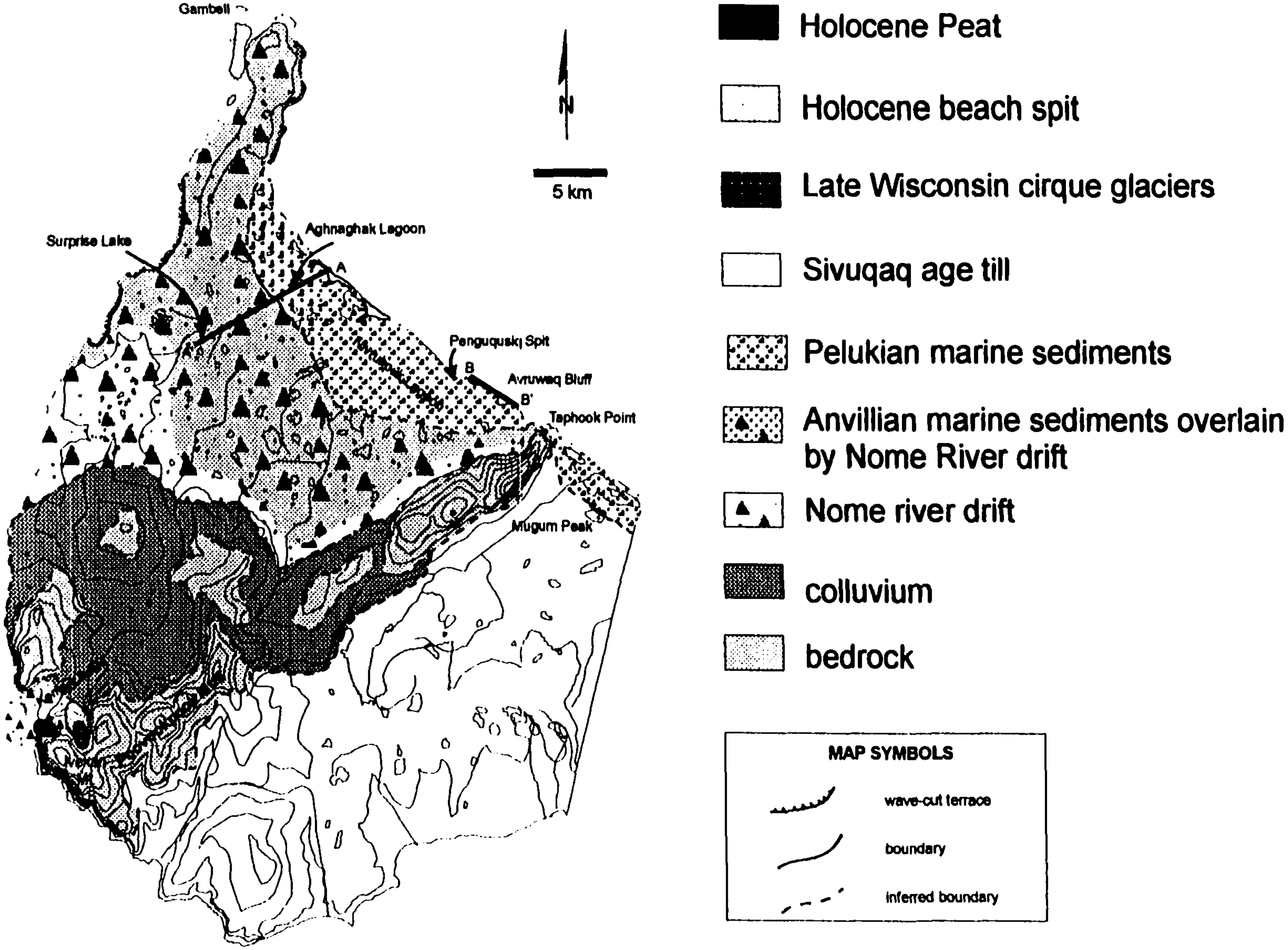

Figure 4.3. Generalized surficial geologic map of northwestern St. Lawrence Island.

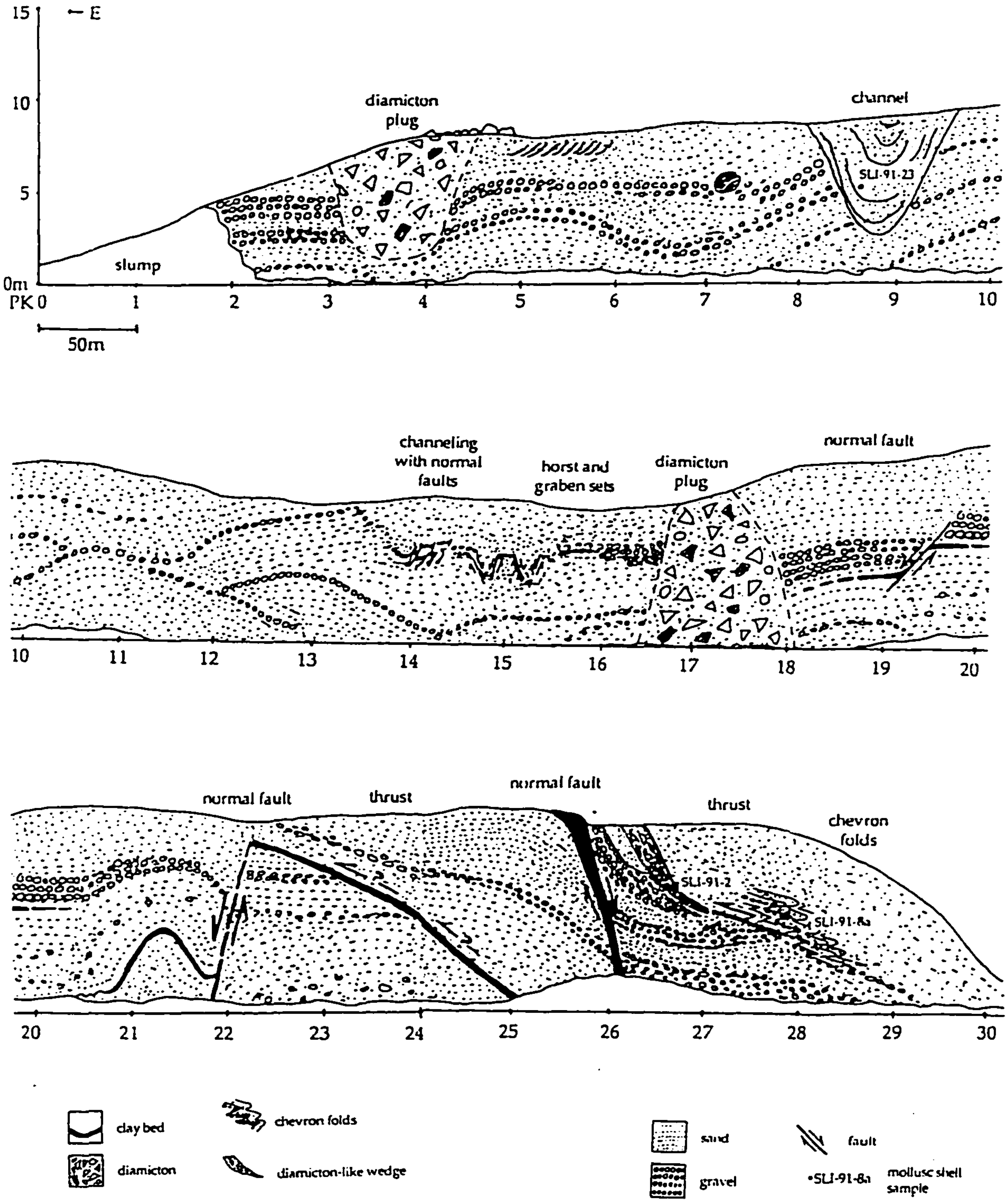

Figure 4.4. Cross section (B-B') of sediments exposed at Avruwaq Bluff. Glaciotectonic structures and till plug suggest a marginal glacial environment. (from Benson 1994). Vertical exageration 10x.

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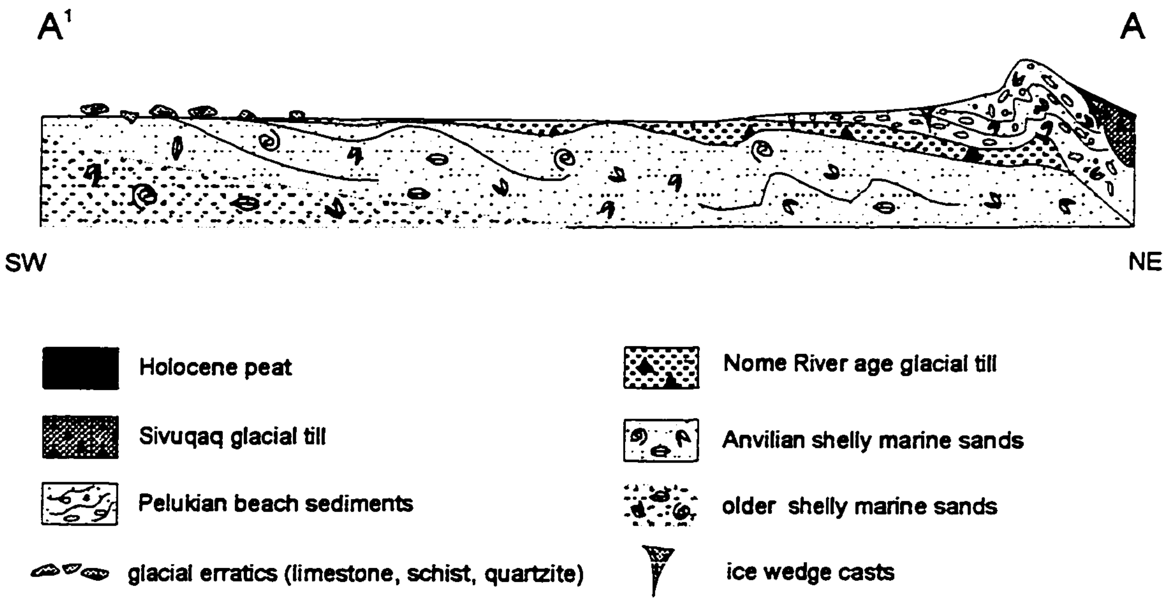

Figure 4.5. Idealized cross-section of stratigraphy exposed along the shores of Aghnaghak Lagoon, St. Lawrence Island.

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CHAPTER 5

CONCLUSIONS AND FUTURE WORK

Summary and Conclusions of this Dissertation:

This study employed a multidisciplinary approach to the investigation of the extent, timing, and potential effects of repeated Pleistocene glaciation in Bering Strait region. The application of satellite remote sensing proved very useful to the study of glacial geology in Russia. The use of synthetic aperture radar (SAR) allowed us to produce the first comprehensive, regional map of glacial features in northeast Siberia. One of the crucial elements in the argument against the southward encroachment of a large ice sheet over Chukotka, was the discovery of several generations of moraines that mark the terminal positions of northward flowing glaciers. The record of re eated mountain glaciation, characterized by radial flow out of high topographic areas, was further evidence against the existence of a continental scale ice sheet in Beringia at any time in the latter part of the Pleistocene.

The utilization of the geometric relationships inherent in SAR technology also proved very useful for estimating the relative age of glacial features. The qualitative and quantitative characterization of moraine morphology, including the determination of surface slope angle, allowed us to make estimates of relative age and thus make

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correlations within northeast Siberia as well as make tentative correlations with the glacial record in Alaska.

The sequences of moraines in Chukotka are similar, in morphology and position, to moraine sequences described in Alaska, and record a succession of glacial events that most likely began in the middle Pleistocene and ended with the Late Wisconsin. An extensive glacial advance, marked by low-angle, degraded moraines located 200-300 km from the source area, is probably correlative with the middle Pleistocene age moraines described from the Seward Peninsula and the Brooks Range of Alaska. Glaciers of this age advanced across the Bering Shelf and encroached on St. Lawrence Island. An intermediate series of moraines, characterized by hummocky and kettle topography, probably marks a separate advance that may be correlative with the Salmon Lake glaciation on the Seward Peninsula and the early Wisonson glacial advance that encroached on St.Lawrence Island. During the youngest glaciation recorded in Chukotka, valley glaciers advanced only 12-60 kilometers from their high mountain cirques. This glacial advance is recorded by moraines exhibiting 'fresh' morphologies with steep slope angles and rough surfaces. These moraines are comparable in appearance and morphology to Late Wisconsin aged moraines in Alaska.

This study has shown that SAR imagery holds great promise as a reconnaissance tool for mapping and moraine morphological investigations in regions of the world that have little or no aerial photo or large-scale map coverage. The unique properties that allow the determination of criteria such as surface slope, may make it applicable as an

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additional component in the study glacial features using established methods. The map created in this study served not only as a direct test of the Marine Ice Trangression Hypothesis (MITH), but also provides an essential background for all continuing studies of glacial history in the region. Although the age assignments are rough estimates, they have enabled us to develop a preliminary framework for future endeavors aimed at correlating glacial history across Bering Strait.

Field mapping and stratigraphic work on St. Lawrence Island revealed that ice advanced onto the island twice in the late Pleistocene, once in the Middle Pleistocene and once after the Last Interglacial, probably during the Early Wisconsin. The record of glaciers advancing from Chukotka onto the island provides an important 'Rosetta Stone' for correlating the glacial histories of northeast Siberia and Alaska.

Much of this study was directed toward the controversy over glacial extent in Beringia during the last glacial maximum. The MITH ice sheet, had it been present should have induced a regional lithshperic response. By modelling the amount of isostatic recovery expected after deglaciation I was able to directly test the extensive glaciation hypothesis postulated by Grossvald and Hughes for the Last Glacial Maximum (LGM). Using several load histories, all of which underestimated both the size and duration of the MITH ice sheet, my results showed that there should be distinct evidence of either isostatic uplift or subsidence, on the scale of meters to tens of meters, around the shores of present-day Bering Strait. Examination of coastlines and careful dating at Cape Espenberg

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show that sea level has remained essentially unchanged in Bering Strait over the last 5000 years.

The reconstruction of paleo-equilibrium line altitudes also provided important evidence contrary to the MITH ice sheet. Snowlines during the LGM rose from southeast to northwest, dipping as low as 400 meters on the tip of Chukotka to approximately 1000 meters on the north shore of the peninsula. This data directly contradicts the assertion that the LGM snowline, used to 'grow' the MITH ice sheet, dipped to the north and was lowered 1000 meters below present-day snowline. The southwest-to-northeast rise in LGM, as well as modern, snowlines in Chukotka also indicates that Bering Sea is the dominant moisture source in the region and that climate patterns in Chukotka roughly mirror the trend in Alaska.

This study has used a multidisciplinary approach to directly test the opposing hypotheses regarding Late Wisconsin glaciation in Beringia. The outcome is clear evidence against the existence of a large marine-based ice sheet in Beringia at any time in the late Pleistocene. The results of this work also provide important base level information, in the form of a glacial geologic map and a preliminary chronologic framework for the glacial history of northeast Siberia. The methodologies and techniques developed in the course of this research have advanced our capabilities to collect information from areas, such as Russia, where the lack of map and photo coverage has hindered such studies before now.

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Future Work:

Clearly more research is needed on the timing of glacial events in Chukotka and other areas of western Beringia. This work is, in fact, continuing at the University of Massachusetts with the efforts Julie Brigham-Grette and her students. They are using cosomgenic isotope dating and ground based studies of moraine morphology to date the sequences present in the Tanuyer and Anadyr valleys. The results of these studies will be an important contribution to the regional glacial history as well as important ground-truthing for the satellite work done here.

This study focused on the Late Wisconsin glacial period, however, similar attention could be directed toward the earlier glacial advances recorded in Beringia. Although difficult to obtain for older glaciations, a reconstruction of snowline during earlier ice ages may provide some clues as to why glaciers were so much more extensive than during subsequent cold periods. Has there been a change in circulation patterns? Were the earlier glaciations initiated 'out of phase' and thus during periods of high sea level and greater moisture availability? This latter question could also be addressed using the isostatic model developed here.

The presence of glaciomarine sediments near present sea level have led several researches to propose that the Middle Plestocene (Nome River) glaciation occurred out of phase with global ice buildup (Huston 1990, Roof 1994). While this is an interesting concept, and one certainly meriting careful consideration, no attempt was made to rule out the potential isostatic influence of the large piedmont-type glaciers that flowed from

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Chukotoka to St. Lawrence Island, and from the Brooks Range to Kotzebue Sound. Is the presence of marine waters, near to today's level, an indication of global sea level or is it due to a local isostatic effect induced by early Beringian glacial advances? An understanding of the mechanisms that might be responsible for a stratigraphic record such as the one found in Kotzebue Sound will be very important to the study of climate change and especially to understanding why the extent of glaciation during ice ages in Beringia has decreased by an order of magnitude since the Middle Pleistocene.

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Appendix I. Isostatic responses to disc loads ranging in size from 5 to 200 kilometers radius. (6 pages)

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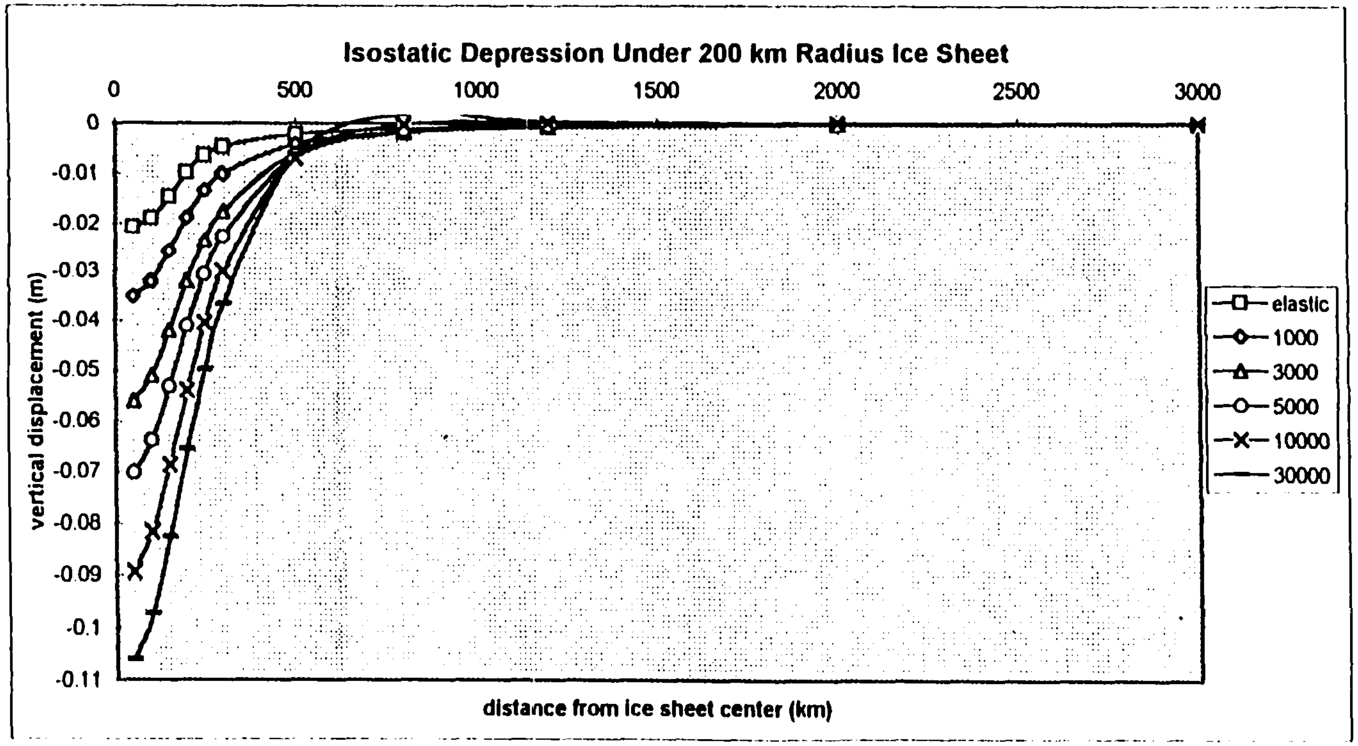
Isostatic Depression Under 200 km radius Ice Sheet (1 meter thick)

calculated for elastic response and response after 1,3,5,10, and 30 thousand years
at distances 50-3000 kilometers from center of ice sheet

	50	100	150	200	250	300	500	800	1200	2000	3000
elastic	-0.02084	-0.01897	-0.01461	-0.00983	-0.00638	-0.00474	-0.00218	-0.00115	-0.00073	-0.000417	-0.00026
1000	-0.03523	-0.03206	-0.0258	-0.01888	-0.0134	-0.01009	-0.00417	-0.00185	-0.00095	-0.000386	-0.00018
3000	-0.0561	-0.05108	-0.04212	-0.03209	-0.02357	-0.01762	-0.00602	-0.00195	-0.00072	-0.000179	-6.7E-05
5000	-0.0702	-0.06399	-0.05329	-0.04121	-0.03061	-0.02276	-0.00673	-0.00153	-0.00039	-5.13E-05	-3.6E-05
10000	-0.08933	-0.08165	-0.06872	-0.05396	-0.04053	-0.02993	-0.00695	-0.00024	0.000118	2.03E-05	-4E-05
30000	-0.10608	-0.09727	-0.0826	-0.06565	-0.04977	-0.03674	-0.00632	0.001825	0.000441	-4.89E-05	-7E-05



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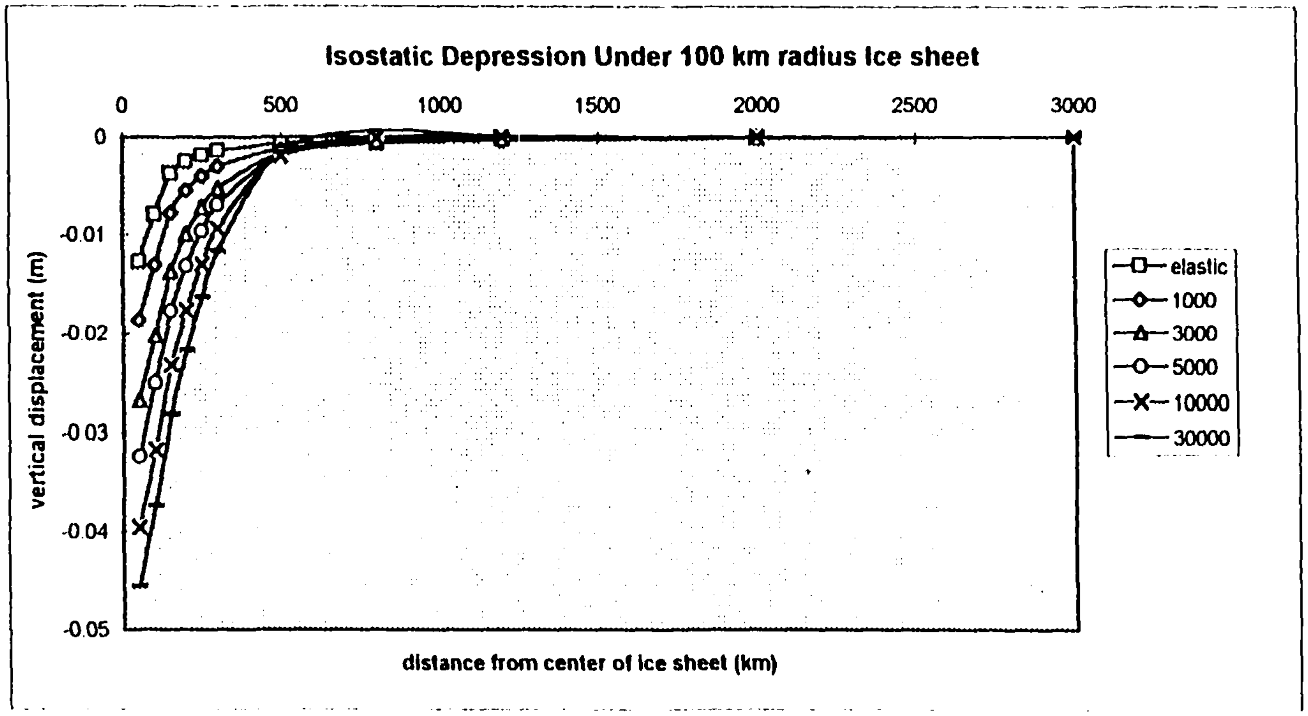
Isostatic Depression Under 100 km radius Ice Sheet (1 meter thick)

calculated for elastic response and response after 1,3,5,10, and 30 thousand years
at distances 50-3000 kilometers from center of ice sheet

	50	100	150	200	250	300	500	800	1200	2000	3000
elastic	-0.01266	-0.00795	-0.00382	-0.00254	-0.00184	-0.00142	-0.00068	-0.00037	-0.00023	-0.000134	-8.4E-05
1000	-0.01855	-0.01297	-0.00777	-0.0055	-0.00403	-0.00304	-0.0013	-0.00059	-0.00031	-0.000125	-5.7E-05
3000	-0.02686	-0.02022	-0.01362	-0.00995	-0.00728	-0.00535	-0.00183	-0.00061	-0.00023	-5.77E-05	-2.2E-05
5000	-0.03231	-0.02507	-0.01764	-0.01309	-0.00959	-0.00697	-0.00199	-0.00047	-0.00012	-1.66E-05	-1.2E-05
10000	-0.03952	-0.03161	-0.02321	-0.01755	-0.01294	-0.0093	-0.00195	-4.3E-05	3.84E-05	6.56E-06	-1.3E-05
30000	-0.0456	-0.03726	-0.0282	-0.02169	-0.01615	-0.01158	-0.00161	0.000648	0.000137	-1.58E-05	-2.3E-05



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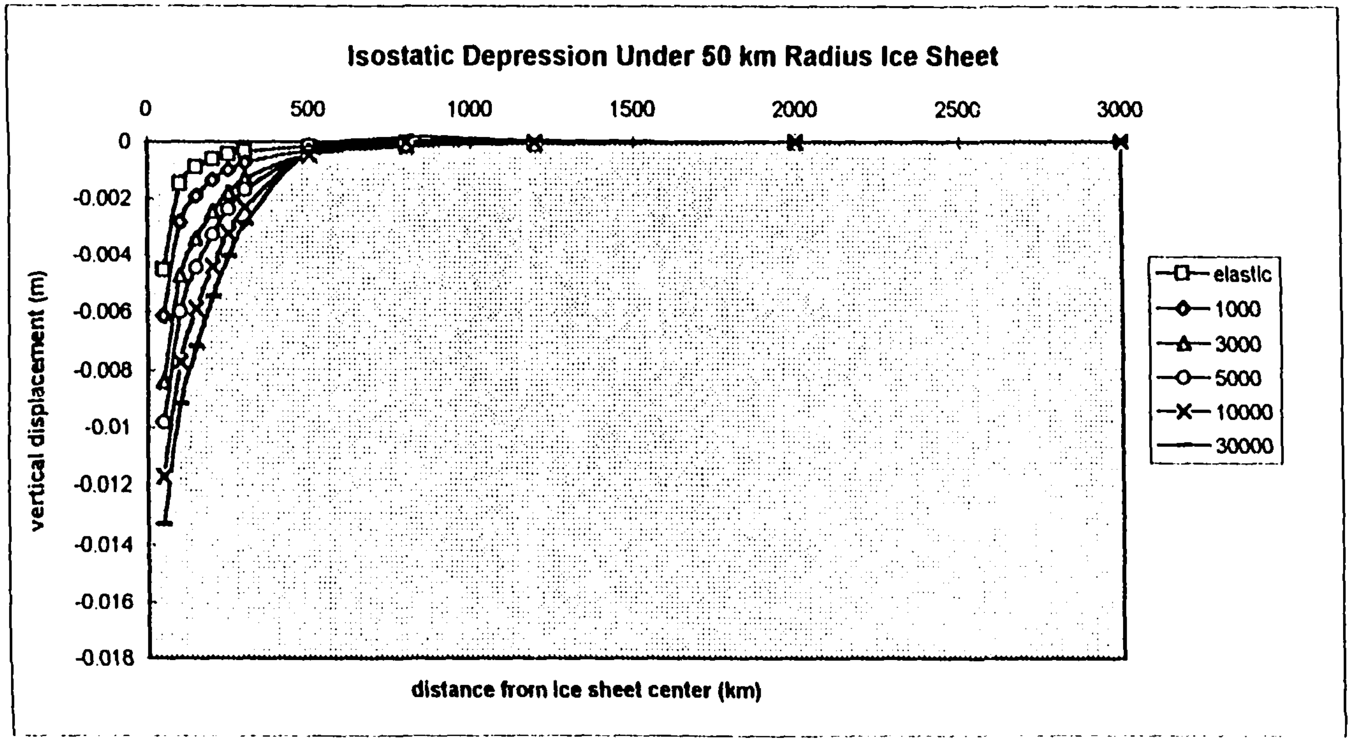
Isostatic Depression Under 50 km radius Ice Sheet (1 meter thick)

calculated for elastic response and response after 1,3,5,10, and 30 thousand years
at distances 50-3000 kilometers from center of ice sheet

	50	100	150	200	250	300	500	800	1200	2000	3000
elastic	-0.00452	-0.00149	-0.00089	-0.00061	-0.00045	-0.00035	-0.00017	-9.2E-05	-5.9E-05	-3.37E-05	-2.1E-05
1000	-0.00613	-0.00282	-0.0019	-0.00135	-0.00099	-0.00074	-0.00032	-0.00015	-7.7E-05	-3.12E-05	-1.4E-05
3000	-0.00838	-0.00473	-0.0034	-0.00247	-0.00179	-0.00131	-0.00045	-0.00015	-5.8E-05	-1.45E-05	-5.4E-06
5000	-0.00983	-0.006	-0.00444	-0.00326	-0.00237	-0.00171	-0.00048	-0.00012	-3.1E-05	-4.16E-06	-2.9E-06
10000	-0.01172	-0.00771	-0.00588	-0.0044	-0.00322	-0.00229	-0.00046	-8.4E-06	9.7E-06	1.65E-06	-3.2E-06
30000	-0.0133	-0.00917	-0.00717	-0.00547	-0.00404	-0.00287	-0.00037	0.000166	3.51E-05	-3.94E-06	-5.6E-06



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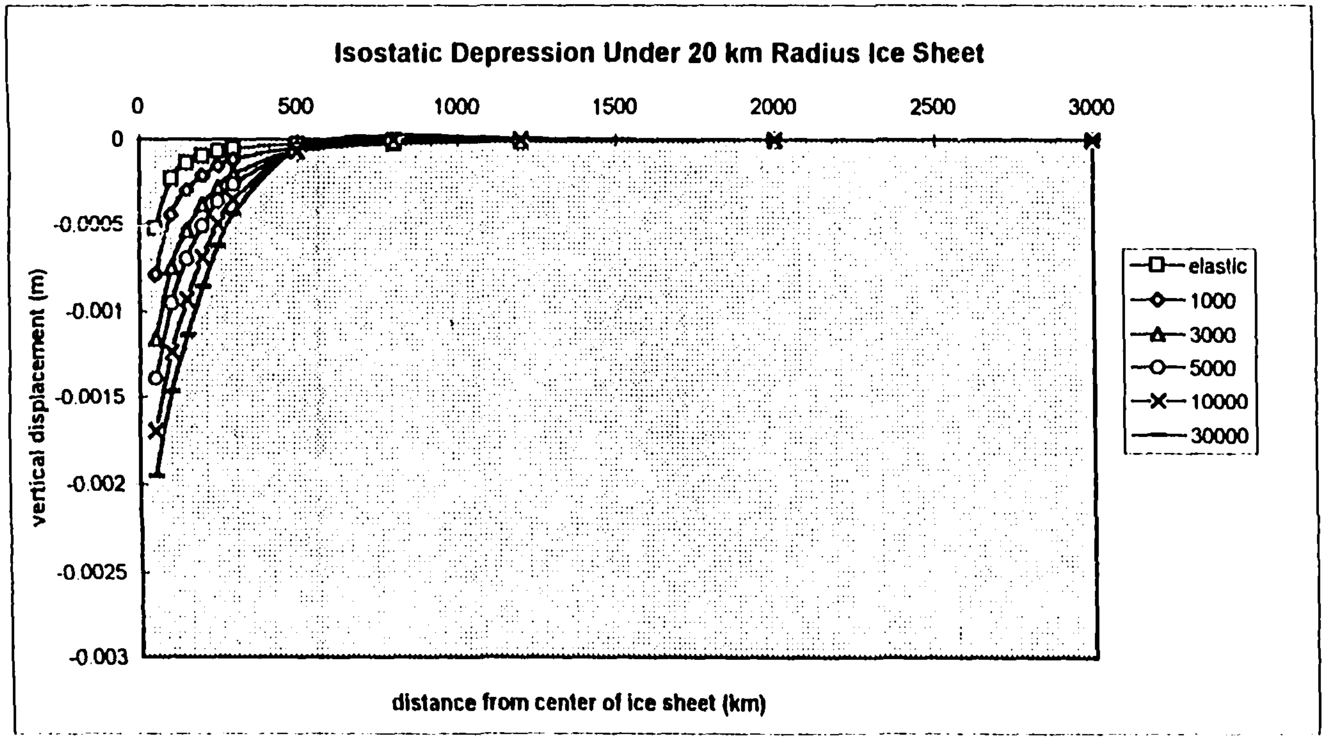
Isostatic Depression Under 20 km radius Ice Sheet (1 meter thick)

calculated for elastic response and response after 1,3,5,10, and 30 thousand years
at distances 50-3000 kilometers from center of ice sheet

	50	100	150	200	250	300	500	800	1200	2000	3000
elastic	-0.00052	-0.00023	-0.00014	-9.4E-05	-6.9E-05	-5.4E-05	-2.7E-05	-1.5E-05	-9.6E-06	-5.12E-06	-3.4E-06
1000	-0.00079	-0.00044	-0.0003	-0.00021	-0.00015	-0.00012	-5E-05	-2.3E-05	-1.3E-05	-4.58E-06	-2.3E-06
3000	-0.00116	-0.00075	-0.00054	-0.00038	-0.00028	-0.0002	-6.9E-05	-2.4E-05	-9.7E-06	-2.02E-06	-8.6E-07
5000	-0.00139	-0.00096	-0.0007	-0.00051	-0.00037	-0.00026	-7.4E-05	-1.8E-05	-5.3E-06	-5.64E-07	-4.7E-07
10000	-0.0017	-0.00123	-0.00093	-0.00069	-0.0005	-0.00035	-6.9E-05	-9.5E-07	1.59E-06	1.63E-07	-5.2E-07
30000	-0.00195	-0.00147	-0.00114	-0.00086	-0.00063	-0.00044	-5.2E-05	2.65E-05	6.18E-06	-6.74E-07	-9E-07



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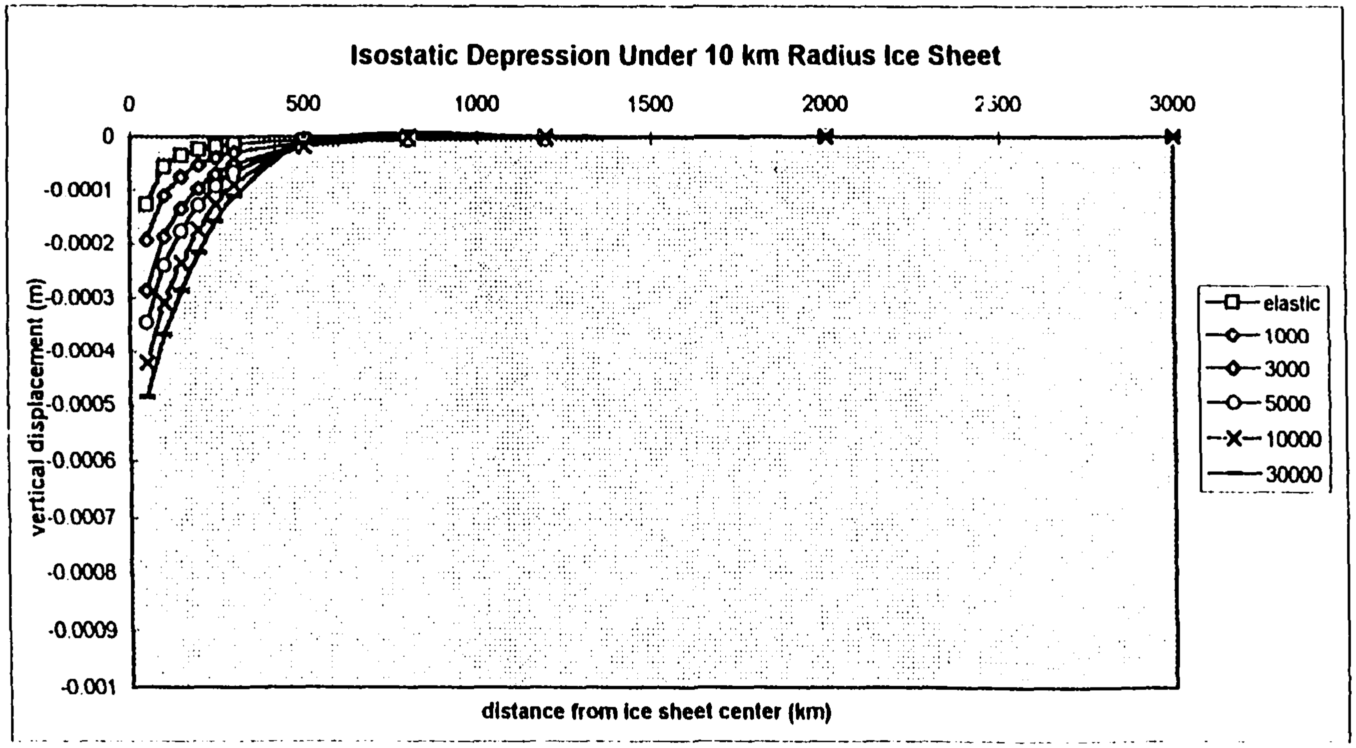
Isostatic Depression Under 10 km radius Ice Sheet (1 meter thick)

calculated for elastic response and response after 1,3,5,10, and 30 thousand years
at distances 50-3000 kilometers from center of ice sheet

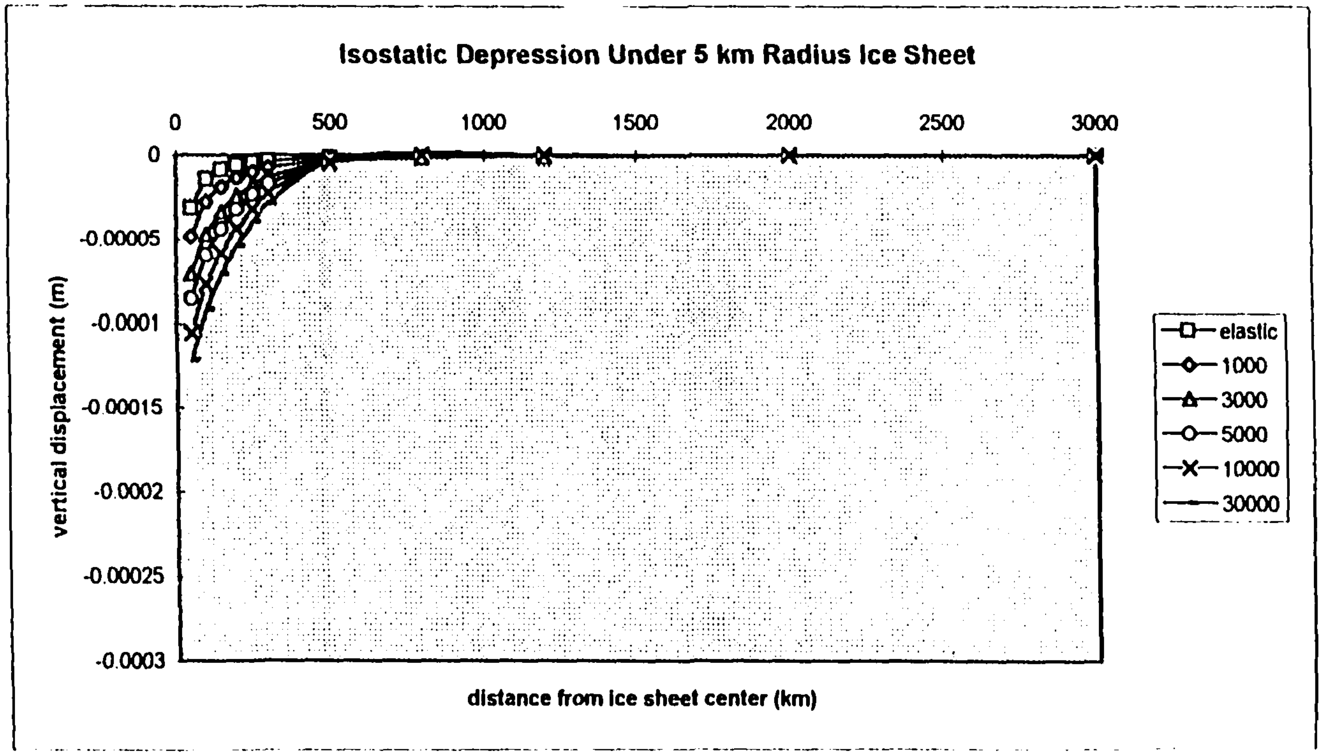
	50	100	150	200	250	300	500	800	1200	2000	3000
elastic	-0.00013	-5.7E-05	-3.5E-05	-2.4E-05	-1.8E-05	-1.4E-05	-6.7E-06	-3.7E-06	-2.4E-06	-1.29E-06	-8.4E-07
1000	-0.00019	-0.00011	-7.5E-05	-5.3E-05	-3.9E-05	-2.9E-05	-1.3E-05	-5.8E-06	-3.2E-06	-1.17E-06	-5.7E-07
3000	-0.00029	-0.00019	-0.00014	-9.7E-05	-7E-05	-5.1E-05	-1.8E-05	-6E-06	-2.4E-06	-5.23E-07	-2.2E-07
5000	-0.00035	-0.00024	-0.00018	-0.00013	-9.3E-05	-6.7E-05	-1.9E-05	-4.6E-06	-1.3E-06	-1.47E-07	-1.2E-07
10000	-0.00042	-0.00031	-0.00023	-0.00017	-0.00013	-9E-05	-1.8E-05	-2.7E-07	3.98E-07	4.66E-08	-1.3E-07
30000	-0.00049	-0.00037	-0.00029	-0.00022	-0.00016	-0.00011	-1.4E-05	6.67E-06	1.54E-06	-1.66E-07	-2.3E-07



Isostatic Depression Under 5 km radius Ice Sheet (1 meter thick)

calculated for elastic response and response after 1,3,5,10, and 30 thousand years
at distances 50-3000 kilometers from center of ice sheet

	50	100	150	200	250	300	500	800	1200	2000	3000
elastic	-3.2E-05	-1.4E-05	-8.7E-06	-6E-06	-4.4E-06	-3.4E-06	-1.7E-06	-9.2E-07	-6E-07	-3.32E-07	-2.1E-07
1000	-4.8E-05	-2.8E-05	-1.9E-05	-1.3E-05	-9.7E-06	-7.3E-06	-3.2E-06	-1.5E-06	-7.9E-07	-3.05E-07	-1.4E-07
3000	-7.1E-05	-4.7E-05	-3.4E-05	-2.4E-05	-1.8E-05	-1.3E-05	-4.4E-06	-1.5E-06	-6.1E-07	-1.39E-07	-5.4E-08
5000	-8.6E-05	-6E-05	-4.4E-05	-3.2E-05	-2.3E-05	-1.7E-05	-4.8E-06	-1.2E-06	-3.3E-07	-3.98E-08	-2.9E-08
10000	-0.00011	-7.7E-05	-5.9E-05	-4.4E-05	-3.2E-05	-2.3E-05	-4.5E-06	-7.1E-08	9.94E-08	1.47E-08	-3.2E-08
30000	-0.00012	-9.2E-05	-7.2E-05	-5.5E-05	-4E-05	-2.8E-05	-3.5E-06	1.67E-06	3.86E-07	-4.02E-08	-5.6E-08



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LEFT TO RIGHT, TOP TO BOTTOM, WITH SMALL OVERLAPS

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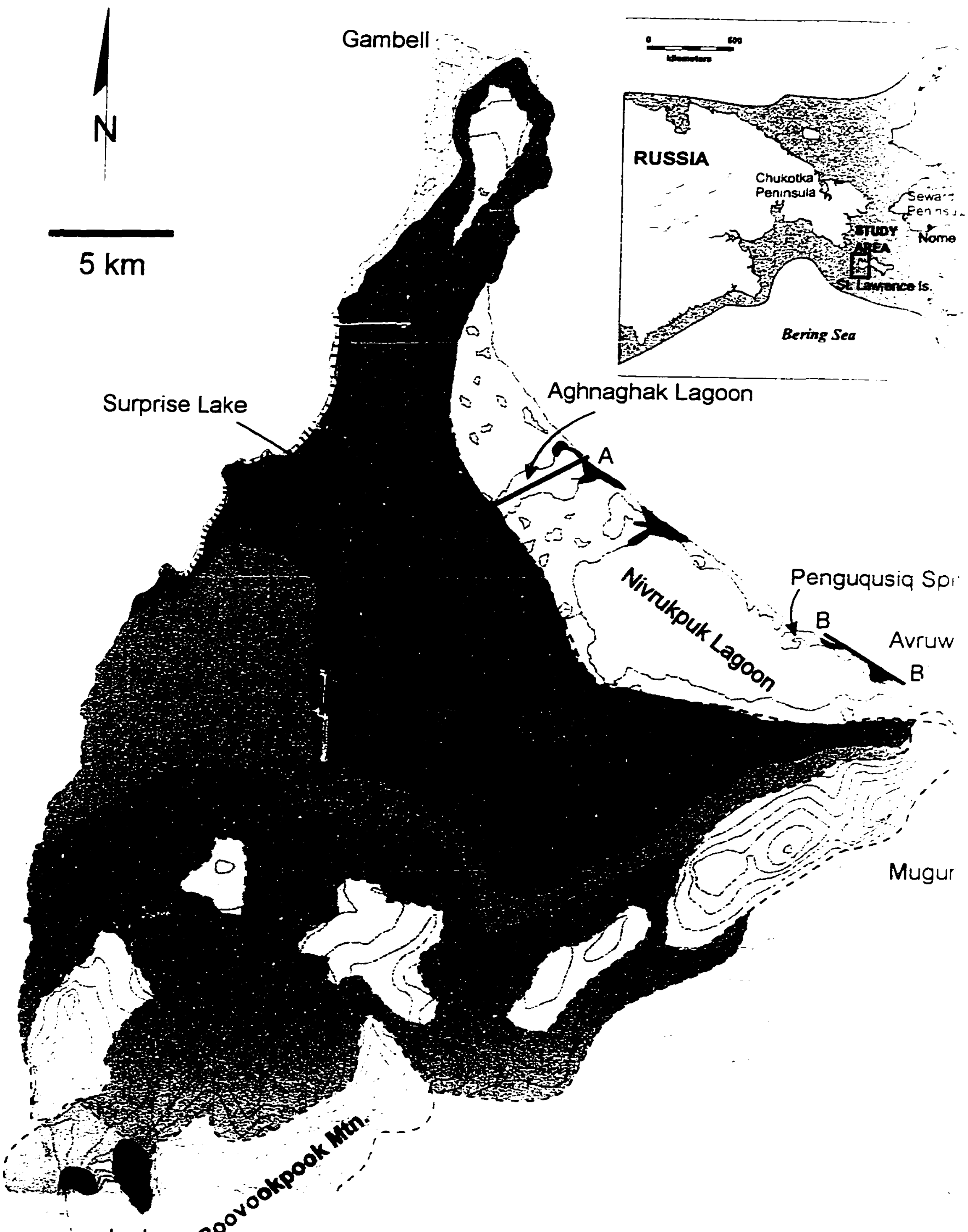
SURFICIAL GEOLOGIC MAP

Patricia A. I



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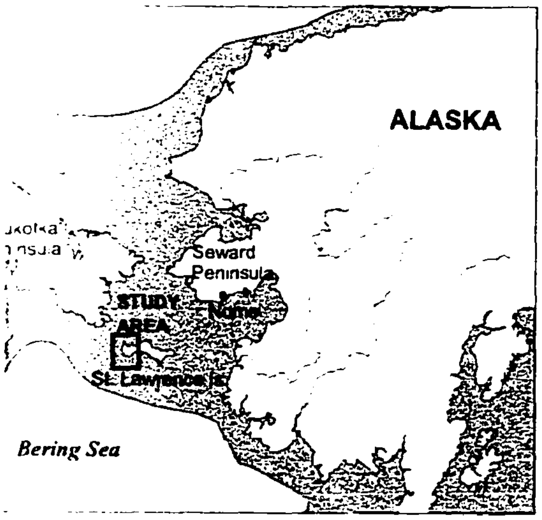
MAP OF NORTHWEST ST. LAWREN

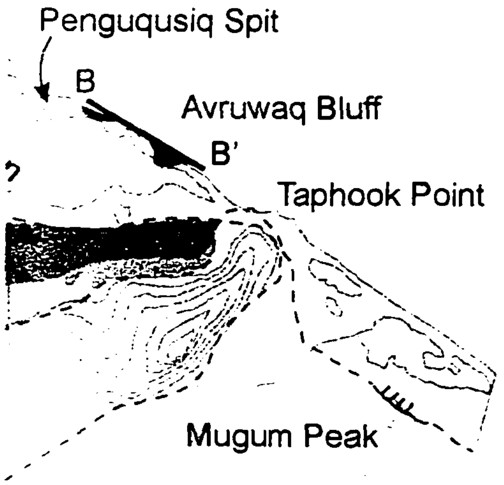
:ricia A. Heiser, Sonja L. Benson, and David M. Hopkir



Holocene Peat
Exposed along the shores of Aghnaghak and Nivrukpuk Lagoons and in the eroding banks of several thaw lakes surrounding the lagoons.

Sivuqaq Glacial Diamicton
The Pelukian gravels exposed in Avruwaq Bluffs are deformed by glacial tectonics and the structure of the deformation features (thrust faults, chevron folding) suggest glacial encroachment from the northwest. In places (see section diagram below), glacial diamicton is injected into the gravels. The diamict is composed of unsorted silty boulder gravels containing angular striated clasts. Granite boulders of up to 1 m are common. Numerous schist and quartzite erratics are present and indicate a source in the mountains of the Chukotka Peninsula to the northwest. In 1973 Hopkins *et al.* reported a thin (~1m) mantle of till on top of the section, above the Pekukian gravels. Today there is no till present on the top of the section. Rapid coastal retreat has evidently removed the sediments seen by Hopkins in 1968. This information combined with the pro-glacial style of deformation indicate that Avruwaq Bluff probably marks the maximum extent of glacial ice during this event.

Clearly this glacial event post-dates the Last Interglacial high sea level stand recorded by the Pelukian gravels and probably dates to the Early Wisconsin (70-80,000 yrs BP) or possibly to a cold period late in oxygen isotope Stage 5.



Pelukian Marine Gravels
Exposed in section along the entire length of Avruwaq Bluffs, along the northern shores of Aghnaghak Lagoon and along the southern shore of Fox Lake.

This unit is composed of stratified shelly pebble gravel and medium to coarse sand in beds ranging in thickness from 5 - 150 centimeters. The lithology of the pebbles is mixed but dominated by the Cretaceous igneous rocks (quartz monzonite) that form much of NW St. Lawrence Island. Near the channel mouth of Aghnaghak Lagoon, frequent, washed (but still distinctly striated and sculpted) erratics of schist and quartzite are found in this unit. These are probably reworked from older glacial sediments. The unit is glaciotectonically deformed in many places (see Figure 3).

Hiatella and *Mya* shells from these deposits yield mean total amino acid ratios of $.04 \pm >013$ and are correlative with ratios from Pelukian sediments near Nome (Benson 1994). These sediments mark the last high sea level event in Beringia, which occurred during the Last Interglacial at 125, 000 years BP.

Nome River Glacial Drift
These glacial sediments are exposed at the channel mouth of Aghnaghak Lagoon and in places along the shores of Nivrukpuk and Aghnagak lagoons. Where found in section these deposits contain unsorted sandy boulder clay with angular and sculpted granites, as well as erratic schists, limestone, and quartzite. Inland, these sediments are present only as a mantle of frost churned, silty, sediments containing angular and erratic pebbles and cobbles. South of Aghnaghak Lagoon, the drift is identified only by a lag of glacially striated sculpted erratic cobbles and boulders

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\WRENCE ISLAND, ALASKA

M. Hopkins

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The sediments in this unit are typically medium to fine tan sands with abundant fossils. They are glaciotectonically deformed at most localities, including Surprise Lake.

Hiatella and *Mya* shells from Nivrukpuk Lagoon and the entrance of Aghnaghagak Lagoon yield amino acid ratios of $.150 \pm .032$ (Benson 1994). These ratios are very similar to the ratios obtained from sediments left by the Anvillian marine transgression recorded at Nome (Kaufman 1991). This marine transgression probably occurred during oxygen isotope Stage 11 (Kaufman et al. 1991).

Older Marine Sediments
Underlying the Anvillian sands at Surprise Lake, is a unit of gray, slightly finer, fossilferous sand. Amino acid ratios from this unit suggest that an older marine transgression my be recorded here as well.

Cirque Glacier Deposits
Glacial deposits are present in and at the mouth of several cirques on the southwest comer of the island in Koguk Basin and also at Northeast Cape (out side of study area). The age of these cirques, and the glacial deposits that accompany them, is difficult to estimate. The cirque at Ivekan mountain has been glaciated several times. Ages assigned on the map are tentative. Fresh headwall morphology suggests the cirques may have been occupied during the Late Wisconsin. However, further research is required before any definitive age can be assigned to these features.

Nome River **Sivuqaq** **Late Wisconsin**

Colluvium
The hills and mountain slopes of the southern part of the study are covered in a mantle of colluvium. This material contains a high percentage of granite pebbles and grus. Solifluction and frost creep are the dominant geomorphic processes in these areas.

Holocene Beach Spit
Beach gravels at present site of Gambell form ridges that date to the Holocene.

Bedrock

MAP SYMBOLS

wave-cut terrace

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IDEALIZED CROSS-SECTION OF STR
ALONG THE SHORES OF AGHN

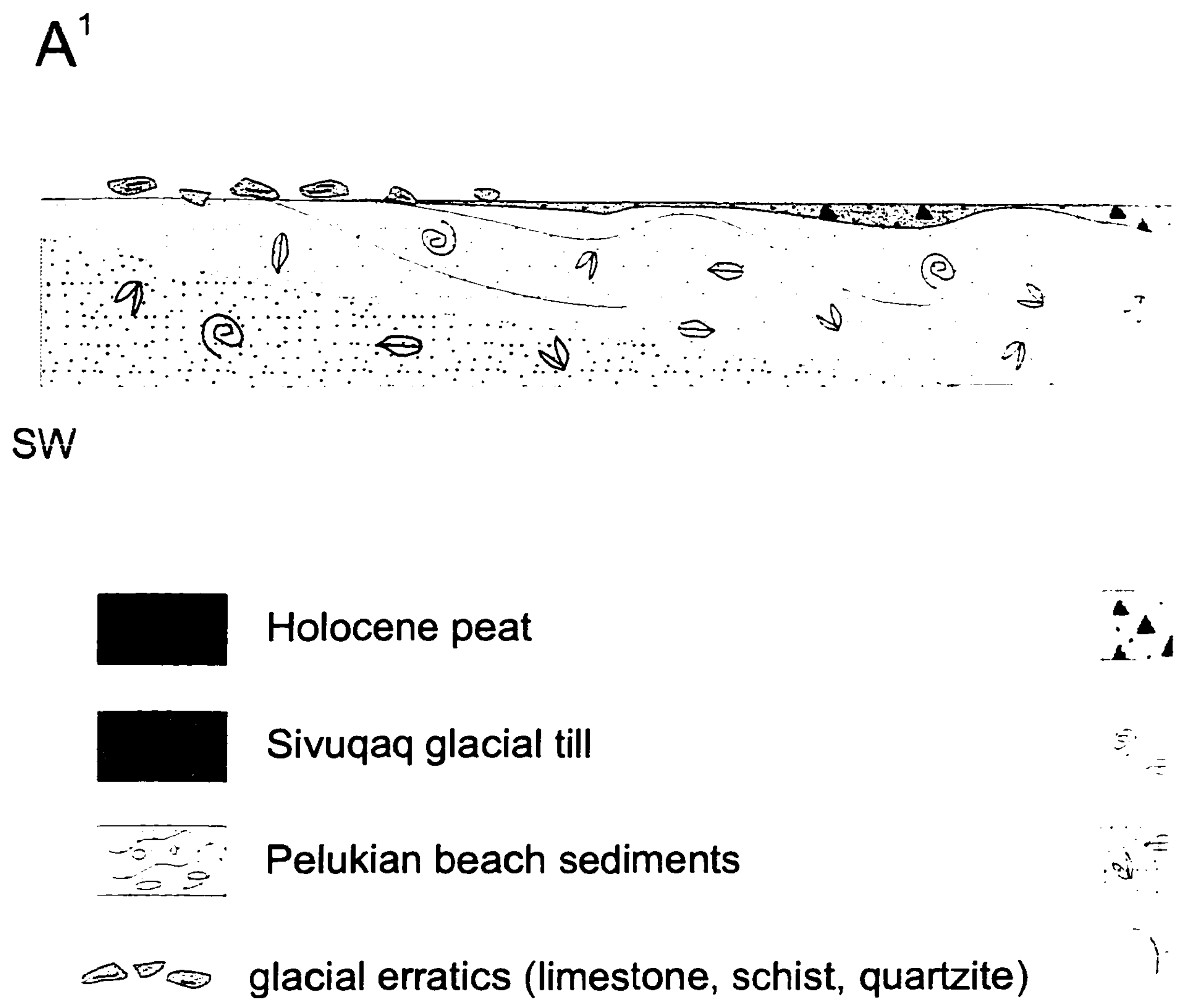

Figure 3. Cross-section A - A'. Idealized cross-section illu
shores of Aghnaghak Lagoon. The sediments show a p

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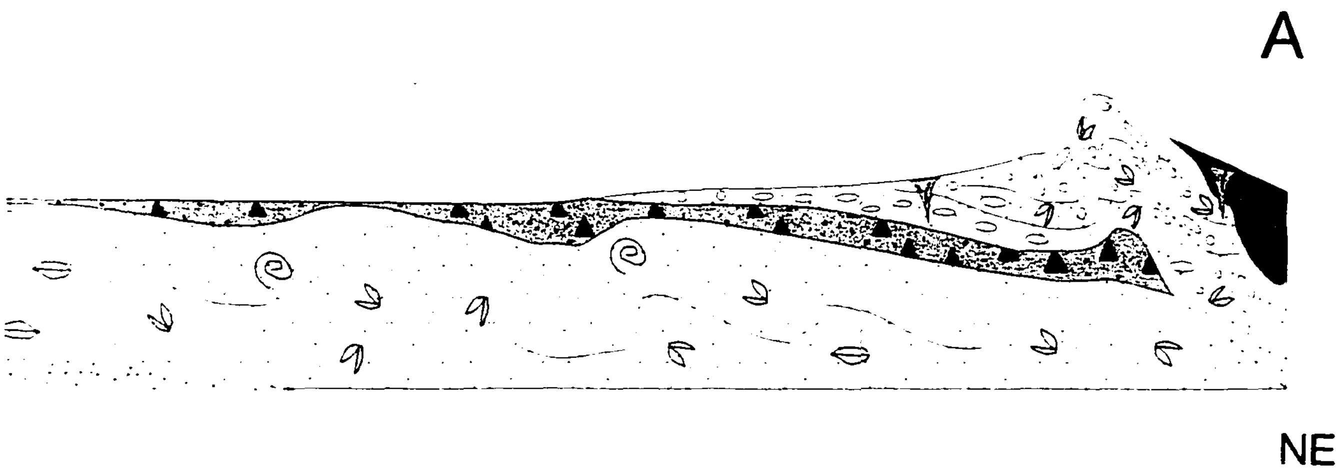
ECTION OF STRATIGRAPHY EXPOSED
ORES OF AGHNAGHAK LAGOON



Nome River age glacial till

Anvilian shelly marine sands

ɜnts

older shelly marine sands

ıe, schist, quartzite)

ice wedge casts

ɑlized cross-section illustrating stratigraphy exposed along both
ıe sediments show a possible Early Pleistocene marine inundation

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Figure 1. Surficial geologic map of northwest St. Lawrence

from Benson 1994.

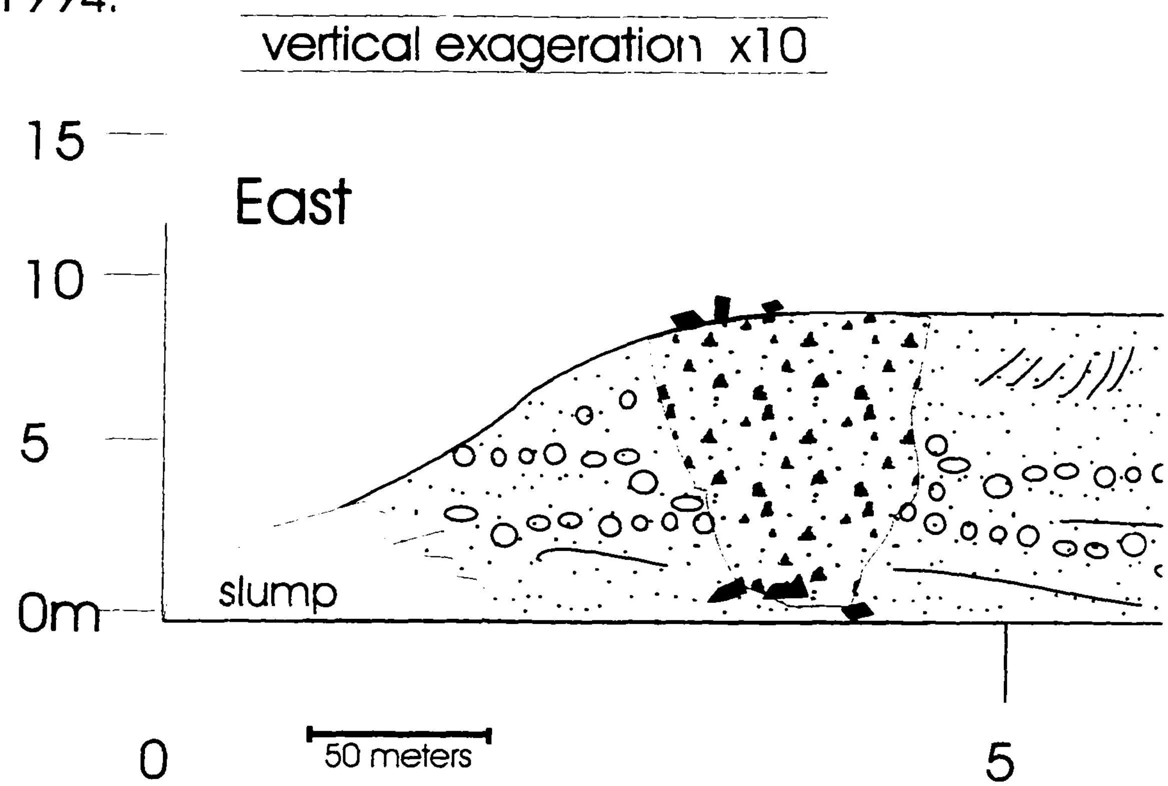

Figure 2. Cross-section B - B" Avruwaq Bluffs. This section is through
and graben sets suggest glacial encroachment from the northwes
PK 17-18) till is injected into the interglacial marine deposit. The strc

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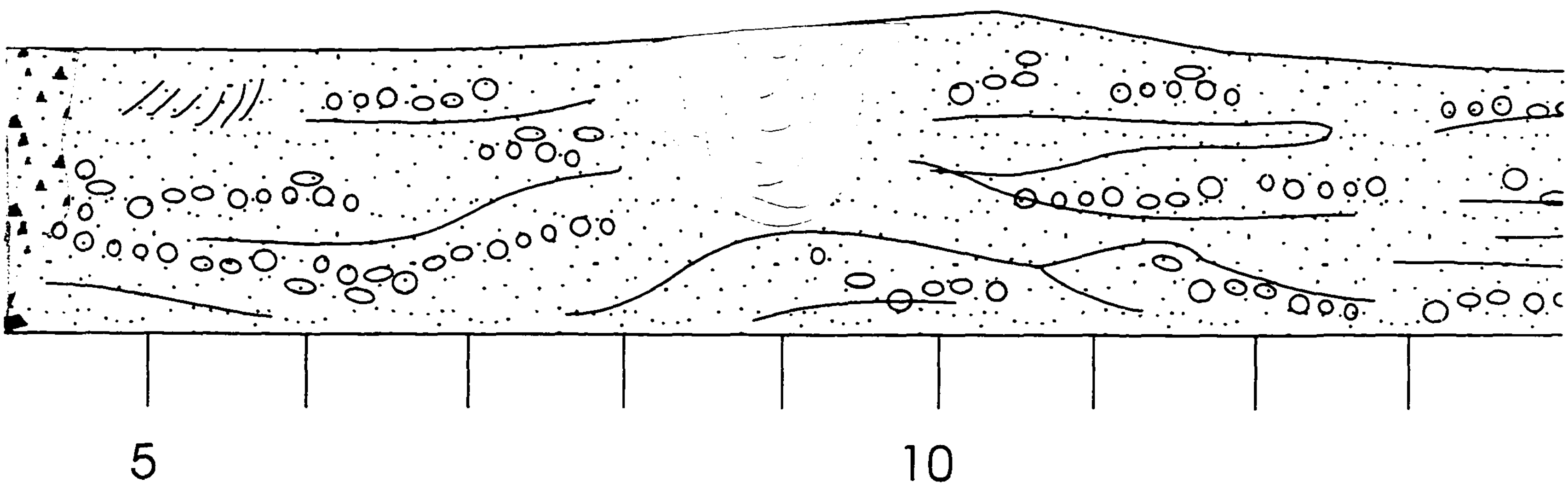
South of Aghnaghak Lagoon, the drift is identified only by a lag of glacially striated, sculpted, erratic cobbles and boulders.

These glacial sediments predate the Last Interglacial and overlie Anvillian aged marine sediments (see below). Comparisons with sediments at Nome and Kotzebue suggest that these sediments correspond to the Nome River Glaciation, and may date to oxygen isotope Stage 10 (~410,000 kyBP) (Kaufman *et al.* 1991).

Anvillian Marine Sediments (under Nome River Drift)
Marine sediements of this age are exposed along the north coast of the island from Aghnaghak Lagoon to the western shore of Nivrukpuk Lagoon and in section along the shores of Aghnagak Lagoon. Marine sediments of this age also crop out along the shore of Surprise Lake.

t St. Lawrence Island showing the limits of marine transgressions and glaciatio



tion is through the deformed Pelukian gravels that form Avruwaq Bluffs along
n the northwest. These structures and the presence of glacial erratics indiger
eposit. The stratigraphy and landforms here suggest that Avruwaq Bluff, and I

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·Drift)
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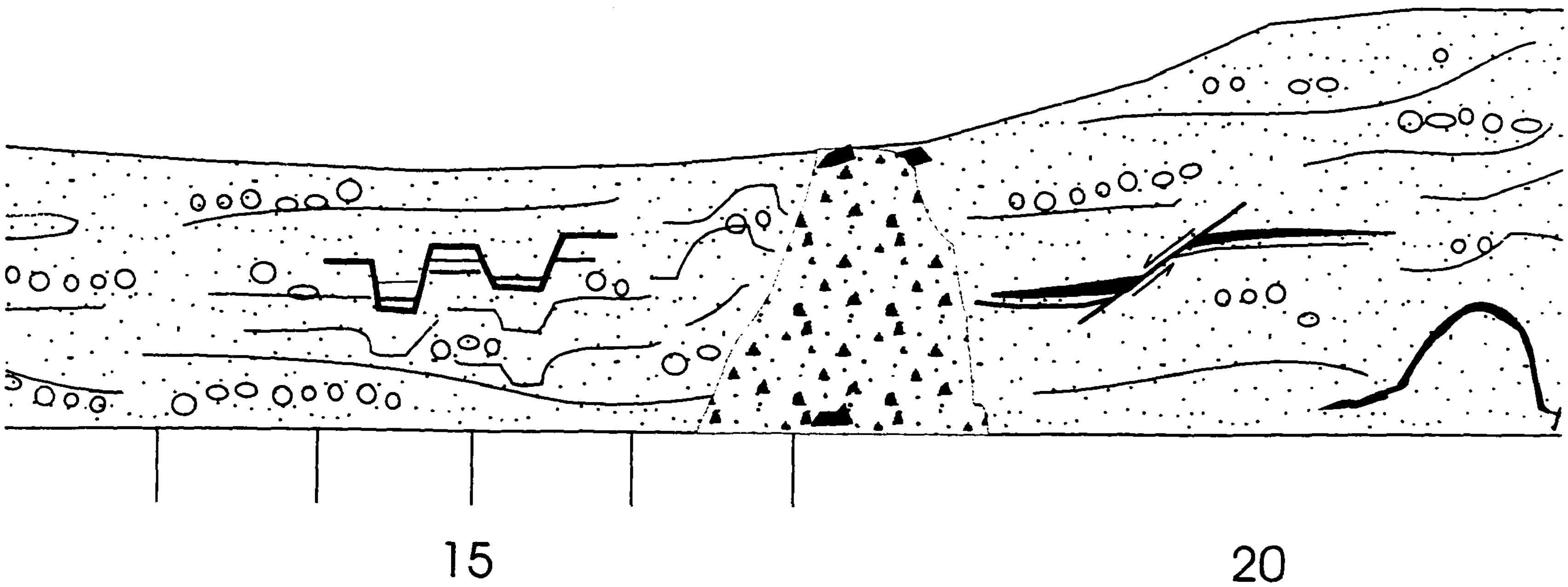
ıs and glaciation since the Middle Pleistocene.



ɔq Bluffs along Penguqusiq Spit. The 'spit' or barrier separates Nivrukpuk Lagoc
erratics indigenous to mainland Chukotka indicate that the mountains of Chı
waq Bluff, and Penguqusiq Spit, were located at the glacial margin during the

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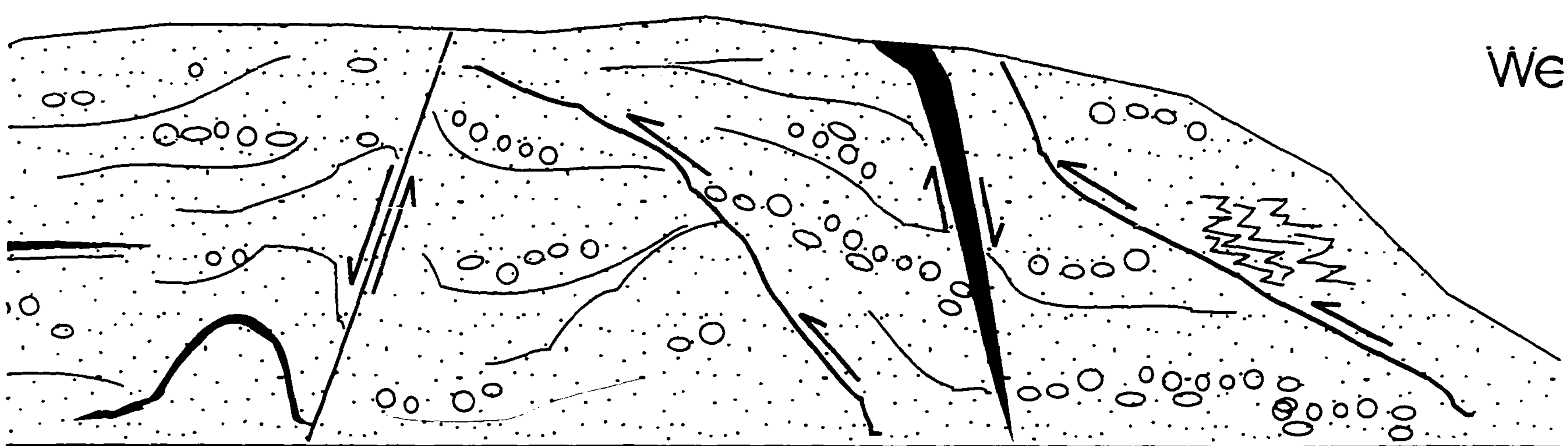
Figure 3. Cross-section A - A'. Idealized cross-sectic
shores of Aghnaghak Lagoon. The sediments show
followed by the Anvillian transgression. Nome River
and deposited drift over the interglacial units. Errati
redeposited in the Pelukian beach deposits. The Pe
along much of the northwest coast of the island. Ti
Sivuqaq advance which probably occurred during



Nivrukpuk Lagoon from the open Bering Sea. Glaciotectonic structures includ
nountains of Chukotka were the source of glaciers that encroached on the isl
argin during the maximum extent of the most recent (probably Early Wiscons

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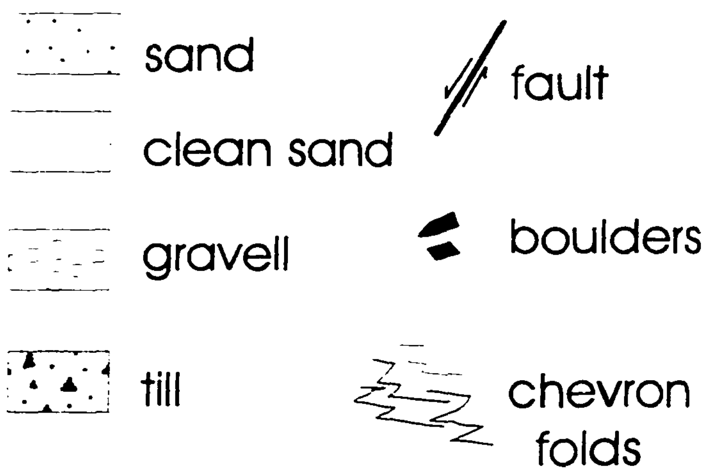
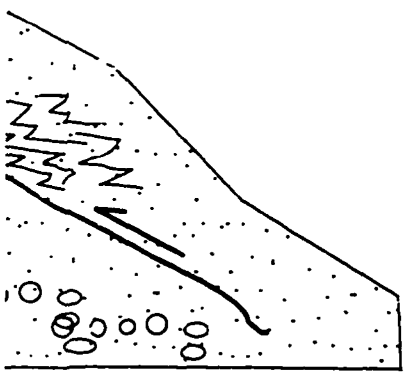
sediments show a possible Early Pleistocene marine inundation on. Nome River age glaciers deformed the marine sediments acial units. Erratics from the Nome River drift are reworked and deposits. The Pelukian fossiliferous sands and gravels are exposed t of the island. These sediments are, in turn, deformed by the occurred during the Early Wisconsin.

Legend



West



30 ← picket number

structures including thrust faults, normal faults, folds, and horst ached on the island (Benson 1994). In two places (PK 3-4 and bly Early Wisconsin) glacial advance onto the island.

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